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Plate Tectonics & Crustal Evolution

Third Edition

by

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*To Carolyn,
Tamara, Linda and
Nathan*

Preface

This book has grown out of a course I teach at New Mexico Tech. The rapid accumulation of data related to plate tectonics and the origin of continents in the past decade has necessitated continued updating of the course. The book is written for an advanced undergraduate or graduate student, and it assumes a basic knowledge of geology, chemistry, and physics that most students in the Earth Sciences acquire during their undergraduate education. It also may serve as a reference book for various specialists in the geological sciences. I have attempted to synthesize and digest data from the fields of oceanography, geophysics, geology, and geochemistry and to present this information in a systematic manner addressing problems related to the evolution of the Earth's crust over the last 3.8 Ga. The role of plate tectonics in the geological past is examined in light of geologic evidence and examples of plate reconstructions are discussed.

Since the first and second editions of the book were published, a wealth of information related to plate tectonics and continental origin has appeared in scientific journals. To accommodate this new information, it was necessary to rewrite more than 75% of the text as it appeared in the second edition. Also, a large number of new figures have been added and the tectonic map of the world has been updated. The third edition includes new sections on meteorites, seismic tomography, mantle convection, accretionary terranes, mantle sources and evolution, continental growth, secular changes in Earth history, Venus, and a new chapter on exogenic Earth systems. In addition, the following topics have been substantially revised: lunar origin, global gravity, origin of the core, metamorphism, plate boundaries, hotspots, tectonic settings, magma associations, Phanerozoic orogenic systems, and crustal origin and evolution.

The general approach is much the same as in the first and second editions. Historical background and major physical properties of the Earth are briefly summarized in Chapter 1 and the origin of meteorites, planets and the Earth/Moon system are discussed in Chapter 2. Chapters 3 and 4 dealing with the mantle (and core) and crust, respectively, are no longer strictly descriptive chapters but include also interpretations. Exciting new constraints on mantle structure from seismic tomography and satellite gravity studies are included in Chapter 3, as well as a discussion of convection and the driving forces of plate tectonics. The origin of the core is also examined. In addition to summarizing geophysical and geochemical properties of both the oceanic and continental crust in Chapter 4, metamorphism and crustal provinces are discussed.

Chapter 5 includes not only the basic factual data that led to the seafloor spreading and plate tectonic models, but also includes new information on magnetic reversals, the evolution and changes of plate boundaries, aseismic ridges and oceanic plateaus, and hotspots and plumes. Major tectonic settings are reviewed in Chapter 6 and the section on collisional orogens is expanded, based on the voluminous literature on this subject that has appeared in the past few years. Mineral and energy deposits are also discussed in light of tectonic settings. Chapter 7 has been completely rewritten and updated in light of major advances in our understanding of magma production and mantle sources from trace element, Sr–Nd–Pb isotope, and rare gas studies of basalts. Chapter 8 includes examples of plate tectonic histories of Phanerozoic orogenic systems. Although emphasis is placed on North American systems, examples are also included from other continents.

Chapter 9 includes a discussion of Archean and Proterozoic crustal provinces and of the possible role of plate tectonics during the Precambrian. In Chapter 10, information is brought together to discuss crustal origin and evolution. Topics include the composition and origin of the primitive crust, magma oceans, crustal growth mechanisms and rates, a summary of the growth of North America, secular changes in crustal composition, and comparative evolution of the terrestrial planets. The new Chapter 11 includes a detailed discussion of the origin and evolution of the atmosphere and oceans and comparison of the evolution of the atmospheres of the terrestrial planets. In addition, the evolution of terrestrial climates, the origin and evolution of life, mass extinctions and impact phenomena are discussed.

In order to keep the book to a reasonable length, and avoid duplicating information that is widely available in other books, some subjects are covered in only a cursory manner and others not at all. For instance, the methods by which geological, geochemical, and geophysical data are gathered are only briefly mentioned, as books on these subjects are readily available. Extensive mathematical treatments are omitted for the same reason. Because the book is designed primarily as a textbook, references are kept to a minimum. I have attempted, however, to reference the major papers and some of the minor ones that have strongly influenced me in regard to interpretations set forth in the text. More extensive bibliographies can be found in these papers and in the references listed under "Suggestions for Further Reading" at the end of each chapter.

I am greatly appreciative of efforts of Debbie Pettengill and Pat Mills who typed and edited the manuscript and dedicated many months to producing the finished copy. The superb figures were drafted by Jessica McKinnis.

KENT C. CONDIE

SOCORRO, NEW MEXICO
MAY 1988

CHAPTER 1

Introduction

A Perspective

THE origin and evolution of the Earth's crust is a tantalizing question that has stimulated much speculation and debate from the early part of the 19th century. Some of the first problems recognized—such as how and when did the oceanic and continental crust form?—remain a matter of considerable controversy even today. Results from the lunar landings and satellite data collected from other planets indicate that the Earth's crust may be a unique feature of bodies in the solar system. The rapid accumulation of data in the fields of geophysics, geochemistry, and geology in the past 25 years has added much to our understanding of the physical and chemical nature of the Earth's crust and of the processes by which it evolved. Evidence favors a source for the materials composing the crust from within the Earth. Partial melting of the Earth's interior appears to have produced magmas that moved to the surface and produced the first crust. The continental crust, being less dense than the underlying mantle, has risen isostatically and has been subjected to weathering and erosion. Eroded materials have been added partially to continental margins, causing the continents to grow laterally, and partially returned to the mantle to be recycled and perhaps again to become part of the crust at a later time. Specific processes by which the crust was created and grown are not well known, but a large amount of data allows important boundary conditions to be invoked. In this book, important physical and chemical properties of the crust and upper mantle are presented and discussed in terms of models for crustal origin and evolution.

The theories of seafloor spreading and plate tectonics that have so profoundly influenced geologic thinking in the past two decades have also provided valuable insight into the mechanisms by which the crust has evolved. One of the major problems regarding crustal evolution is that of when plate-tectonic and seafloor spreading processes began. Some scientists consider the widespread acceptance of sea floor spreading and continental drift as a “revolution” in the Earth Sciences (Wilson, 1968). Scientific disciplines appear to evolve from a stage primarily of data gathering, characterized by transient hypotheses, to a stage where a new unifying theory or theories are proposed that explain a great deal of the accumulated data. Physics and chemistry underwent such revolutions around the beginning of the 20th century, whereas the Earth Sciences entered such a revolution in the late 1960s. As with scientific revolutions in other fields, new ideas and interpretations do not invalidate earlier observations. On the contrary, the theories of seafloor spreading and plate

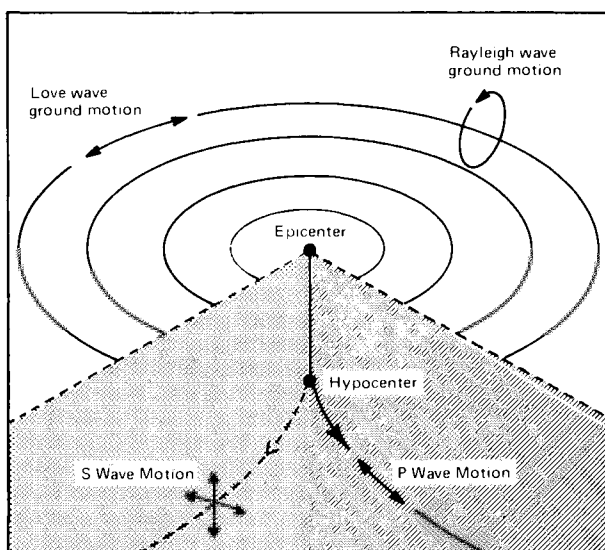


FIG 1.1 Types of body and surface wave motion related to the hypocenter and epicenter of an earthquake. After Davies (1968).

tectonics offer for the first time a unified explanation for heretofore seemingly unrelated observations in the fields of geology, paleontology, geochemistry, and geophysics.

Sources of Data

Seismic Methods

Before discussing the physical and chemical properties of the Earth, it may be useful to review some basic definitions and techniques used in the geosciences. The most definitive information on the structure of the Earth's interior comes from *seismology*. When an earthquake or an explosion occurs in the Earth, two types of elastic waves are produced—body waves and surface waves. *Body waves* travel through the Earth and are reflected and refracted at interfaces. They are of two types: *P waves* (or compressional waves), which are characterized by alternate compression and expansion in the direction of propagation, and *S waves* (or shear waves), with particle motion normal to the direction of propagation (fig. 1.1). *P waves* are always faster than *S waves* and *S waves* cannot be transmitted through a liquid. Surface waves are propagated along or near the surface of the Earth and also are of two types: Rayleigh and Love waves. *Rayleigh waves* exhibit elliptical particle motion confined to a vertical plane containing the direction of propagation, while *Love waves* are characterized by horizontal motion normal to the propagation direction. The region in the Earth where elastic waves are produced by an earthquake (or explosion) is defined as the *hypocenter* or *focus*, and the point on the Earth's surface vertically above as the *epicenter*. Elastic waves are detected by seismometers, which respond to ground movements. Computerized arrays of seismometer stations make it possible to separate interfering signals, to improve signal-to-noise ratio, and to measure wave velocities directly.

Several seismic methods are used in investigating the interior of the Earth. The

gross features of the Earth's interior are determined from travel-time distance studies of body waves travelling through the Earth. Detailed structure of the crust and uppermost mantle is determined by seismic refraction and reflection methods. Large underground explosions are particularly useful in these studies because the time and location of such explosions are known more accurately than earthquake times and hypocenter locations. The refraction method, which is used both on land and sea, is based on measuring the travel times of P waves between shot points and seismic recorders located various distances apart, usually along straight-line profiles. The method is limited, in that very detailed crustal structure cannot be determined. Evidence for low-velocity layers is obtained from modeling of surface-wave data and from amplitude studies of refraction data. The use of supercritical reflections (i.e. reflected waves that have incident angles greater than the critical angle) can enhance the interpretation of refraction data. Vertical incidence reflections occur only at sharp discontinuities and may allow a distinction to be made between sharp and gradational discontinuities.

Travel-time anomaly studies are valuable in evaluating upper-mantle structure. A *travel-time anomaly* (or residual) is the difference between observed and calculated body wave arrival times at a given seismograph station. Calculated arrival times are azimuthally corrected and based on idealized models. Maps constructed by contouring travel-time anomalies are useful in relating such anomalies to geological and other geophysical features.

Earthquakes produce natural vibrations in the Earth known as *free oscillations*. Two types of oscillations occur: torsional oscillations, involving particle displacements normal to the Earth's radius, and spheroidal oscillations, which are radial or tangential displacements. Long-period free oscillations are detected with strain seismometers and Earth-tide gravimeters. Free-oscillation studies have resulted in improved resolution and detection of interfaces within the Earth, as well as determination of density and seismic anelasticity of parts of the mantle. Rayleigh- and Love-wave dispersion provide a basis for detailed studies of crustal and upper-mantle structure.

Free oscillations produced by major earthquakes do not last indefinitely and the vibrational energy is gradually converted to heat. The oscillations are attenuated, and the process is known as *anelasticity*. Body waves passing through the Earth are also attenuated. Anelastic attenuation is measured with a unitless factor Q , the *specific attenuation factor*, and low values of Q mean high seismic-wave attenuation. Measured Q values in the Earth range from about 10 to greater than 1000. Anelasticity in the Earth appears to result from some combination of grain boundary damping, stress-induced ordering of crustal defects, and damping caused by vibration of dislocations (Gordon and Nelson, 1966). Q decreases rapidly as temperature and degree of melting increase in the Earth.

Seismic reflection profiling is used both in oceanic and continental areas (Brewer and Oliver, 1980; Stommel and Graul, 1978). In oceanic areas, acoustic sources on ships are used to produce energy to study the stratigraphy and structure of sediments around continental margins and in ocean basins. The major reflection profiling studies on the continents are those of the COCORP research group in the United States and the BIRPS group in the United Kingdom. Vibrating trucks transmit energy into the Earth, and return echoes are recorded by arrays of

geophones. Data are collected and computer processed so as to produce a section through the crust as a function of seismic-wave travel times.

Magnetic Methods

The Earth's magnetic field is defined by its strength and direction. The direction is expressed in terms of the horizontal angle between true north and magnetic north, the *declination*, and the angle of dip with the horizontal, the *inclination*. The inclination becomes vertical at the two magnetic poles. The total magnetic field strength is strongest near the magnetic poles ($70\ \mu\text{T}$ at the South Pole) and weakest at the equator (about $30\ \mu\text{T}$). Both short- and long-term variations occur in the direction and strength of the magnetic field. Short-term variations (with periods of hours to years) are related chiefly to interactions of the magnetic field with the strongly conducting upper layers of the atmosphere. Variations with periods of hundreds of years or more are known as secular variations and are interpreted to support an origin for the magnetic field in terms of fluid motions in the outer part of the Earth's core. Approximately 90% of the present field can be described by a magnetic dipole at the Earth's center, which makes an angle of about 11.5° with the rotational axis. A general westward drift of the field is noted at a rate of about $0.18\ \text{deg/a}$.

Local and regional variations in the magnetic field reflect, for the most part, rocks beneath the surface with varying degrees of magnetization. These variations are measured with fluxgate or proton magnetometers on land or sea, or in the air. Significant deviations from a magnetic background either on a local or regional scale are known as magnetic anomalies, the intensities of which are expressed in gammas (γ) or teslas (T) ($1\ \gamma = 100\ \mu\text{T}$). Small-scale anomalies extending over thousands of square kilometers reflect variations in the lower crust or upper mantle.

Rocks may become magnetized in the Earth's magnetic field by several mechanisms, which are described in Chapter 5. Such magnetization is known as *remanent magnetization* and is measured in the laboratory with spinner, astatic, or cryogenic magnetometers. The maximum temperature at which a mineral can possess remanent magnetization is known as the *Curie temperature*. *Paleomagnetism* is the study of remanent magnetism in rocks of various geologic ages. If rock samples can be accurately oriented and the date of magnetization determined, it is often possible to determine the locations of earlier magnetic pole positions. Paleomagnetic studies have shown that the magnetic poles have reversed themselves many times in the geologic past, and such reversals are thought to be produced by instability in the outer core.

Gravity Methods

Gravity is the force of attraction between the Earth and a body on or in the Earth divided by the mass of the body. The average gravitational force of the Earth is $980\ \text{gals}$ ($1\ \text{gal} = 1\ \text{cm/sec}^2$). Gravity is measured with a gravimeter and can be determined both on land and at sea. Accuracies are typically about $1\ \text{mgal}$ on land and $5\text{--}10\ \text{mgal}$ at sea. The standard reference for gravity on the Earth is the gravitational field of a spheroid, and is dependent only on latitude. The gravity field

on the Earth can be described using data derived from the directions and rates of the orbital shift of artificial satellites. From such data it is possible to determine how much the Earth's average surface, or *geoid*, which is roughly equal to sea level, actually deviates from a spheroid. Data indicate that the Earth is pear-shaped, with an average equatorial radius of 6378 km and an average polar radius of 6357 km. Gravity distribution on the Earth can be calculated from spherical harmonic coefficients of the satellite gravitational data.

Local and regional gravity data must be corrected for latitude and elevation before interpretation. On land, gravity measurements are usually above the geoid surface, and hence an increase in gravity must be added to the observed value to account for the difference in elevation. This is known as the *free-air correction*. If the standard gravity value of the spheroid is now subtracted (i.e., the latitude correction), the *free-air anomaly* remains. If the attraction of rock between the geoid and the gravity station is subtracted (the *Bouguer correction*) and a correction is made for nearby topographic variations, we obtain the *Bouguer anomaly*. Measurements at sea require no free-air correction, since they are made at sea level, and the Bouguer correction, where used, is added to account for the change in gravity that would result if the oceans were filled with rock instead of water.

Early gravity measurements in the mid-1700s indicated that large mountain ranges exhibit smaller-than-expected gravitational attractions. Such data led to the principle of *isostasy*, introduced about 1900 by Dutton. This principle suggests that an equilibrium condition exists in the Earth whereby the load pressure due to overlying columns of rock is equal at some depth of compensation. Two main theories have been proposed to explain isostasy. Pratt's theory assumes that the density of rock columns in the outer shell of the Earth varies laterally above a constant depth of compensation and is expressed as a function of elevation on the Earth's surface. Airy's theory proposes that the outer shell is composed of low, rather constant-density columns, and that the depth of compensation varies as a function of the thickness of the columns. Both mechanisms probably contribute to isostatic compensation. Models suggest compensation depths of the order of 50–100 km for both the Airy and Pratt theories. *Isostatic gravity anomalies* may be calculated by subtracting from Bouguer anomalies the mass distribution within a segment of the upper part of the Earth as determined from some combination of the Airy and Pratt compensation mechanisms.

Electrical Methods

The Earth's magnetic field induces electrical currents, known as *telluric currents*, which flow in the crust and mantle. Most short-period variations in the magnetic field are produced by interactions with the strongly conducting ionosphere (upper atmosphere). A magnetic storm produces large magnetic variations lasting for a few days and is caused by strong currents of high-energy particles emitted by solar flares that are trapped in the ionosphere. Magnetic variations can be used to estimate conductivity in the Earth, since the strength of induced currents depends on electrical conductivity distribution. Short-period variations of such currents penetrate only to shallow depths, while longer periods penetrate to greater depths.

Four methods have been used to estimate conductivity distribution in the crust

and mantle (Keller, 1971; Creer, 1980): (a) direct-current sounding; (b) magnetotelluric sounding; (c) electromagnetic sounding; and (d) geomagnetic deep-sounding. Direct-current sounding involves driving a current into the ground between widely spaced electrodes, and the depth of penetration of this method is limited to only several tens of kilometers. In the magnetotelluric method, both electric and magnetic variations in the Earth's field are measured simultaneously. An artificial electromagnetic field is generated, driven into the Earth, and measured in the electromagnetic method. The geomagnetic deep-sounding method involves measuring variations of naturally induced currents caused by magnetic storms. This provides the best method for estimating mantle conductivity distributions.

Geothermal Methods

Heat flow determinations on the Earth involve two separate measurements, one of the thermal gradient (dT/dx) and one of thermal conductivity (K). From these measurements, heat flow (q) is calculated as follows:

$$q = K \frac{dT}{dx} \quad (1.1)$$

Heat flow may be expressed as $\mu\text{cal}/\text{cm}^2 \text{ sec}$ or as mW/m^2 where $1 \mu\text{cal}/\text{cm}^2 \text{ sec}$ is defined as one heat flow unit (1 HFU) and $1 \text{ HFU} = 0.0239 \text{ mW}/\text{m}^2$. Thermal gradient is measured with thermistors, which on land are attached to a cable and lowered down a borehole, and at sea are attached to core barrels or mounted in a long thin probe that is inserted into deep-sea sediments. In both cases, time is allowed for thermal equilibration before measurements are taken. Thermal conductivity of water-saturated rocks is usually measured with a divided-bar apparatus in which a known heat flow is passed through a sandwich of copper discs, two standards, and a rock sample; thermal conductivity is calculated from the temperature difference across the sample and its thickness. The thermal conductivity of unconsolidated sediments is usually measured with a needle probe, which consists of a thermistor, an electrical heating element, and a hypodermic needle inserted into the sediment. Thermal conductivity is obtained from the rate at which the needle temperature rises for a given energy input to the heater.

In continental areas, significant ground water movement can produce anomalously low heat flow. Also, measured heat flow in areas that were covered by Pleistocene glaciers may be lower than actual heat flow. Although glacial corrections up to 30% have been proposed by some investigators, evidence is conflicting regarding the general importance of this effect.

The radiogenic heat production of a rock or of a geologic terrane may be calculated from the concentrations of U, Th, and K and the heat productivities of ^{235}U , ^{238}Th , and ^{40}K . The concentrations of these isotopes can be determined by counting the natural radioisotopes with a gamma-ray spectrometer in the laboratory. Airborne gamma-ray spectrometers have been used to estimate concentrations of U, Th, and K over large areas of the crust. Radiogenic heat generation (A) is expressed in $10^{-13} \text{ cal}/\text{cm}^3 \text{ sec}$ or as $\mu\text{W}/\text{m}^3$. One heat generation unit (1 HGU) is defined as $10^{-13} \text{ cal}/\text{cm}^3 \text{ sec}$ and is equivalent to $0.0239 \mu\text{W}/\text{m}^3$.

High-pressure Studies

For many years it has been possible to reconstruct in the laboratory static pressures up to about 300 kbar, which is equivalent to about 1000 km burial depth in the Earth. A new era of high-pressure research began in 1972 with the development of the double-stage split-sphere apparatus and the diamond-anvil pressure cell (Liu and Bassett, 1986). With these systems it is possible to study phase relations at pressures up to 1.7 Mbar and temperatures up to 3500°C, which allows direct investigation of lower mantle and core compositions. High-pressure experiments can be performed with solid or liquid media at a large range of temperatures. It is also possible to measure a considerable number of properties of rocks at high pressures and temperatures: phase equilibria boundaries, elastic properties including P- and S-wave velocities, electrical and thermal properties, and fracture and flow characteristics are but a few. From such measurements, in conjunction with geophysical data, it is possible to place limitations on the composition, mineralogy and melting behavior of the crust and upper mantle, and to evaluate the origin of magmas. From the results of high-pressure and high-temperature rock-deformation studies, it is also possible to understand more fully earthquake mechanisms and flow characteristics within the Earth.

Possible mineral assemblages and compositions of deeper parts of the Earth, including the core, also can be studied using the results of shock-pressure experiments. The method involves generating a strong shock (up to several megabars) in a material with explosives, producing a wave front that moves through the material at a velocity greater than sound and greater than the particle velocity of the shocked material (Ahrens and Peterson, 1969). The pressure and density within the wave can be deduced by measuring the shock and particle velocities, and results are generally expressed in terms of the hydrodynamic sound velocity plotted against density. Various elements, minerals, and rocks are examined, and the results are compared with hydrodynamic velocity data deduced from body-wave studies of the Earth. Such comparisons provide limitations on the composition of the lower mantle and core.

Geochemistry and Geochronology

Geochemical data from rocks and minerals provide important information bearing on the composition of the upper mantle and evolutionary changes in the crust and mantle. Geochemical and isotopic research have advanced rapidly in the past decade in response to the development of new analytical methods and geochemical modeling. Trace element and radiogenic isotope geochemistry, in particular, have been useful in studying the evolution of the crust and mantle. Important advances in our understanding of planetary evolution also have come from geochemical and isotopic studies of lunar and meteorite samples and of ultramafic inclusions from the Earth's mantle. Isotopic studies are important not only in terms of geochronology but also for identifying and mapping crustal and upper-mantle sources for magmas. Pb, Nd, and Sr isotopes are important in both applications. Geochronology involves the study of time relationships in orogenic belts and in the evolution of continents and ocean basins. Tracer studies make use of daughter isotopes as "fingerprints" to

study the origin of igneous rocks and to trace the evolution of the mantle and crust through geologic time (see Chapt. 7).

Refinements in radiometric dating methods have improved estimates of the beginning and duration of the various subdivisions of geologic time. A current version of the geologic time scale is given in table 1.1.

Other Sources of Information

The viscosity of the mantle has been estimated by studies of isostatic recovery rates of large segments of the crust after removal of a surface load such as icecaps or large lakes, and from estimates of the seismic anelasticity Q . The mass of the Earth can be estimated from surface gravity data after a rotational correction. The Earth's two principal moments of inertia—one about the polar axis and the other about an equatorial axis—can be estimated from rotational axis precessional data and the observed flattening of the Earth. Other physical properties as a function of depth within the Earth are estimated from measurements made on the Earth's surface and models of the Earth's interior.

Information from oceanic and continental drill cores allows a reliable projection of compositional data to shallow depths in the crust. The Deep Sea Drilling Project (DSDP), which began in 1968, has now recovered many cores from the sediment layer on the ocean floors, some up to several hundred meters in length. A specially designed drilling ship, the *Glomar Challenger*, is used as a floating drilling platform. Deep holes on the continents, other than oil wells, are rare. However, deep drilling into the continents in a variety of geologic environments is currently in the planning stages.

Last but not least are the conventional and well-established geological methods. Perhaps the most commonly overlooked yet extremely important source of data is field geology. The results of widespread geological mapping on the continents are of critical importance to the evaluation of the roles of seafloor spreading and plate tectonics in the geologic past. Stratigraphy, tectonics, volcanology, experimental petrology, sedimentation, and paleontology are other important fields of investigation.

Seafloor Spreading

Seafloor spreading is the process by which the lithosphere splits at ocean ridges and moves away from ridge axes with a motion like that of a conveyor belt as new lithosphere is created and fills the resulting crack. The lithosphere can be considered as a mosaic of plates bounded by ocean ridges, subduction zones, and transform faults (boundaries along which plates slide by each other) (fig. 1.2), and the study of the interactions of lithospheric plates is known as *plate tectonics*. Oceanic lithosphere is consumed in the asthenosphere at subduction zones to accommodate the newly created lithosphere such that the surface area of the Earth remains constant. The alternate possibility, that the Earth is expanding to accommodate growth of the lithosphere, seems unlikely, as discussed later. Hess (1962) is credited with proposing the theory of seafloor spreading, although the name was suggested by Dietz (1961).

The most definitive evidence for seafloor spreading comes from the study of linear magnetic anomalies on the seafloor (Chapt. 4). Vine and Matthews (1963) first

TABLE 1.1

GEOLOGIC TIME SCALE																														
CENOZOIC					MESOZOIC					PALEOZOIC					PRECAMBRIAN															
AGE (Ma)	PERIOD	EPOCH	AGE		AGE (Ma)	PERIOD	EPOCH	AGE		AGE (Ma)	PERIOD	EPOCH	AGE		AGE (Ma)	EON	ERA													
5 10 15 20 25 30 35 40 45 50 55 60 65	QUATERNARY TERTIARY	PLIOCENE	L	CALABRIAN	70 80 90 100 110 120 130 140 150 160 170 180 190 200 210 220 230 240	TRIASSIC	LATE	NORIAN	520 540 560	CAMBRIAN	LATE	TREMPERALEAN FRANCONIAN DRESBACHIAN	3250 3500 3750	MIDDLE	EARLY		ARCHAEO	EON	ERA											
																				PALEOCENE										
		Eocene	M	LUTETIAN			EARLY	PLEISTOCENE	500		EARLY	TREMADOCIAN		MIDDLE	EARLY															
		Oligocene	E	RUPELIAN		LATE	KIMMERIDGIAN	400	420	DEVONIAN	MIDDLE	GIVETIAN	2000	EARLY	SIEGENIAN	2250														
		Neogene	L	CHATTIAN		AQUITANIAN		NEOCENE	HAUTERIVIAN	360	380	MISSISSIPPIAN	LATE	FRASNIAN	1750	EARLY	TOURNAISIEN	1500												
		Miocene	M	SERRAVALLIAN		LANGHIAN		EARLY	ALBIAN	320	340	MISSISSIPPIAN	LATE	BASHKIRIAN	1250	MIDDLE	SERPUKHOVIAN	1000												
		Pliocene	L	ZANCLEAN		MESSINIAN		LATE	CAMPAÑIAN	280	260	PERMIAN	EARLY	SARMATIAN	750	LATE	KAZANIAN	500												
		Quaternary	E	CALABRIAN		PIACENZIAN		LATE	MAASTRICHTIAN	70	260	PERMIAN	LATE	TATARIAN	750	LATE	KAZANIAN	500												
												PROTEROZOIC					ERA													
												ARCHAEO					EON													

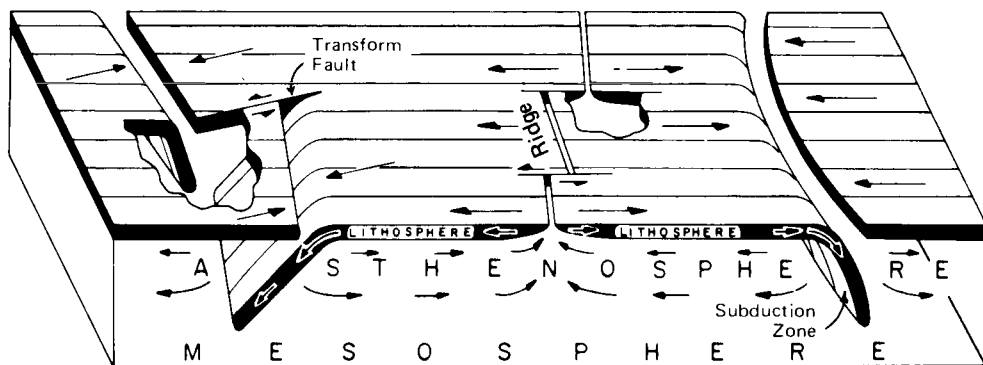


FIG 1.2 Schematic diagram showing the major features of seafloor spreading. After Isacks et al. (1968), copyright by the American Geophysical Union.

suggested that the alternate stripes of positive and negative magnetic anomalies were caused by bands of basaltic rock in the oceanic crust that were alternatively magnetized in normal and reversed directions of the Earth's magnetic field. They proposed that new lithosphere is formed over convective upcurrents beneath ocean ridges by magmatic processes. As magmas emplaced in the axial rifts cool through the Curie temperature of magnetic minerals, they acquire a magnetization in the direction of the existing magnetic field of the Earth. The newly magnetized crust then splits and is forced apart to make room for fresh injections of magma.

Many seemingly unrelated facts from the fields of geology, paleontology, geophysics, and geochemistry are consistent with, and find a unified explanation in, the seafloor spreading model. For instance, the observed increase in age and thickness of deep-sea sediments with distance from ocean ridges is predicted by the model, as are the compositional changes found in basalts retrieved from the oceanic crust. First-motion studies of earthquakes (see Chapt. 5) are also consistent with the model. The absence of deep-sea sediments older than Jurassic reflects seafloor spreading rates of a few centimeters per year, which indicates that the ocean floors should be completely renewed every 150–200 Ma. Satellite laser measurements verify plate motions (Christodoulidis et al., 1985) and results agree in magnitude and direction with those deduced from seafloor spreading reconstructions. Data indicate that Hawaii on the Pacific plate and locations on the North American plate are separating at about 1 cm/a, Hawaii and Australia are converging at a rate of 4.6 cm/a and Australia and North America are closing on each other at about 1.8 cm/a.

Continental Drift

Although continental drift was first suggested in the 17th century it did not receive serious scientific investigation until the beginning of the 20th century. Wegener (1912) is usually considered the first one to have formulated the theory precisely. In particular, he pointed out the close match of opposite coastlines of continents and the regional extent of the Permo-Carboniferous glaciation in the Southern Hemisphere. DuToit (1937) was the first to propose an accurate fit for the continents based on geological evidence. Continental drift, however, did not receive wide acceptance among geoscientists, and especially among Northern-Hemisphere geoscientists, until the last two decades. The chief reasons, aptly summarized by MacDonald

(1964), related to lack of an acceptable mechanism and to geophysical data that indicate that continents have roots in the mantle (Chapts. 3 and 4) and hence could not be moved at the Moho, as was commonly proposed. The breakthrough came in the early 1960s, when the seafloor spreading theory became widely accepted. Seafloor spreading offers a mechanism for continents to drift and yet retain their deep roots, by drifting at the base of the lithosphere, not at the Moho.

It is possible, using a variety of geological and geophysical data, to reconstruct the positions of continents prior to their last break-up about 200 Ma ago. Matching of continental borders, stratigraphic sections, and fossil assemblages are some of the earliest methods used to reconstruct continental positions. Today, in addition to these methods, we have polar wandering paths, seafloor spreading directions, and hotspot traces. The geometric matching of continental shorelines was one of the first methods used in reconstructing continental positions. Both Wegener (1912) and DuToit (1937) employed this method. Later studies have shown that matching of the continents at the edge of the continental shelf or continental slope results, as it should, in better fits than the matching of shorelines, which reflect chiefly the flooded geometry of continental margins. The use of computers in matching continental borders has resulted in more accurate and objective fits. As an example, figure 1.3 shows a computerized, least-square fit of continents on both sides of the Atlantic Ocean at 500 fathoms depth. Areas of overlap and gaps are minimal. Many areas of overlap can be explained by additions to the edges of the continents after their rupture. Examples are the Niger Delta and the carbonate reefs of the Bahama Bank and Blake Plateau. Similar computerized fits have been proposed for Australia, Antarctica, India, Madagascar, and Africa.

Similarities in age and stratigraphic sections on opposite continental margins is another feature used by early investigators to support continental drift. One of the most striking similarities is that of the late Paleozoic (Gondwana) sections on continents in the Southern Hemisphere. When continents in the Southern Hemisphere are considered in their predrift positions, late Precambrian–Paleozoic terranes define a continuous belt from eastern Australia through Antarctica into southern Africa and Argentina. A belt of Triassic and Jurassic basalts and diabases of similar composition, which extends from South Africa through Antarctica into Tasmania, and a clustering of Precambrian anorthosites in East Africa, Madagascar, and India also support the geometric fit of the continents in the Southern Hemisphere. Radiometric dating of Precambrian terranes bordering the Atlantic Ocean provides strong evidence for continental drift. The fit of South America to Africa suggested by the matching of crustal provinces and structural trends is very close to the 500-fathom computerized fit of these continents shown in figure 1.3. Most striking is the continuation of the late Precambrian provinces from west-central Africa into eastern South America. Also, the Archean provinces in western and central Africa continue into South America. Precambrian provinces around the North Atlantic are considerably more complex and appear to involve extensive reworking of older terranes.

The Expanding Earth Hypothesis

An alternate explanation for some of the evidence of continental drift is that the Earth has expanded, with new surface area created at ocean ridges (Carey, 1976).

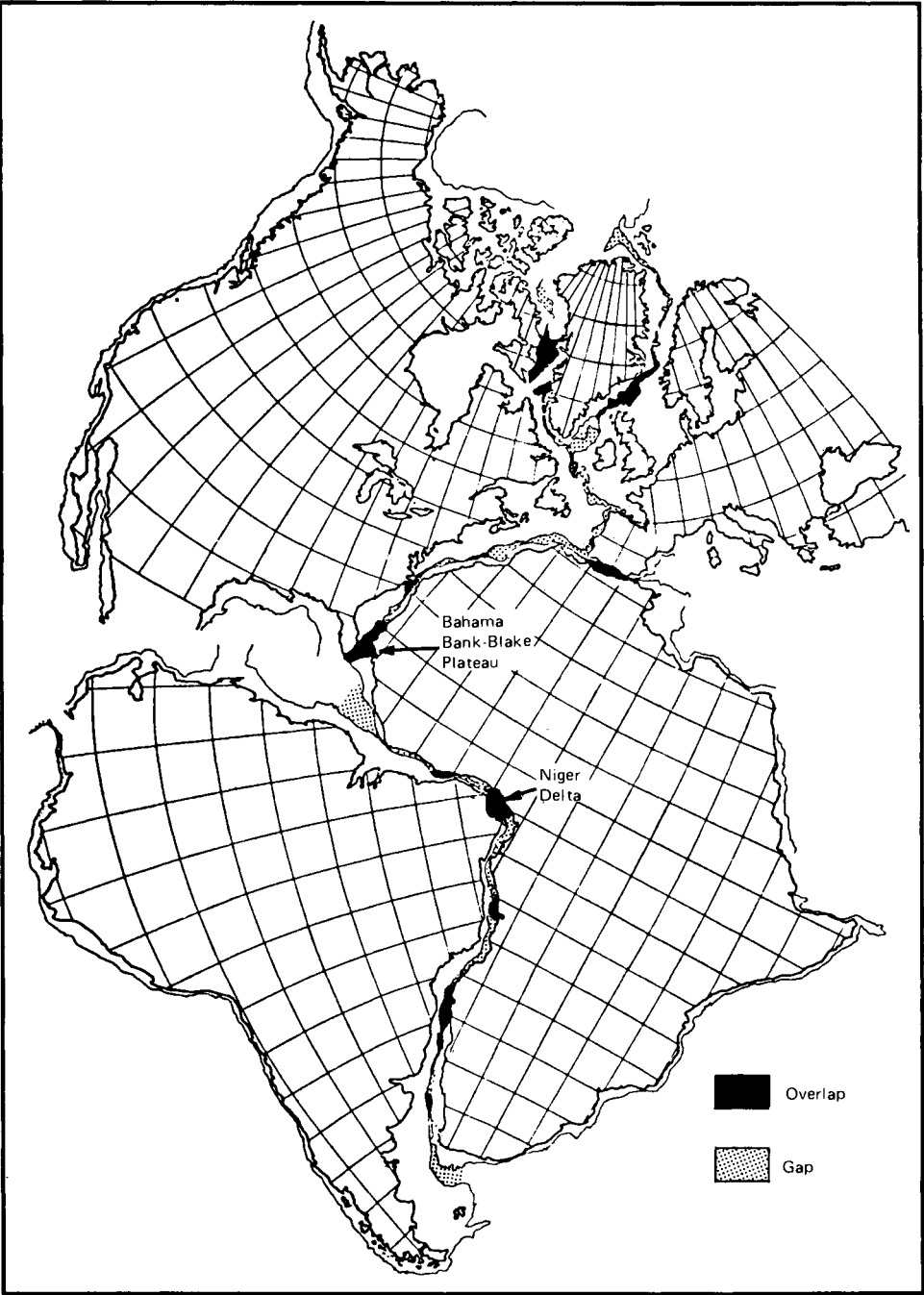


FIG 1.3 Computerized least-square fit at 500 fathoms of continents around the Atlantic Ocean. After Bullard et al. (1965). Used with permission of the Royal Society of London.

This hypothesis generally presumes that the area of the continents has remained constant and that ocean basins have grown as the Earth expands. An expanding Earth and seafloor spreading are not mutually exclusive processes, however, and the overwhelming evidence for seafloor spreading and continental drift seems to necessitate that, if the Earth expanded, these processes have gone on concurrently.

Evidences for an expanding Earth are either ambiguous or are based on tenuous and ad hoc assumptions. One of the earliest evidences cited is the idea that the continents have become progressively emergent with time, necessitating an increase in the Earth's radius of about 0.5 mm/a (Egyed, 1956). These results, based on data collected from paleogeographic maps, suggest that the percentage of the continents flooded by shallow seas has decreased with time. Some investigators have interpreted this to indicate that the Earth is expanding and causing seawater to move off the continents into growing ocean basins. However, several sources of inherent error render this interpretation uncertain. First, the continents are only 88% emergent today, and some seawater is tied up in polar ice caps. Also, determining the emergent area of continents by scaling data from paleogeographic maps has many sources of error, including inadequate maps for most continents. Estimates of the areas covered by seawater as a function of time in North America indicate a rather constant relationship between average elevation of the continents and sea level since the Cambrian (Wise, 1973). It is also likely that the volume of ocean ridges, which has changed with time, is the most important factor in controlling flooding of the continents. Data are not sufficient to document progressive emergence (or submergence) of continents with time.

Carey (1958, 1976) has discussed a variety of geometric arguments for an expanding Earth. For instance, large gaps are left on some reconstructions of the Permian supercontinent Pangaea, which disappear on a globe of smaller radius. Such problems, however, may or may not be real, in that other investigators have managed to obtain acceptable reconstructions. Glikson (1980) has suggested that the lack of evidence for extensive oceanic crust in the Proterozoic can be explained by having an Earth with a smaller radius and Proterozoic supercontinents covering nearly the entire surface.

Major obstacles to an expanding Earth, however, seem to outnumber evidences in favor of expansion (McElhinny et al., 1978; Schmidt and Clark, 1980). New estimates of the radius of the Earth during the last 400 Ma from paleomagnetic results limit the amount of expansion over this time to $<0.8\%$. Also, crater distributions and ages on the lunar surface rule out any significant expansion of the Moon ($<0.06\%$) during the last 4 Ga, and a similar conclusion applies to Mars, Mercury and Venus. By analogy, the Earth would not be expected to have undergone significant expansion during the same time period. Biogeographic distributions of invertebrates during the early Paleozoic are also incompatible with an expanding Earth, in that they demand large expanses of oceanic crust between continental blocks. It also has not as yet been possible to find a satisfactory mechanism for planetary expansion. Results from the Moon seem to limit a decrease in the gravitational constant over the last 4 Ga to $<2 \times 10^{-10} \text{ a}^{-1}$, which eliminates one of the more popular causes suggested for expansion. Considering all sources of data, it would appear that the Earth has not expanded more than 1% since its formation 4.6 Ga ago.

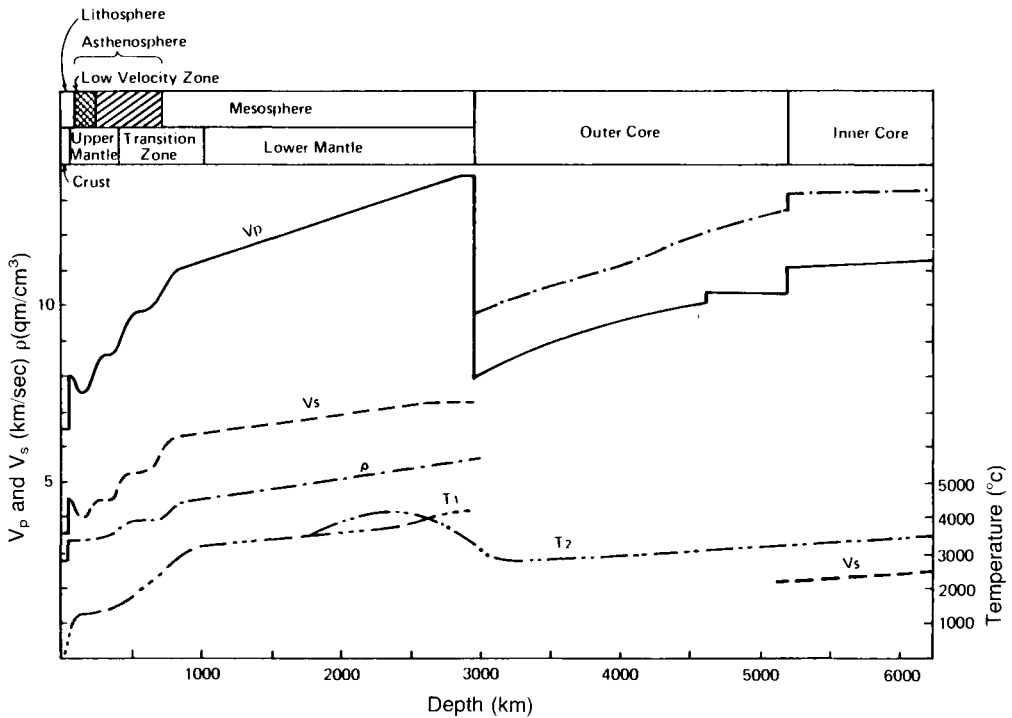


FIG 1.4 P-wave (V_P), S-wave (V_S), density (ρ) and temperature (T_1 , T_2) distributions in the Earth.

Variation of Physical Properties in the Earth

Internal Structure

The internal structure of the Earth is revealed primarily by body-wave studies and by free-oscillation data. The variation of P- and S-wave velocities and density in the Earth reflects changes in pressure, temperature, mineralogy, composition, and degree of partial melting. Although the overall features of seismic-wave velocity distributions have been known for some time, refinement of data has been possible in the last ten years. A model consistent with the Earth's mass, moment of inertia, periods of free oscillation and travel times of body waves is given in figure 1.4. Three major first-order discontinuities occur: the largest, at 2900 km, is the *core-mantle interface*; at 10–12 km beneath the oceans and 30–50 km beneath the continents is the *Mohorovicic discontinuity*, or *Moho*; and at about 5200 km is the *inner-core-outer-core interface*. The Moho separates the Earth's crust from the mantle, and the core composes about 16% of the Earth by volume and 32% by mass. The discontinuities reflect changes in composition or phase, or in both. Smaller velocity change provides a basis for further subdivision of the mantle, as discussed in Chapter 3.

The major regions of the Earth can be summarized as follows with reference to figure 1.4.

1. The *crust* consists of the region above the Moho, and ranges in thickness from about 3 km at some oceanic ridges to about 80 km in collisional orogens.

2. The *upper mantle* extends from the Moho to 400 km, and includes the lower part of the lithosphere and the upper part of the asthenosphere. The *lithosphere* (50–200 km thick) is the strong outer layer of the Earth, including the crust, that reacts to stresses as a brittle solid. The *asthenosphere*, extending from the base of the lithosphere to about 700 km, is by comparison a weak layer that readily deforms by creep. A region of low seismic-wave velocities and high attenuation (low Q), the *low-velocity zone* (LVZ), generally occurs at the top of the asthenosphere and is from 50 to 100 km thick. Significant lateral variations in density and in seismic-wave velocities are common at depths of less than 400 km.
3. The *transition zone*, which extends from 400 to 1000 km, is characterized by three rather rapid increases in velocity that are considered in more detail in Chapter 3. Significant lateral variations in S-wave velocities appear to extend to about 1000 km.
4. The *lower mantle* extends from about 1000 km to the 2900 km discontinuity. For the most part, it is characterized by rather constant increases in velocity and density until just above the core–mantle interface, where a flattening of velocity and density gradients occurs. The lower part of the transition zone and the lower mantle are collectively referred to as the *mesosphere*, a region that is strong and is relatively passive in terms of deformational processes.
5. The outer core will not transmit S waves and is interpreted to be liquid. It extends from 2900 km to the discontinuity at 5200 km.
6. The inner core, which extends from 5200 km to the center of the Earth, transmits S waves, although at very low velocities, suggesting that it is near the melting point.

Gravity distribution can be calculated from the density profile and the pressure variation with depth can be determined from density and gravity distributions. Gravity changes very little to the core–mantle discontinuity and falls off rather rapidly thereafter. Pressure increases slowly to about 1.5 Mbar at this discontinuity and rapidly thereafter, to about 3 Mbar at the 5200 km discontinuity; the pressure in the inner core is about 3.5 Mbar.

Temperature Distribution

Considerable uncertainty exists regarding the temperature distribution in the Earth. It is dependent upon such features of the Earth's history as (1) the initial temperature distribution, (2) the amount of heat generated as a function of both depth and time, and (3) the process of core formation. Most estimates of the temperature distribution in the Earth are based on one or a combination of three approaches:

1. models of the Earth's thermal history involving various mechanisms for core formation;
2. variations of seismic-wave velocities and probable variations of electrical conductivity, thermal conductivity, and other physical and chemical properties with depth;

3. models involving redistribution of radioactive heat sources in the Earth by melting and convection processes.

Estimates using various models seem to converge on a temperature at the core–mantle interface of $3500 \pm 500^\circ\text{C}$.

Two examples of calculated temperature distributions in the Earth are shown in figure 1.4. Tozer (1959) calculated the temperature distribution in the mantle using electrical conductivity data. The results are shown as curve T_1 and indicate a rapid temperature increase in the lithosphere, a flattening of gradient in the asthenosphere, and a moderate increase to the base of the transition zone. The lower mantle is characterized by a steady but gradual increase in temperature. Other models indicate that the process of core formation and fractionation of radioactive heat sources in the Earth by convection and melting are critical in estimating the Earth's temperature distribution. If the core formed during or soon after the accretion of the Earth, it should absorb gravitational energy; hence, a maximum in the temperature distribution may occur in the lower mantle, as shown by curve T_2 . Movement of radioactive heat sources into the upper regions of the mantle early in the Earth's history would have the effect of producing a temperature maximum at even shallower depths in the mantle.

Summary Statements

1. Three first-order seismic discontinuities occur in the Earth: the Moho (10–50 km), the core–mantle interface (2900 km), and the 5200-km discontinuity. These discontinuities define the crust, mantle, outer core and inner core, in order of increasing depth.
2. In terms of strength and mode of deformation the Earth is divided into the lithosphere (extending from the surface to 50–200 km depth), a strong outer layer that behaves as a brittle solid; the asthenosphere (extending from the base of the lithosphere to about 700 km), a region of low strength with a low-velocity zone in the upper part; and the mesosphere, a probably strong and homogeneous region extending to the base of the mantle.
3. A transition zone characterized by three rapid velocity changes occurs from 400 to 1000 km depth, and separates the upper from the lower mantle.
4. Temperature increases in the Earth most rapidly in the transition zone and reaches a possible maximum in the lowermost mantle. The probable temperature at the core–mantle interface is $3500 \pm 500^\circ\text{C}$.

Suggestions for Further Reading

- Brown, G. C., and Mussett, A. E. (1981) *The Inaccessible Earth*. London: Allen and Unwin, 235 pp.
 DeBremaecker, J. C. (1985) *Geophysics: The Earth's Interior*. New York: J. Wiley, 342 pp.
 Liu, L., and Bassett, W. A. (1986) *Elements, Oxides, and Silicates: High-Pressure Phases with Implications for the Earth's Interior*. Oxford: Oxford Univ. Press, 250 pp.
 Mason, B., and Moore, C. B. (1982) *Principles of Geochemistry*, fourth ed. New York: J. Wiley, 352 pp.
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CHAPTER 2

Origin of the Earth–Moon System

Introduction

As members of the solar system, the origin of the Earth and Moon must be considered within the broader framework of planetary origin. Most data favor an origin for the planets by condensation and accretion of a gaseous solar nebula in which the Sun forms at the center. Considering that the age of the universe may be more than 20 Ga, the formation of the solar system at about 4.5 Ga is a relatively recent event. Since it is not possible as yet to observe planetary formation in other gaseous nebulae, we depend upon a variety of indirect evidence to reconstruct the conditions under which the planets in the solar system formed. Geophysical and geochemical data provide the most important constraints. Because the interiors of planets are not accessible for sampling, we often rely on meteorites to learn more about planetary interiors. Most meteorites appear to represent fragments of small planets from the asteroid belt between Mars and Jupiter. Hence, before discussing planetary origin, it is important to review some of the important features of meteorites and to discuss their origin.

Meteorites

Meteorites range in size from dust particles to bodies hundreds of meters across. Those with masses more than 500 gm fall on the Earth at a rate of about one per 10^6 km² per year. Calculated meteorite orbits from photographic networks on the Earth's surface indicate sources in the asteroid belt. A small proportion of meteorites, however, may come from the surfaces of the Moon or Mars, as indicated by geochemical data from meteorite finds. Meteorites are classified as *stones*, including chondrites and achondrites, *stony-irons* and *irons*, depending on the relative amounts of silicates and Fe–Ni metal phases present. *Chondrites*, the most widespread meteorites, are composed in part of small silicate spheroids known as *chondrules* (fig. 2.1), and have chemical compositions similar to the Sun. One group of chondrites, the *Type 1 carbonaceous chondrites* (C1) contain hydrous phases and carbonaceous compounds and are thought to approximate the composition of the primitive solar nebula from which the planets formed. *Achondrites*, which lack chondrules, commonly have igneous textures and appear to have crystallized from magmas. Some achondrites are breccias that probably have formed on asteroid surfaces by impact. Stony-irons and irons have textures and chemical compositions

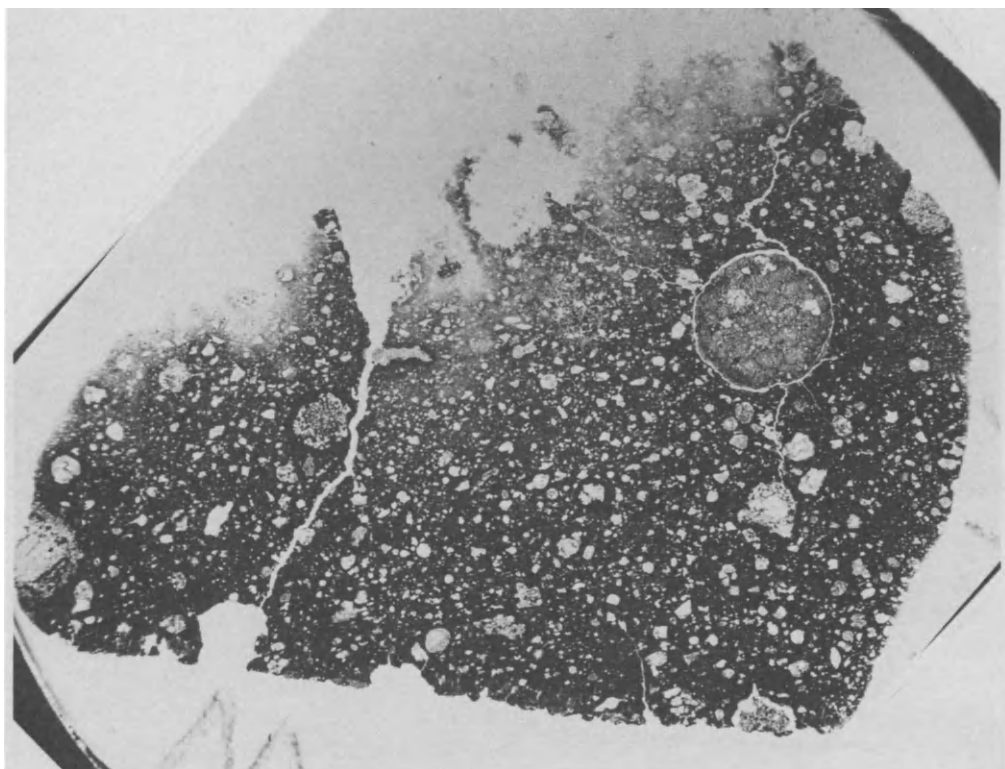


FIG 2.1 Thin section of the Murray meteorite, a Type II carbonaceous chondrite. Shown are chondrules (the largest is about 1 mm across) and irregular aggregates of olivine and pyroxene in a matrix composed largely of serpentine and carbonaceous matter. Courtesy of K. Keil.

that suggest they formed in asteroid interiors by fractionation and segregation of melts. Three minor groups of achondrites, the shergottites, nakhlites, and Chassigny meteorites, are mafic and ultramafic cumulates that have compositions similar to terrestrial basalts. They also have crystallization ages considerably less than other meteorites and appear to have been ejected from the surface of a major planet, perhaps Mars. The metal in meteorites is composed of two phases of Fe–Ni, which are intergrown in iron meteorites, producing the Widmanstätten structure visible on polished surfaces.

It is possible, using a variety of isotopic dating schemes, to learn a great deal about the history of meteorites. Sm–Nd and Rb–Sr isochron ages (see Chapt. 4) appear to record the times of accretion and partial melting in meteorite parent bodies. These ages cluster between 4.45 and 4.55 Ga with the upper extreme representing an estimate of the age of the solar system (fig. 2.2). Using *extinct radiogenic isotopes* with short half-lives, i.e. isotopes that completely decay in times less than 100 Ma, it is possible to estimate the *formation interval*, which is the time span between nucleosynthesis (element formation in stars) and planetary formation. ^{129}I , which decays to ^{129}Xe with a half-life of 17 Ma, is useful in estimating the formation interval. An excess of ^{129}Xe in the Earth and meteorites indicates a formation

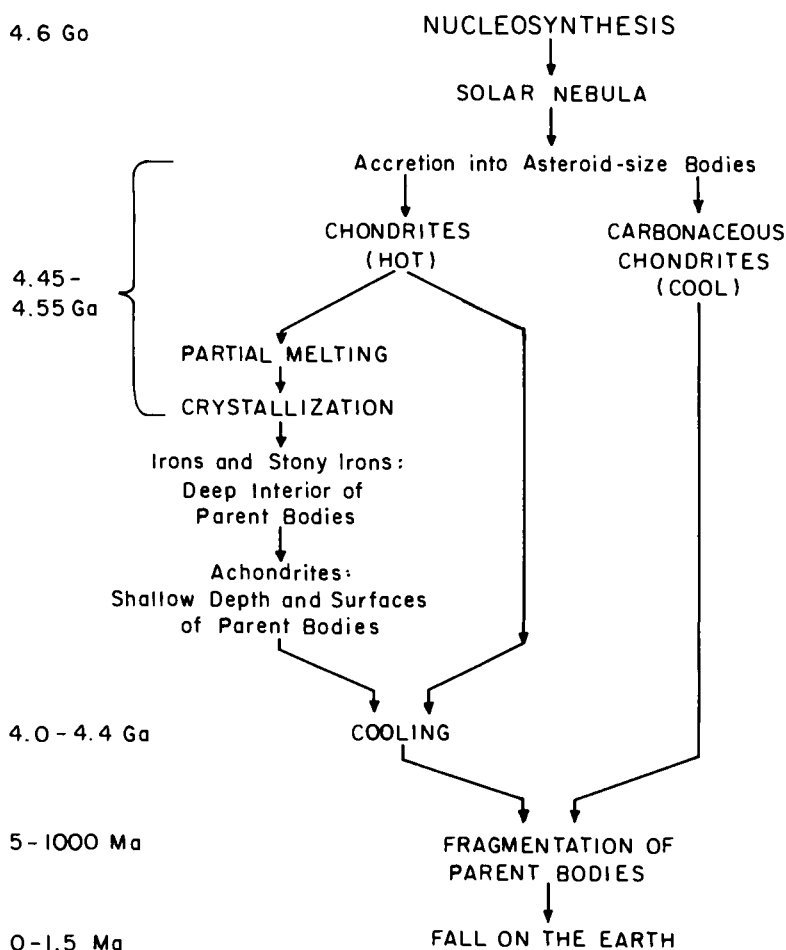


FIG 2.2 Summary of the history of meteorites. After Wasson (1985).

interval for the solar system less than 100 Ma. K-Ar dates from meteorites reflect cooling ages and generally fall in the range of 4.0–4.4 Ga (fig. 2.2). Fragmentation of asteroids by continual collisions exposes new surfaces to bombardment with *cosmic rays*, which are high-energy particles ubiquitous in interplanetary space. Interactions of cosmic rays with elements in the outer few millimeters of meteorites produce radioactive isotopes that can be used to date major times of parent-body breakup. Stone meteorites have cosmic-ray exposure ages of 5–100 Ma while irons are chiefly 200–1000 Ma. These differences reflect chiefly the fact that irons are more resistant to collisional destruction than stones. After a meteorite falls on Earth it is shielded from cosmic rays and the contents of cosmic-ray produced isotopes can be used to calculate *terrestrial ages*, the times of fall onto the Earth's surface. Although most terrestrial ages are less than 100 years, some as old as 1.5 Ma have been reported.

Most cosmochemical data suggest that iron meteorites come from the cores of asteroid-size bodies (radii of 300–1000 km). For such cores to form, the parent bodies must have melted soon after or during accretion and molten Fe-Ni settled to

the center of the bodies. It is possible to constrain the cooling rates of iron meteorites from the thickness and Ni content of kamacite (α -iron) bands. This, in turn, provides a means of estimating the size of the parent body, since smaller parent bodies cool faster than larger ones. Published cooling rates are generally in the range of 1–10°C/Ma, which indicates an upper limit for the radius of parent bodies of 300 km with most lying between 100 and 200 km. Results clearly eliminate the possibility that a single large planet was the parent body for meteorites and asteroids.

The origin of chondrites has been the subject of a great deal of research since these meteorites appear to have formed in the solar nebula and may be pristine remnants of this nebula. Although numerous studies have addressed the question of how chondrules form, none is completely definitive. The uniformity of chondrite formation ages (about 4.55 Ga) and the fact they are 10–30 Ma greater than planetary formation ages suggests that they formed before the planets and that they formed nearly simultaneously (within a few million years) throughout the inner solar system. Because various groups of chondrites (in terms of their compositions) seem to have formed at various distances from the Sun, it is possible from the study of these meteorites to learn more about conditions within the early solar nebula. Chondrules show features suggesting they formed by cooling of molten or partly molten droplets. They are composed chiefly of olivine and pyroxene, and exhibit a variety of textures with porphyritic textures being most common. Many models have been proposed for the origin of chondrules, including direct condensation from the solar nebula, condensation in the atmosphere of early planets, melting (and cooling) of dust particles caused by giant lightning bolts in the nebula, and melting of solids caused by impacts on the surfaces of asteroid-size bodies or impacts between these bodies. Unfortunately, there are problems with all of the models. For instance, direct condensation models cannot account for the high FeO/FeO+MgO ratios found in mafic silicates and the impact models seem to require unreasonably impact velocities (>3 km/sec). Clearly more research is needed to understand the origin of chondrules.

Chemical Composition of the Earth and Moon

Since it is not possible to sample the interior of the Earth and Moon, indirect methods must be used to estimate their composition. It is generally agreed that the Earth and other bodies in the solar system formed by condensation and accretion from a solar nebula, and that the composition of the Sun roughly reflects the composition of this nebula. Nucleosynthesis models for the origin of the elements also provide limiting conditions on the composition of the planets. It is possible to enhance our knowledge of the composition of the Earth's crust and upper mantle with geochemical data obtained from crustal igneous rocks and from ultramafic rocks, some of which may represent samples of upper mantle material (see Chapt. 3). Meteorite compositions and high-pressure experimental data, however, provide the most definitive information regarding the overall composition of the Earth.

Shock-wave experimental results indicate a mean atomic weight for the Earth of about 27 (mantle = 22.4 and core = 47.0), and show that it is composed chiefly of iron, silicon, magnesium, and oxygen. As evidenced by the following atomic weights of meteorites, however, no single meteorite class has the appropriate composition to

TABLE 2.1 *Estimates of the bulk composition of the Earth and Moon (in weight percent)*

	Earth			Moon
	1	2	3	4
Fe	34.6	29.3	29.9	8.3
O	29.5	30.7	30.9	44.7
Si	15.2	14.7	17.4	20.3
Mg	12.7	15.8	15.9	19.3
Ca	1.1	1.5	1.9	3.2
Al	1.1	1.3	1.4	3.2
Ni	2.4	1.7	1.7	0.6
Na	0.6	0.3	0.9	0.06
S	1.9	4.7	—	0.3

1. 32.4% iron meteorite (with 5.3% FeS) and 67.6% oxide portion of bronzite chondrites (Mason, 1966).
2. 40% Type I carbonaceous chondrite, 50% ordinary chondrite, and 10% iron meteorite (containing 15% sulfur) (Murthy and Hall, 1970).
3. Nonvolatile portion of Type I carbonaceous chondrites with FeO/FeO+MgO of 0.12 and sufficient SiO₂ reduced to Si to yield a metal/silicate ratio of 32/68 (Ringwood, 1966).
4. Based on Ca, Al, Ti = 5 × Type I carbonaceous chondrites: FeO = 12% to accommodate lunar density; and Si/Mg = chondritic ratio (after Taylor, 1979b, 1982).

match the shock-wave data for the entire Earth: carbonaceous chondrites (H₂O-free), 23.4 to 24.0; ordinary chondrites, 24.4; high-iron chondrites, 25.1; enstatite chondrites, 25.6; and iron meteorites, 55 (Anderson et al., 1971). However, meteorite classes can be mixed in such a way as to give the correct core/mantle mass ratio (32/68) and mean atomic weight, three examples of which are given in table 2.1. All three estimates indicate that iron and oxygen are the most abundant elements, followed by silicon and magnesium. Ninety percent or more of the Earth is composed of these four elements. Calcium, aluminium, nickel, sodium, and perhaps sulfur account for most of the remainder.

From lunar heat flow results, correlations among refractory elements, density and moment of inertia, it is possible to estimate bulk lunar composition (table 2.1). Compared to the Earth, the Moon is depleted in Fe, Ni, Na and S, and enriched in other major elements. The bulk composition of the Moon is commonly likened to the composition of the Earth's mantle because of similar densities. Data indicate, however, that the Moon is enriched in refractory elements (TiO₂ = 0.4%; Al₂O₃ = 8%; CaO = 6%) compared to the Earth's mantle (see table 2.3).

Trace-element concentrations in the Earth and Moon are difficult to estimate and depend on *ad hoc* assumptions. One approach involves estimates of upper and lower limits of potassium in the Earth (Hurley, 1968a, b). A lower limit is set by assuming that all of the Earth's ⁴⁰Ar is in the atmosphere and was produced from the radioactive decay of ⁴⁰K over geologic time, while an upper limit is set by establishing a lower limit for the fractional release of Ar from the Earth. By comparison with probable fractional release factors of H₂O, a lower limit for Ar release appears to be about 50%. The results suggest that the concentration of potassium in the Earth is about 200 ppm. By considering the lowest permissible K/Rb ratio for the Earth as deduced from meteorite data and the maximum permissible Rb/Sr ratio as determined from ⁸⁷Sr/⁸⁶Sr ratios of crustal and upper-

TABLE 2.2 *Element Classification according to Relative Volatilities from Silicate Melts under High-temperature Moderately Reducing Conditions*

Refractory group	Volatile group	
A. Oxyphile elements Mg, Al, So, P, Ca, Sc, Ti, Sr, Y, Zr, Nb, Ba, Rare earths, Hf, Ta, Th, U	H, C, N	Probable volatile species H ₂ O, CO, N ₂
	F, Cl, Br, I	Silicon and metal halides
	S, Se, Te, Na, K, Rb, Cs, Zn, Cd, Hg, Tl, Pb, As, Sb, Bi, Se, Te	Hydrides; elements
B. Siderophile elements Fe, Co, Ni, V, Cr, Mn, Ca, Ag, Au, Mo, W, Rh, Pd, Re, Ir, Pt	Ga, Ge, Sn, In	Oxides; sulfides

After Ringwood (1966).

mantle rocks, Rb and Sr contents are calculated. From element ratios in meteorites and igneous rocks, and from a knowledge of the general geochemical behavior of various trace elements, the concentrations of many trace elements in the Earth and Moon can be estimated.

It is necessary to discuss volatile and refractory elements and to establish a basis for defining element enrichments and depletions in planetary bodies. A rather arbitrary division exists between refractory and volatile elements. In general, *volatile elements* are those elements that can be volatilized from silicate melts under moderately reducing conditions at temperatures below 1400°C, while *refractory elements* are not volatilized under the same conditions. Refractory elements can be further subdivided into *oxyphile* and *siderophile elements*, depending on whether they follow oxygen or iron, respectively, under moderately reducing conditions. Selected elements together with their probable chemical form under these conditions are given in table 2.2. The Earth and Moon do not resemble either ordinary or Type I carbonaceous chondrites (C1) in terms of the trace elements presented in table 2.3. Compared to C1, the Earth is depleted in volatile elements and enriched in refractory elements. The Moon is depleted in volatile elements and siderophile refractory elements and enriched in oxyphile refractory elements compared to both C1 and the Earth (fig. 2.3). These differences present a major problem for models of lunar origin which derive the Moon from the Earth's mantle.

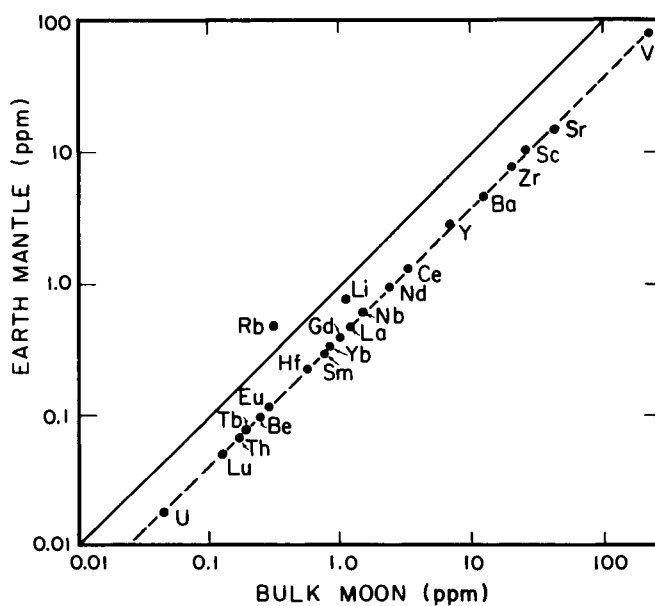
Vertical zoning of elements has occurred in the Earth and Moon. This zoning is the result of element *fractionation*, which is the segregation of elements with similar geochemical properties. Fractionation results from physical and chemical processes such as condensation, melting, and fractional crystallization. Elements such as K, Rb, Th, and U have been strongly enriched in the Earth's upper mantle and even more so in the crust in relation to the lower mantle and core. The crustal enrichment factors, expressed as percentages and calculated by dividing estimated element concentrations in the crust by estimated concentrations in the bulk Earth, are shown in figure 2.4. Those elements with enrichment factors less than 2% (Mn, Fe, Co, Ni, Si, etc.) are concentrated chiefly in the mantle or core. Fractionation of elements with large ionic size or charge into the crust probably reflects the difficulty with

TABLE 2.3 *Trace-element abundances in Meteorites, Earth, and Moon*

Element	Ordinary Chondrites	Type 1 Carbonaceous Chondrites* (CI)	Earth	Moon
Volatile				
Na (%)	0.68	0.79	0.5	0.06
K (ppm)	850	885	200	83
Rb (ppm)	3.5	3.5	1.0	0.3
Cs (ppb)	290	290	25	12
Oxyphile Refractory				
Ca (%)	1.2	1.4	2.0	3.2
Sr (ppm)	11	11	20	30
Ba (ppm)	4.1	3.6	5.5	9
Sc (ppm)	8.0	7.8	17	20
Y (ppm)	2.0	2.0	4	5
La (ppm)	0.24	0.37	1	1
Th (ppb)	43	50	65	125
U (ppb)	12	14	20	33
Siderophile Refractory				
Cr (%)	0.28	0.35	0.26	0.42
Fe (%)	25	27	3	8.3
Co (ppm)	600	765	1000	220
Ni (%)	1.2	1.5	2.0	0.04

*Volatile-free

After Shaw (1972), Taylor (1982).

**FIG 2.3** Comparison of oxyphile refractory element distributions in the Moon and Earth (mantle + crust). Note the uniform enrichment in the Moon. After Taylor (1982).

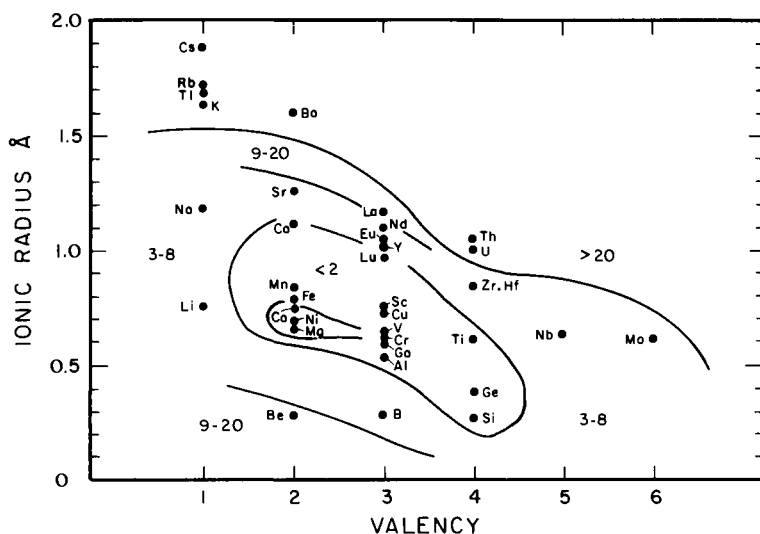


FIG 2.4 Estimated elemental crustal enrichment factors (in percent) as a function of ionic radius and valency. After Taylor (1964a).

which these elements enter minerals in the mantle and core, which are in a state of extreme close packing. These elements have been concentrated at shallow depths in the Earth by rising magmas or fluids.

Age of the Earth and Moon

The oldest dated rocks on the Earth, which have been recognized on most continents, are about 3.8–3.9 Ga in age. Detrital zircon ages suggest the existence of small amounts of continental crust by 4.2 Ga (see Chapt. 10). Gabbroic anorthosite samples from the Moon have isotopic dates of 4.4–4.5 Ga and model lead ages of lunar soils give similar results. Rb–Sr, U–Th–Pb, and Sm–Nd dates of meteorites suggest an age for the solar system of about 4.55 Ga. Model lead ages of volcanic rocks and ores on the Earth suggest an age for the Earth of 4.52–4.54 Ga. It would appear that the major melting events following accretion of the Earth, Moon and other planetary bodies in the solar system occurred at about 4.5 Ga.

Condensation and Accretion of the Planets

The origin of the Earth–Moon system is part of the general problem of the origin of the solar system. Although many models have been proposed for the origin of the solar system, only those that begin with a gaseous solar nebula appear to be consistent with data from both astrophysics and cosmochemistry. Stars form by contraction of gaseous nebulae, and planets probably form by condensation and accretion as nebulae cool (Hartmann, 1983). Just how the solar system formed can be considered in terms of three questions:

1. how the Sun acquired the gaseous material from which the planets formed;

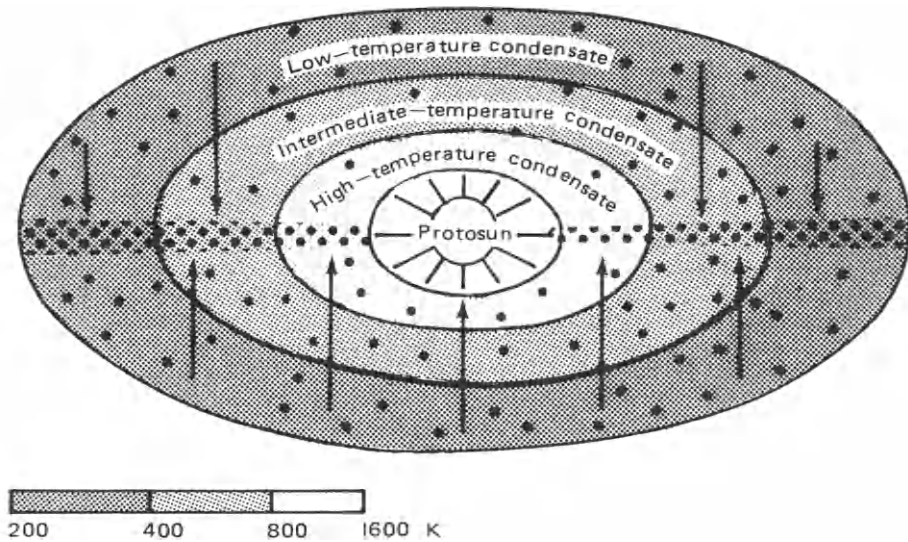


FIG 2.5 Schematic cross-section of the primitive solar nebula (after Ringwood, 1979). Temperatures in degrees K. Arrows indicate direction of gravitational sinking of solid particles.

2. the history of condensation of the gaseous material; and
3. the processes and history of planetary accretion.

Regarding the first question, one viewpoint is that the Sun, already in existence, attracted material into a gaseous nebula about itself. Another proposes that the pre-existing Sun captured a solar nebula of appropriate mass and angular momentum to form the solar system. Most theories, however, call upon condensation and accretion of the Sun and planets from the same cloud at approximately the same time. All models have in common a gaseous nebula from which the planets form. The minimum mass of such a nebula is about 1% of a solar mass. The mechanisms by which the nebula is distributed into a disc with the Sun at the center are not well understood. One mechanism is the transferring of angular momentum from the Sun to the nebula, caused either by hydromagnetic coupling during rotational instability of the Sun or by turbulent convection in the cloud. Condensing matter rapidly collapses into a disc about the Sun (Safronov, 1972) or into a series of Saturn-like rings which condense into the planets (fig. 2.5). Small planetesimals form within the cloud and spiral toward the ecliptic plane. Actual paths taken by planetesimals are complex. Regardless of the specific mechanism, the collapse of the nebula results in a cold cloud ($<0^{\circ}\text{C}$) except in the region near the protosun within the orbit of Mercury.

Two processes are important in planet formation. *Condensation* is the production of solid dust grains as the gases in the solar nebula cool. *Accretion* is the collision of the dust grains to form clumps and progressively larger bodies, some of which grow into planets. Condensation begins during collapse of the nebula into a disc, and leads to the production of silicate and oxide particles as well as to other compounds, all of which are composed chiefly of Mg, Al, Na, O, Al, Si, Fe, Ca, and Ni. In the cooler

parts of the nebula ices of C, N, O, Ne, S, Ar, and halogens form. The remaining gaseous mixture is composed chiefly of H and He. The ices are especially abundant in the outer part of the disc where the *Jovian planets* (those from Jupiter outwards) form; the silicate-oxide particles are concentrated in the inner part and give rise to the *terrestrial planets* (those from Mars inward). Thermodynamic considerations indicate that iron should be present in the cloud initially only in an oxidized state. It has been suggested the Type I carbonaceous chondrites (C1) may represent a sample of this primitive nebula. These meteorites contain only oxidized iron and large amounts of volatile components of the ices listed above (including organic compounds), and they have not been heated to more than 100°C to preserve volatile constituents.

Although it is commonly assumed that the solar nebula was well mixed, differences in isotopic abundances and age of various meteorites suggest that it was not (Lee, 1979; Wasson, 1985). Variations in the abundance of rare gases and isotopic variations of such elements as Mg, O, Si, Ca, and Ba indicate large mass-dependent fractionation effects. The isotopic anomalies appear to require a sudden injection of neutrons into the early solar nebula. The source of such a neutron burst may have been one or more supernova. It has also been suggested that such an injection event triggered the collapse of the solar nebula.

Planetary accretion models fall into two general categories. *Homogeneous accretion models* call upon formation of the planets from a generally cool (<500°C) largely condensed cloud, perhaps similar to C1 in composition. *Inhomogeneous accretion models*, on the other hand, involve simultaneous condensation and accretion of compounds from an initially hot cloud as it cools. Each of these models is considered in more detail in the following sections.

The specific mechanisms by which particulate matter evolve into small meter-sized bodies and these bodies, in turn, grow into planetesimals (small planets) and planets are not well understood. Most experimental data suggest that grains stick together during the early stages of accretion by weak electrostatic forces (Van der Waal forces) which are important in producing centimeter-sized objects. Theoretical studies indicate that a large fraction of the gaseous cloud will condense and accrete into several hundred asteroid-size bodies rather than into a swarm of small particles. The model of Goldreich and Ward (1973) shows that the first generation of planetesimals will increase in radius to the order of 100 m in times of about 1 year after condensation. Clusters of these planetesimals contract to form second-generation planetesimals with radii of the order of 5 km in a few thousand years, and these planetesimals accumulate into the planets by collisions. Wetherill (1977, 1986) simulated planetary growth rates by computer modeling. The results indicate that the accretion rate of the Earth and Venus increases rapidly with these planets, reaching a value within 50 km of the final radius in the first 25 Ma. Accretion is 98% complete for both planets in times of the order of 10^8 years and large impact velocities (≥ 10 km/sec) during the final stages of accretion are capable of producing extensive melting in both planets. The giant planets, Jupiter and Saturn, which consist chiefly of H and He accreted onto rocky cores, may also have grown by planetesimal collisions. Once a planet reaches 10–50 Earth masses it is able to attract H and He from the nebula and to grow into a giant planet. Solar nebula thermal models produce an amazingly good match between predicted and observed planetary

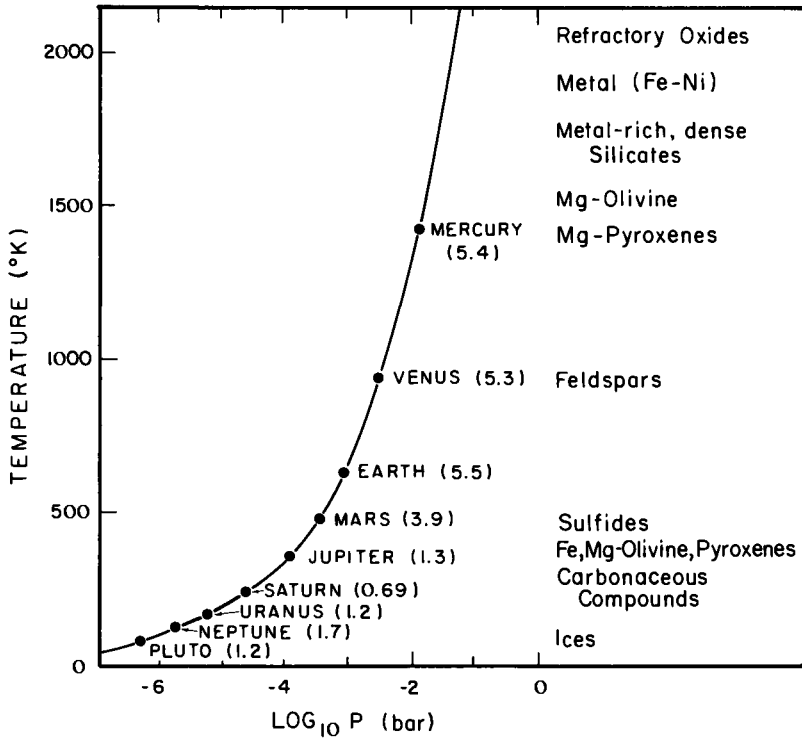


FIG 2.6 Calculated temperature and pressure distribution in the solar nebula late in planetary accretion history. Shown also is the distribution of the planets and their densities, and the predicted composition from condensation experimental data (table 2.4). After Hartmann (1983).

compositions (fig. 2.6). If, for instance, the gas cleared at the time when Mercury was accreting refractory elements, Venus and Earth are silica-rich with lower densities as observed. Mars has a still lower density, enriched in oxidized iron compounds, and the giant outer planets are composed chiefly of ices.

There are important geochemical constraints which must be accounted for by any acceptable planetary accretion model. One constraint, as described in the previous section, is the depletion in volatile elements in the Earth and Moon relative to C1 meteorites. Another is the enrichment of the refractory siderophile elements in the Earth compared to C1. Such enrichment strongly indicates that the Earth's mantle has not equilibrated with the core, because if it had the siderophile elements would be carried into the core and depleted in the mantle. Although the mantle appears to be well mixed in terms of major elements, it is not well mixed in terms of Sr, Pb, and Nd isotope distributions.

Inhomogeneous Accretion Models

Inhomogeneous accretion models call upon planetary growth by simultaneous condensation and accretion of various compounds as the temperature falls in an originally hot solar nebula. The zoned structure of the Earth, with an iron core

TABLE 2.4 *Condensation Sequence from Nebula of Solar Composition at 10^{-4} atm*

Phases	Temperature (°C)
Ca, Al, Ti, Zr, U, Th, REE oxides and some silicates	1250–1600
Metallic iron-nickel	1030–1200
Forsterite (Mg_2SiO_4) and enstatite (MgSiO_3)	1030–1170
Ca-plagioclase	900–1100
Na–K feldspars	~730
Troilite (FeS)	430
Fe, Mg pyroxenes and olivine	300
Magnetite	130
Carbonaceous compounds	100–200
Hydrated Mg silicates	0–100
Ice (H_2O)	<0

After Grossman (1972).

surrounded by a silicate mantle, may have been produced by sequential condensation and accretion of the Earth from a cooling solar nebula. Thermodynamic calculations of condensation (Grossman, 1972) have provided a quantitative basis for the model. Cameron (1973) has suggested a model whereby the planets develop in a rotating, disc-shaped nebula amounting to two solar masses. Contraction of such a nebula produces high temperatures (1000–1700°C) in the region where the terrestrial planets form. Rapid dissipation of the nebula results in planetary accretion on a very short time scale, on the order of 10^3 – 10^4 years. Cooling of the gas as dissipation proceeds results in condensation over a wide range of temperature. The sequence of compounds condensed from the nebula at a pressure of 10^{-4} atm is summarized in table 2.4. Beginning with a temperature of about 1700°C, Clark et al. (1972) propose that the Earth accreted simultaneously with condensation as temperature fell over a period of 10^4 years, resulting in a zoned Earth with an iron core surrounded by a mantle composed dominantly of Fe–Mg silicates. The upper mantle accreted over a longer time scale (10^5 – 10^7 years), involving the addition of compounds with oxidized iron, iron sulfides and hydrated silicates mixed with earlier, high-temperature condensates that failed to accrete (similar to the mineralogy of C1).

The inhomogeneous accretion models, however, are faced with the following problems which render them less appealing than homogeneous accretion models:

1. Thermodynamic data suggest a large degree of overlap in the condensation of iron and Mg-silicates (table 2.4), and hence the accretion of an early iron core is not clearly predicted.

2. The model predicts an Earth with a small core of Ca–Al–Ti oxides, which is not consistent with geophysical data. This problem may be avoided if U and Th, also concentrated in the small core, produce enough heat to melt the surrounding iron and allow the Ca–Al–Ti-rich material to migrate upwards into a zone at the base of the lower mantle. This model also predicts that the upper mantle and crust should be depleted in Ca, Al, and Ti, a feature which is not observed.

3. Iron does not enter condensing Mg silicates except at temperatures below 400°C, which is far less than the temperature at which the lower mantle accreted (1000–1200°C). Yet available geophysical and geochemical data indicate that the FeO/FeO+MgO ratio of the lower mantle must lie in the range of 0.1–0.2.

4. Sulfur should be a major component in the upper mantle of an inhomogeneously accreted Earth, yet data indicate that it is extremely rare.

5. Because of the large mass of the Earth, late volatile condensates should collide with the Earth at such high velocities that they are evaporated and lost to the atmosphere. Hence there is no way to account for the volatile constituents found in the upper mantle today.

6. It is necessary to have 10–20% of a low-atomic-weight element in the core to explain seismic wave velocity distribution (see Chapt.3) but the elements considered as possible diluents (Si, S, O) condense only after the iron core forms.

Homogeneous Accretion Models

The major difference between homogeneous and inhomogeneous accretion models is the timing of accretion. In the inhomogeneous models, condensation and accretion occur simultaneously, while in the homogeneous models condensation is essentially complete before accretion begins. In most homogeneous accretion models the planets and the Moon accrete from a cool ($<100^{\circ}\text{C}$), well-mixed, oxidizing solar nebula, similar perhaps in average composition to C1 meteorites. Time scales for accretion are 10^3 – 10^8 years. In these models the Earth and other planets become zoned by later melting of the mantle. Results suggest that more than enough heat was available to completely melt the Earth soon after or during the late stages of accretion (see Chapt. 3).

An example of a homogeneous accretion model for the Earth is the autoreduction model of Ringwood (1979). In this model, reduction of oxidized iron to metal, loss of volatiles, and fractionation resulting from core formation occur simultaneously during the late stages of accretion. It is assumed that most of the Earth was accreted in $\leq 10^6$ years and before the Sun evolved through a T-Tauri stage. A *T-Tauri stage* of stellar evolution is a late stage during which stars lose large amounts of mass. An important consequence of the T-Tauri stage is a strong solar wind that can clear the inner part of the solar system of many or most volatile elements and also blow away early planetary atmospheres. During planetary accretion in the Ringwood model, most of the gravitational energy is converted to heat and radiated away. The general features of the model are summarized in fig. 2.7 in terms of five evolutionary stages.

Stage I. Accretional energy is small and the temperature is less than 700°C maintained, in part, by evaporation of volatile components. A small nucleus (5–10% of the Earth mass) of cool, oxidized and volatile-rich material is formed during this stage. Silicates, hydrated silicates and Fe–Ni sulfides coexist in the nucleus.

Stage II. As accretion continues, the impact energy of accreting bodies becomes high enough to cause near-surface heating, leading to reduction of oxidized iron to metal by carbonaceous matter. Incoming bodies are also completely degassed on impact, resulting in the formation of a primitive reducing atmosphere.

Stage III. By the time the mass is equal to 0.2 of an Earth mass, the temperature has risen to about 1200°C . Reduction of iron oxides to metal occurs in the atmosphere, and as temperature increases to 1500°C , volatile elements (table 2.2) are baked out of the Earth and enter the atmosphere. The accreting material is a mixture of Fe–Ni metal and iron-free silicates depleted in volatile elements which remain in the atmosphere.

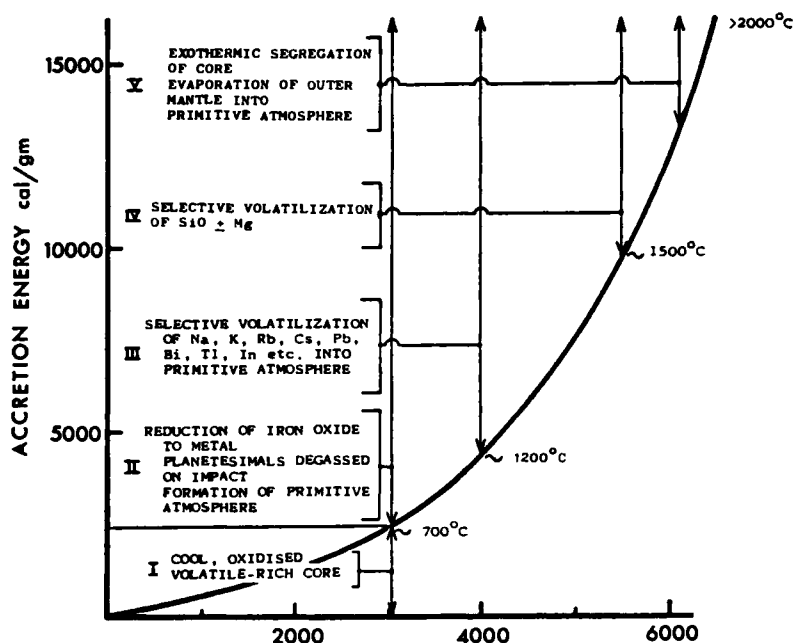


FIG 2.7 Relationship between accretion energy and radius for homogeneous accretion of the Earth. Stages of accretion are shown in relation to accretion energy and approximate surface temperatures. After Ringwood (1979). Used with permission of McGraw-Hill Book Co.

Stage IV. Increasing temperatures ($>1500^{\circ}\text{C}$) from increasing accretional energy reduce and volatilize silicate phases in accreting planetesimals. Silicon (as SiO) and to a lesser extent MgO are partially evaporated at such high temperatures and enter the primitive atmosphere. Less volatile constituents, such as Ca and Al silicates, continue to accrete with olivine and smaller amounts of pyroxenes.

Stage V. Melting begins at shallow depths in the Earth and iron, mixed with FeO or FeS , which reduce its melting point (Chapt. 3), segregates into bodies which move toward the center of the Earth, forming the core. This process is highly exothermic, as well as probably catastrophic, liberating 600 cal/gm over the entire Earth, resulting in the evaporation of the outer part of the Earth into the atmosphere. This material collects in a sediment ring about the Earth, and the Moon accretes from this ring (fig. 2.8). The early reducing atmosphere is either lost as it accumulates or it is blown away by a solar wind coming from the Sun as it passes through a T-Tauri stage of evolution.

Thermal Histories of The Earth and Moon

The evolution of planetary crusts and the role of plate tectonics in planetary evolution are critically dependent upon planetary thermal history. Today the Earth is cooling at a rate of about $200^{\circ}\text{C}/10^9 \text{ a}$, and about 80% of the surface heat flow is from the decay of radioactive isotopes in the Earth and 20% is from initial heat (Turcotte, 1980). The thermal evolution of planets is controlled by the amount of

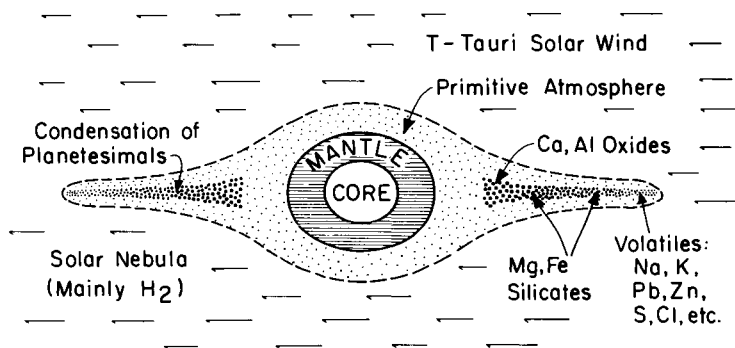


FIG 2.8 Structure of the Earth immediately after core formation, with its primitive atmosphere and sediment ring. After Ringwood (1972), with permission.

initial heat available during accretion and by the mechanism of planetary cooling. The principal sources of initial heat in the terrestrial planets are as follows:

1. *Accretional energy.* This energy is dependent upon such poorly known parameters as particle velocity distribution during accretion and the amount of input energy retained by the growing planet.
2. *Gravitational collapse.* As a planet grows, the interior is subjected to higher pressures, and minerals undergo phase changes to phases with more densely packed atoms (Chapt. 3) These changes are exothermic and large amounts of energy are liberated into planetary interiors.
3. *Radiogenic heat sources.* Radioactive isotopes liberate significant amounts of heat during decay. Short-lived radioactive isotopes, such as ^{26}Al and ^{244}Pu , may have contributed significant quantities of heat to planets during accretion. Long-lived isotopes, principally ^{40}K , ^{235}U , ^{238}U and ^{232}Th , are important heat producers throughout planetary history.
4. *Inductive heating by the early solar wind.* This source of initial heat appears to have been relatively small for the Earth and Moon.
5. *Core formation.* Core formation is a strongly exothermic process and appears to have occurred over a relatively short period of time (≤ 100 Ma), beginning during the late stages of planetary accretion.

Although most of these sources of early planetary heating are not known very precisely, order of magnitude estimates indicate that there was enough heat available during accretion of both the Earth and Moon if retained to completely melt both bodies. Because volatiles are retained in both bodies, however, it is unlikely that they were ever completely molten. Hence, it is necessary to remove the early heat very rapidly and convection seems the only process capable of bringing heat from planetary interiors to the surface in a short enough time interval (≤ 100 Ma) to escape complete melting.

It is possible to estimate the heat productivity in the Earth as a function of time from a knowledge of the concentration of U, Th, and K in the Earth. Although such concentrations are not well known, constraints can be placed on heat production from U, Th, and K contents of ultramafic rocks and meteorites. Results indicate that

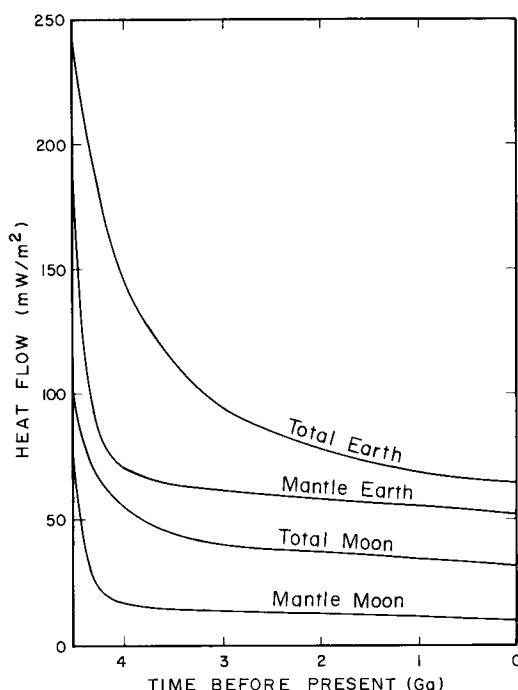


FIG 2.9 Variation of heat flow in the Earth and Moon over geologic time. Modified after Sharpe and Peltier (1978) and Schubert (1979). Mantle curves for both planetary bodies are heat flow into the base of the lithosphere (i.e. mantle heat flow).

heat production and surface heat flow were a factor of 3 to 5 greater in the early Archean than today (fig. 2.9). However, metamorphic mineral assemblages in Archean terranes indicate that Archean geothermal gradients were similar to modern geotherms in provinces with high heat flow such as the Rocky Mountains. Hence, the additional terrestrial heat in the Archean must have been lost from oceanic areas (Burke and Kidd, 1978). Whether or not oceanic Archean geothermal gradients were steeper than present gradients is a subject of considerable discussion, and evidence for steep Archean oceanic geotherms is ambiguous. For instance, the existence of abundant komatiites in the Archean may reflect steeper geotherms, or it may reflect abundant mantle plumes, in which komatiite magmas are produced by adiabatic melting as plumes move rapidly to the base of the lithosphere (Chapt. 10). It is not necessary to call upon steeper oceanic geotherms in the Archean, in that the additional heat may have been liberated by faster seafloor spreading and/or by a greater total length of ocean ridge systems.

The temperature dependence of mantle viscosity appears to be the most important factor controlling planetary thermal history (Schubert, 1979). Initially, when a planet is hot and viscosity is low, mantle convection rapidly cools the planet. This is apparent from calculated models for the thermal histories of the Earth and Moon, which show rapid decreases in mantle heat flow in the first 100 Ma after accretion (fig. 2.9). As mantle viscosity increases, beginning 100–300 Ma after accretion, convection cools both bodies at much reduced rates. As shown by the heat flow

curves, the Earth and Moon began with different initial heat contents caused by their different masses. After the early period of rapid cooling, differences in heat flow and average temperature of the two bodies became less pronounced as interior temperatures became less dependent on planetary size and reflected chiefly mantle viscosities, which are similar in the Earth and Moon. The fact that the total heat flow of both bodies drops less rapidly than the mantle heat flow in the first few hundred million years is due to the increasing contribution of radiogenic heat supplied by K, U, and Th isotopes. The time lag between the total heat flow and the mantle heat flow curves reflects the radiogenic heat contribution from the lithosphere where most of the K, U, and Th reside.

Origin of the Moon

Scientific results from the Apollo missions to the Moon have provided a voluminous amount of data on the structure, composition, and history of the Moon (Taylor, 1979b, 1982). Seismic data are generally interpreted in terms of five zones within the Moon, which are, from the surface inwards: a plagioclase-rich crust (dominantly gabbroic anorthosite), 60–100 km thick; an upper mantle, composed chiefly of pyroxenes and olivine, 300–400 km thick; a lower mantle of the same composition (with olivine dominating), about 500 km thick; 400 ± 100 km of partly melted mantle; and 200–400 km of iron core.

Among the more important constraints that any model for lunar origin must satisfy are the following:

1. The Moon does not revolve in the equatorial plane of the Earth, nor in the ecliptic. The lunar orbit is inclined at 5.1° to the ecliptic, whereas the Earth's equatorial plane is inclined at 23.4° .
2. Tidal dissipation calculations indicate that the Moon is retreating from the Earth, resulting in an increase in the length of the day of 15 sec/a. Orbital calculations suggest that the Moon had not been closer to the Earth than about 240,000 km (Hansen, 1982).
3. The Moon is enriched in refractory oxyphile elements and depleted in refractory siderophile and volatile elements relative to the Earth. Particularly important is the low density of the Moon (3.34 gm/cm^3) compared to other terrestrial planets, which indicates the Moon is significantly lower in iron than these planets.
4. The Earth-Moon system has an anomalously large amount of angular momentum ($3.45 \times 10^{41} \text{ gm/cm}^2/\text{sec}$) compared to the other planets.
5. The oxygen isotope composition of lunar igneous rocks collected during the Apollo missions is the same as that of mantle-derived rocks from the Earth. Because oxygen isotope composition seems to vary with position in the solar system, the similarity of oxygen isotopes in lunar and terrestrial igneous rocks suggests that both bodies formed in the same part of the solar system at approximately the same distance from the Sun.
6. Radiometric dates from igneous rocks on the lunar surface range from about 3.1 to 4.46 Ga. Model ages indicate that most lunar rocks formed during a major magmatic event at about 4.46 Ga.

TABLE 2.5 *Summary of Major Constraints for Models of Lunar Origin*

	Fission	Capture	Double-planet	Giant Impactor
1. Angular momentum	I	C (I)	I	C
2. Orbit	I	C (C)	I	C
3. Oxygen isotopes	C	C (C)	C	C
4. Composition	I	I (I)	C	C

Values in parentheses for disintegrative capture.

C, consistent; I, inconsistent or requires *ad hoc* assumptions

Models for the origin of the Moon generally fall into one of four categories: (1) fission from the Earth; (2) precipitation from a sediment ring around the Earth; (3) capture by the Earth; and (4) impact on the Earth's surface by a Mars-size body. Any acceptable model must account for the constraints listed above, and thus far none of these models is completely acceptable. Each of the hypotheses will be briefly discussed, and a summary of just how well each model complies with major constraints is given in table 2.5.

Fission Model

Fission models involve separation of the Moon from the Earth during an early stage of rapid spinning when tidal forces overcome gravitational forces. One version of the fission hypothesis (Wise, 1963) suggests that formation of the Earth's core reduced the amount of inertia, increasing the rotational rate spinning off material to form the Moon. Such a model is attractive in that it accounts for the similarity in density between the Earth's mantle and the Moon, and for the absence of a large metallic lunar core. The hypothesis is also consistent with the fact that the Moon revolves in the same direction as the Earth rotates, the circular shape of the lunar orbit, the existence of a lunar bulge facing the Earth, and similar oxygen isotope ratios of the two bodies. Also, the model explains the iron-poor character of the Moon if fission occurred after core formation in the Earth, since the Moon would be formed largely from the Earth's mantle.

Fission models, however, face several major obstacles. They do not explain the inclination of the lunar orbit, and they require more than four times the total angular momentum than is available in the present Earth–Moon system. If the Earth–Moon system ever had this excess angular momentum, no acceptable mechanism has been proposed to lose it (Wood, 1986). Also, lunar igneous rocks are more depleted in siderophile and volatile elements than terrestrial igneous rocks, indicating that the lunar interior is not similar in composition to the Earth's mantle (Taylor, 1987).

Although some investigators still favor a fission model (Binden, 1986), the above problems seem to render the model implausible.

Double Planet Models

The double planet models involve an accreting Earth with simultaneous accretion of the Moon from orbiting solid particles (Harris and Kaula, 1975). A major advantage of these models, also known as coaccretion or precipitation models, is that they do

not invoke special, low-probability events. The models assume that as the Earth accreted, solid particles accumulated in orbit about the Earth and accreted to form the Moon. One version of the model is part of the homogeneous accretion model of Ringwood (1972a, 1975). During stage V of the model (fig. 2.7), silicates precipitate as the hot atmosphere expands and cools, forming a sediment ring of planetesimals around the Earth (fig. 2.8). This sediment ring becomes unstable and coagulates to form the Moon. Because the volatile elements are largely lost with the removal of the early atmosphere by intense solar radiation from the T-Tauri wind, the material from which the Moon accretes is enriched in oxyphile refractory elements and depleted in volatile elements, thus accounting for the major geochemical differences between the Earth and Moon. Equilibrium between silicates and metal in the lunar interior accounts both for the low oxygen fugacity in the Moon and for depletion of siderophile elements in lunar igneous rocks.

The most serious problems with the double planet models are that they do not seem capable of explaining the large amount of angular momentum in the Earth–Moon system (Wood, 1986) and they do not readily explain the inclined lunar orbit.

Capture Models

Capture models propose that the Moon and Earth formed in different parts of the solar nebula, and that early in the history of the solar system the Moon or its predecessor approached the Earth and was captured. Both catastrophic and non-catastrophic models of lunar capture have been described, involving retrograde and prograde orbits for the Moon prior to capture.

Capture models fall into two categories; *intact capture*, where a fully accreted Moon is captured by the Earth, and *disintegrative capture*, where a planetesimal comes within the Earth's Roche limit, is fragmented by tidal forces with most of the debris captured in orbit about the Earth, and the debris reaccretes to form the Moon (Wood and Mitler, 1974). Although intact capture models may explain the high angular momentum and inclined lunar orbit, they cannot readily account for geochemical differences between the two bodies. The similar oxygen isotopic ratios between lunar and terrestrial igneous rocks suggests that both bodies formed in the same part of the solar system, yet the capture model does not offer a ready explanation for the depletion of siderophile and volatile elements in the Moon. Also, intact lunar capture is improbable because it requires a very specific approach velocity and trajectory. Disintegrative capture models cannot account for the high angular momentum in the Earth–Moon system. Also, calculations seem to indicate that a planetesimal coming within the Earth's Roche limit will not fragment in the short amount of time it remains there (Mizuno and Boss, 1985).

Giant Impactor Model

The giant impactor model involves a glancing collision of the Earth with a Mars-sized body during which debris from both the Earth and the colliding planet collect and accrete in orbit about the Earth to form the Moon (Hartmann, 1986; Cameron, 1986). Such a model has the potential of eliminating the angular momentum and nonequatorial lunar orbit problems, as well as providing a means to explain chemical

differences between the Earth and Moon. One of the major factors that led to the giant impactor model is that computer calculations indicate that numerous large bodies (up to Mars-sized) existed in the inner solar system during the early stages of planetary accretion (Wetherill, 1985). Results suggest the existence of at least 10 bodies larger in size than Mercury, and several equal or larger than Mars. Wetherill (1986) estimates that about one-third of these objects collided with and accreted to the Earth, providing 50–75% of the Earth's mass. The obliquities of planets and the slow retrograde motion of Venus are also most reasonably explained by late-stage impact of large planetesimals. Because the material that makes up the Moon comes chiefly from the mantle of the impacting body, in which iron had already segregated into a core, a low iron (and other siderophile element) content of the Moon is predicted by the model. At least part of the collisional debris is vaporized during the collision, and that remaining from which the Moon accretes is more depleted in volatile elements than the Earth. Such vaporization should also leave the orbiting debris relatively enriched in oxyphile refractory elements, as observed in the Moon.

Although the giant impactor model needs to be more fully evaluated, it appears to be capable of accommodating more of the constraints related to lunar origin than competing models.

The Earth's Rotational History

Integration of equations of motion of the Moon indicate there has been a minimum in the Earth–Moon distance in the geologic past; that the inclination of the lunar orbit has decreased with time; and that the eccentricity of the lunar orbit has increased with time as the Earth–Moon distance has increased (Lambeck, 1980). It has long been known that angular momentum is being transferred from the Earth's spin to lunar orbital motion, which results in the Moon moving away from the Earth. The current rate of this retreat is about 5 cm/a. The corresponding rotational rate of the Earth has been decreasing at a rate of about 5×10^{-22} rad/sec. Tidal torques cause the inclination of the Moon's orbit to vary slowly with time as a function of the Earth–Moon distance. When this distance is greater than 10 Earth radii, the lunar orbital plane moves towards the ecliptic, and when it is less than 10 Earth radii it moves towards the Earth's equatorial plane. For an initially eccentric orbit, the transfer of angular momentum is greater at perigee than apogee, and hence the degree of eccentricity increases with time. These evolutionary changes in angular momentum of the Earth–Moon system result in changes in the length of the terrestrial day and month. Estimates of the number of days per year N_1 , the number of days per synodic month N_2 , and the number of synodic months per year N_3 , as a function of time, are given in figure 2.10.

Many groups of organisms secrete sequential layers that are known to be related to cyclical astronomical phenomena. Such organisms, which are known as *paleontological clocks* (Pannella, 1972), provide a means of independently evaluating the curves given in figure 2.10. The most important groups are corals and bivalves. Daily increments in these organisms are controlled by successive alternation of daylight and darkness. Seasonal increments reflect changes in the length of sunlight per day, seasonal changes in food supply, and, in some instances, tidal changes. Actual growth patterns are complex, and local environmental factors may cause difficulties

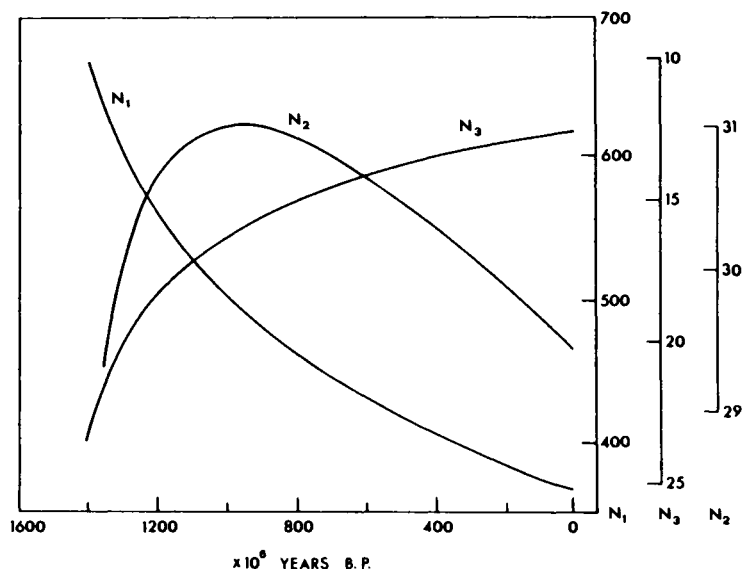


FIG 2.10 Number of solar days per year (N_1) and per synodic month (N_2), and number of synodic months per year (N_3) as a function of geologic time (after Lambeck, 1980). Model assumptions: phase-lag of semi-diurnal tides = 6 degrees; length of the year and mass of the Earth assumed constant; and tidal evolution of Earth's orbit is assumed small.

in identifying periodic growth patterns. Results from Phanerozoic bivalves and corals, however, are consistent and suggest a decreasing rotational rate of the Earth of about 5.3×10^{-22} rad/sec, which is in good agreement with astronomical values. This corresponds to 420–430 days/a at 500–600 Ma ago. Stromatolites (Chapt.11) also deposit regular bands, although a quantitative relationship of banding to astronomical rhythm has not yet been established. The interpretation of stromatolite bands from the Biwabik Iron Formation in Minnesota is that at 2000 Ma ago there were 800–900 days/a (Mohr, 1975). This is in good agreement with the extrapolation of N_1 in figure 2.10.

If the Moon came within the Roche limit of Earth (~ 2.9 Earth radii) after 4.5 Ga, a record of such a close encounter should be preserved. Even if the Moon survived this encounter without disintegrating, the energy dissipated in the two bodies would largely melt both bodies and completely disrupt and recycle any crust. The preservation of crust 4.46 Ga in age on the Moon strongly indicates that if such a close encounter did occur, it was prior to this time.

Summary Statements

1. Most meteorites appear to be derived from the asteroid belt and to represent fragments of asteroids formed at 4.45–4.55 Ga. Chondrites may have condensed directly from the solar nebula, whereas other groups of meteorites come from asteroids that have undergone partial melting and fractionation.
2. Radiometric dates from igneous rocks on the lunar surface range from about 3.1 to 4.46 Ga. Model ages indicate that most lunar rocks formed during a major magmatic event at about 4.46 Ga.

3. Isotopic dating indicates that planetary accretion was largely or entirely completed by 4.5 Ga, the first major period of melting in the terrestrial planets.
4. The Earth, Moon and other terrestrial planets condensed from a gaseous solar nebula and accreted from relatively high temperature condensates in the inner part of the nebula. Homogeneous accretion models are faced with fewer problems than inhomogeneous models.
5. Isotopic anomalies in meteorites suggest that the early solar nebula was injected with neutrons, perhaps from a nearby supernova.
6. Models for lunar origin must account for: (1) the large amount of angular momentum in the Earth–Moon system; (2) the fact that the lunar orbit is not coincident with the Earth's equatorial plane; (3) the similarity in oxygen isotopic composition of the Earth and Moon; and (4) the depletion in volatile and siderophile refractory elements in the Moon relative to the Earth.
7. Of the models for lunar origin, only the giant impactor model appears capable of accommodating most major scientific constraints.
8. Astronomical and paleontological data indicate that the Moon has been retreating from the Earth at a rate of about 5×10^{-22} rad/sec for at least 1000 Ma. If the Moon came within the Earth's Roche limit, it must have occurred within the first 50 Ma after planetary accretion.

Suggestions for Further Reading

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CHAPTER 3

The Mantle and Core

Detailed Structure of the Mantle

The Moho

A great deal has been learned about the Moho since it was first reported by Andrija Mohorovicic in 1910. It is generally well defined by an abrupt increase in P-wave velocity from about 6.6 km/sec to about 8.0 km/sec. It occurs from 10 to 12 km beneath oceans to 30 to 50 km beneath continents (ranging up to 80 km). Compressional waves that travel along the top of the mantle at the Moho are designated P_n waves. P_n velocities commonly range from 7.9 to 8.1 km/sec, although some areas exhibit velocities as low as 6.5 km/sec and other areas as high as 9.0 km/sec. Seismic anisotropy of P_n velocities has been recognized in oceanic areas. In general, the most rapid P velocities occur at approximately right angles to ocean ridges. The sharpness of the Moho is not well defined, but ranges from as little as 0.1 km in some oceanic areas and 0.5 km in Precambrian shield areas to over 1 km in some tectonically active regions. Estimates of the Moho temperature beneath the continents range from 500°C to 700°C and beneath oceans from 150°C to 200°C.

In some areas, one or both of the following problems have been encountered in defining the Moho (Knopoff, 1969): (a) two major discontinuities are detected at a depth less than 60 km, and (b) anomalously low P_n velocities. In parts of the Basin and Range Province, discontinuities occur at 25 km ($V_P = 7.7$ m/sec) and about 50 km ($V_P = 8.1$ km/sec), and in western Italy discontinuities are reported at 10 km ($V_P = 7.4$ km/sec) and 55 km ($V_P = 8.1$ km/sec). The question of which interface defines the crust has significance in terms of crustal evolution and the origin of both discontinuities. In regions where two discontinuities occur, the shallower one is usually accepted as the base of the crust. Low P_n velocities ranging from 6.5 to 7.8 km/sec characterize the upper mantle beneath ocean ridges and beneath continental rifts. Low P_n velocities may also occur beneath island arcs, marginal-sea basins and young orogenic areas. These low velocities correlate well with high surface heat flow, thin lithosphere and high electrical conductivity and seismic wave attenuation (low Q) immediately beneath the Moho. Unusually high P_n velocities of 8.6 km/sec or greater have been reported from some ocean basins and Precambrian shields. A correlation of P_n velocity distribution with crustal type suggests that the crust and upper mantle are not decoupled, and that they evolved together as part of the same system.

Lateral variations in P_n velocity in the central and western United States are shown

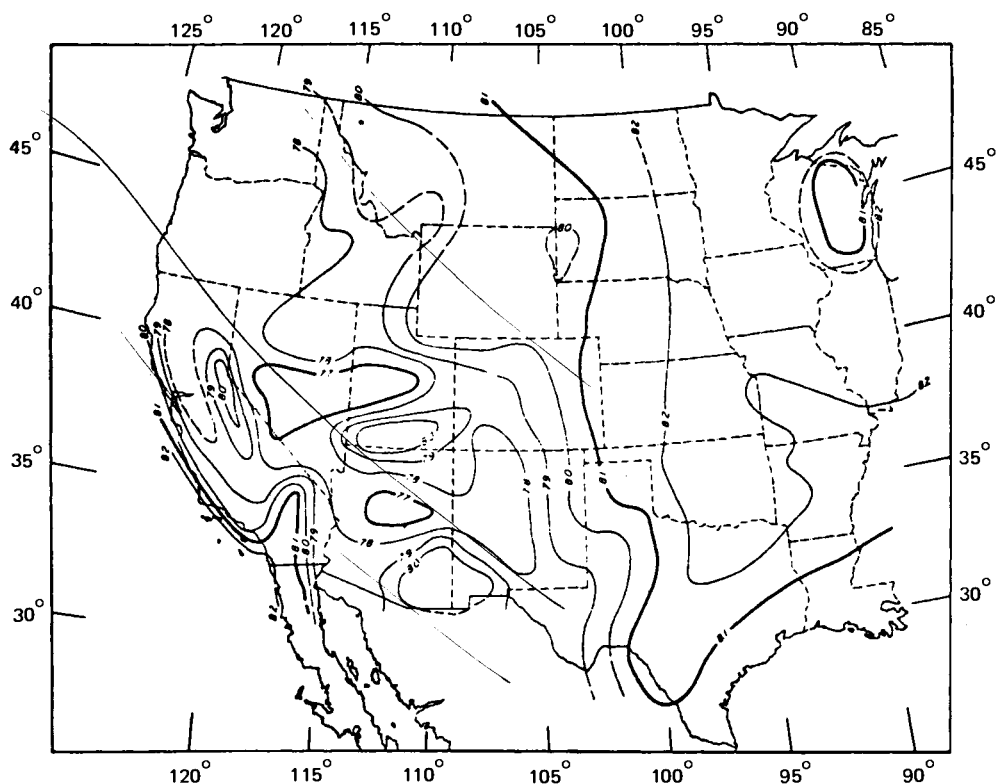


FIG 3.1 P_n velocity distribution in the western and central United States. From Archambeau et al. (1969), copyright by the American Geophysical Union.

in figure 3.1. Variations in P_n velocity in this region correlate with surface topography, average crustal velocities, and in some cases geologic provinces. In general, mountainous areas have lower P_n velocities than regions with little surface topography. Also, the Cordilleran orogenic belt in the western United States exhibits variable and, on the whole, lower P_n velocities than does the stable interior. P_n velocities of 7.7 to 7.8 km/sec characterize the Basin and Range Province, as does thin crust. The Great Plains, Colorado Plateau and stable continental areas of the eastern United States generally exhibit P_n velocities of 8.0 to 8.2 km/sec. Velocity contours outline the Sierra Nevada Range in California, where P_n ranges from 7.8 to 8.0 km/sec. The E-W trending contours in Utah and Nevada, however, do not seem to reflect young surficial tectonic features, the trends of which are generally N-S.

The Upper Mantle and Transition Zone

From studies of spectral amplitudes and travel times of body waves it is possible to refine details of the structure and of the upper mantle and transition zone (Hales et al., 1980; Walck, 1985). The P- and S-wave velocity structure beneath several crustal types is shown in figures 3.2 and 3.3. Some segments of the crust are characterized by a high-velocity lid ($V_p \geq 7.9$ km/sec) underlain by a low-velocity

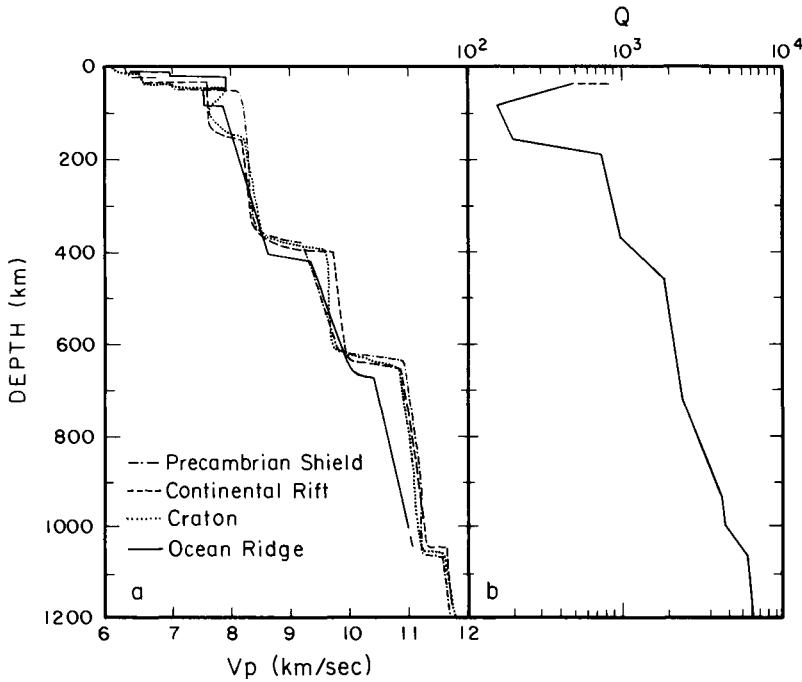


FIG 3.2 (a) P-wave velocity distribution in the mantle. After Archambeau et al. (1969); Hales et al. (1980); Walck (1985). (b) Specific attenuation factor (Q) distribution in the mantle (excluding regions beneath Precambrian shields. After Archambeau et al. (1968).

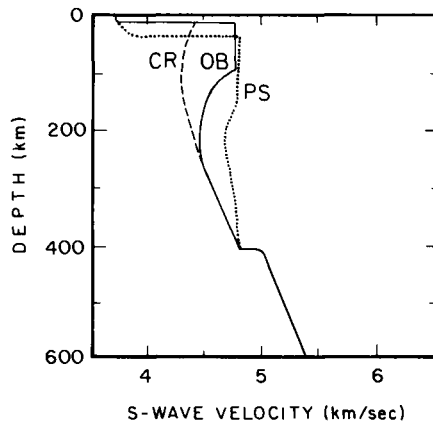


FIG 3.3 S-wave velocity distribution beneath three crustal types: CR, continental rift; OB, ocean basin; and PS, Precambrian shield. After Grand and Helmberger (1984).

zone (the LVZ) beginning at a depth as shallow as 60–70 km and extending to depths of 100–300 km. S-wave low-velocity zones may extend to 400 km (fig. 3.3) The LVZ is generally assumed to mark the base of the lithosphere and averages about 100 km in depth beneath oceans. Beneath ocean ridges and most continental rifts the LVZ

may extend to the Moho at depths as shallow as 25 km. Low velocity zones are very minor or not detected beneath Precambrian shields. Hence, the thickness of the lithosphere beneath shields is not well known. Most data indicate thicknesses greater than 200 km and perhaps as great as 400 km. It is of interest to note that in those regions where a LVZ is poorly developed or does not exist, free-air gravity anomalies are negative; where a well-developed LVZ exists, gravity anomalies are weakly to strongly positive. Figure 3.2b shows an idealized model for the distribution of Q in the Earth for both body and surface waves. A prominent minimum in Q at 150–200 km characterizes the LVZ.

Below 250 km for P-waves and below 400 km for S-waves, seismic velocity distributions show less variability and are broadly similar beneath both oceanic and continental areas. From the base of the LVZ to 400 km, P-wave velocities increase only slightly. At about 400 km a major discontinuity occurs, which defines the base of the upper mantle. A second velocity discontinuity is observed at about 670 km and a third one at about 1050 km, marking the base of the transition zone. Velocity profiles also suggest a minor discontinuity between 500 and 550 km in some regions. Although P-wave velocity distributions at depths from 400 to 1000 km are very similar, S-wave distributions show lateral variability in this depth range, indicating lateral heterogeneity. S-wave travel times in the upper mantle beneath continents that have undergone Phanerozoic tectonic activity are generally greater than times in the upper mantle beneath continents that were stabilized during the Precambrian. Also, travel times beneath oceanic regions decrease with crustal age away from ocean spreading centers. Changes in the gradient of Q for body waves characterize both the 400 km and 1050 km discontinuities (fig. 3.2b).

Another way to study lateral variations in the upper mantle is by the use of travel-time anomaly distributions (Hales, 1972; Dziewonski and Anderson, 1983). Figure 3.4 is a P-wave travel-time anomaly map of the United States. In general, P-wave arrivals are late in areas of young orogenic activity (0.2 to 0.8 sec) and early in shield and craton areas (0 to minus 0.5 sec). Both P_n velocity distributions and travel-time anomaly distributions are consistent with fast upper-mantle P-wave velocities under shields and platforms and slow ones beneath young orogenic areas.

Slab Penetration Depth

The depth to which descending slabs of lithosphere sink into the mantle is important in terms of the style of mantle convection. The cessation of earthquakes at about 700 km in descending slabs has traditionally been interpreted as evidence that slabs do not penetrate the lower mantle, and that the mantle above the 670-km discontinuity convects separately from the lower mantle. However, seismic travel-time studies clearly indicate the presence of seismic anomalies in subducting slabs that extend to depths of several hundred kilometers below the 670-km discontinuity (Creager and Jordan, 1984, 1986). These anomalies indicate that lithospheric slabs sink into the lower mantle and these results do not favor layered convection models. The cessation of earthquakes in slabs at 700 km probably reflects increased slab ductility at and below this depth. Results indicate that the flux of lithosphere descending across the 670-km discontinuity is large, at least $300 \text{ km}^3/\text{a}$. At this rate the volume of the upper mantle is cycled into the lower mantle once every billion years.

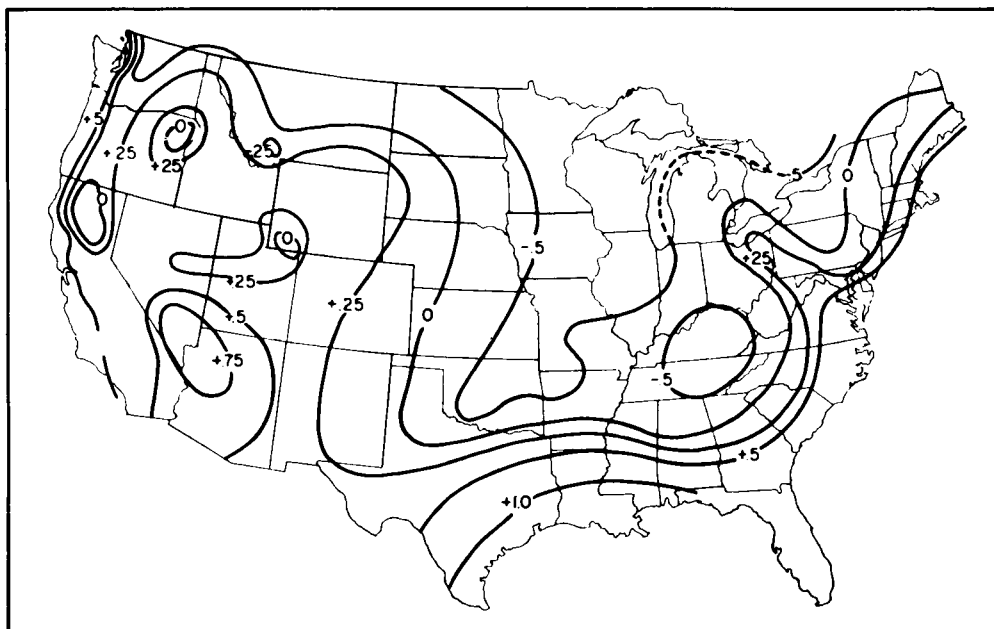


FIG 3.4 Travel-time anomaly map of the United States (contours in seconds). From Herrin and Taggart (1968).

The Lower Mantle

Although seismic velocities, in general, are rather uniform in the lower mantle, several increased velocity gradients are recognized (fig. 3.5). Relatively high gradients occur at 1160–1220, 2180–2370, and perhaps at 2700–2750 km, while low gradients are recognized at 1260–1330, 1750–1850 and 2460–2600 km. Although it is small in magnitude compared with that in the upper mantle and transition zone, some lateral variation in both P- and S-wave velocities in the lower mantle is indicated by existing data. Localized heterogeneities of the order of 1000 km in size and involving several percent velocity change have been detected in parts of the lower mantle.

The structure of the lower 200 km of the mantle, known as the *D''* region, is poorly known. Interpretations of the S- and P-wave velocity structure differ among investigators, due perhaps to strong lateral heterogeneity of the region (Young and Lay, 1987). Most seismic analyses indicate a general decrease of velocity gradients in *D''*, but it is not clear if absolute velocities decrease with depth. As well as being laterally heterogeneous, *D''* is also probably stratified with a possible discontinuity occurring at about 280 km above the core–mantle interface. Seismic data seem to rule out the presence of high relief on the core–mantle interface, but will permit relief with amplitudes not larger than 5 km (Cox and Cain, 1972). This relief likely results from convection in the overlying mantle and provides coupling between the mantle and core. Data suggest that the *D''* region differs from overlying mantle by either a phase change or a compositional change.

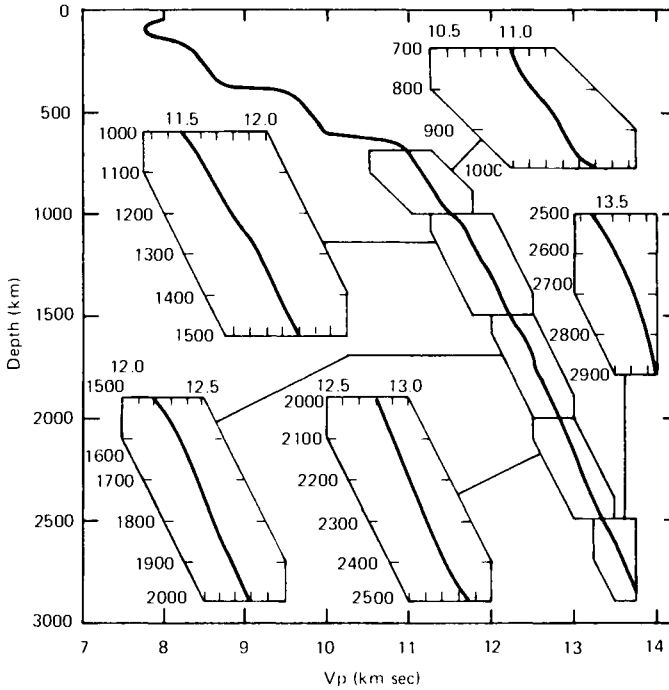


FIG 3.5 P-wave velocity distribution in the lower mantle. After Johnson (1969), copyright by the American Geophysical Union.

Seismic Tomography

Seismic tomography, like its medical analogue, combines information from a large number of criss-crossing waves to construct three-dimensional images of the interior of the Earth. Relatively hot and cool regions of the Earth's interior can be mapped since seismic velocities vary inversely with temperature. Orientation of minerals in convecting mantle can also increase seismic velocities parallel to 'fast' crystallographic axes. The accumulation of digital data from global seismic networks now makes it possible to begin to construct reliable three-dimensional models of mantle structure. Such information provides important constraints on the style of convection in the Earth.

Combining data from surface and body waves it is possible to compare the seismic structure of the upper and lower mantle. Some of the exciting results from tomographic studies are illustrated in two vertical cross-sections in figure 3.6 (after Woodhouse and Dziewonski, 1984). Results for the upper mantle (25–670 km) are based on surface-wave data and those for the lower mantle (670–2900 km) on travel-time residuals of P-waves, which accounts for the mismatch at the 670-km discontinuity. Noteworthy in section a are the deep high-velocity roots beneath the Brazilian and African Shields which appear to merge below 400 km. This high-velocity region continues into the lower mantle, perhaps at a steeper angle. Not shown on these sections are similar deep roots for the Canadian and Baltic Shields.

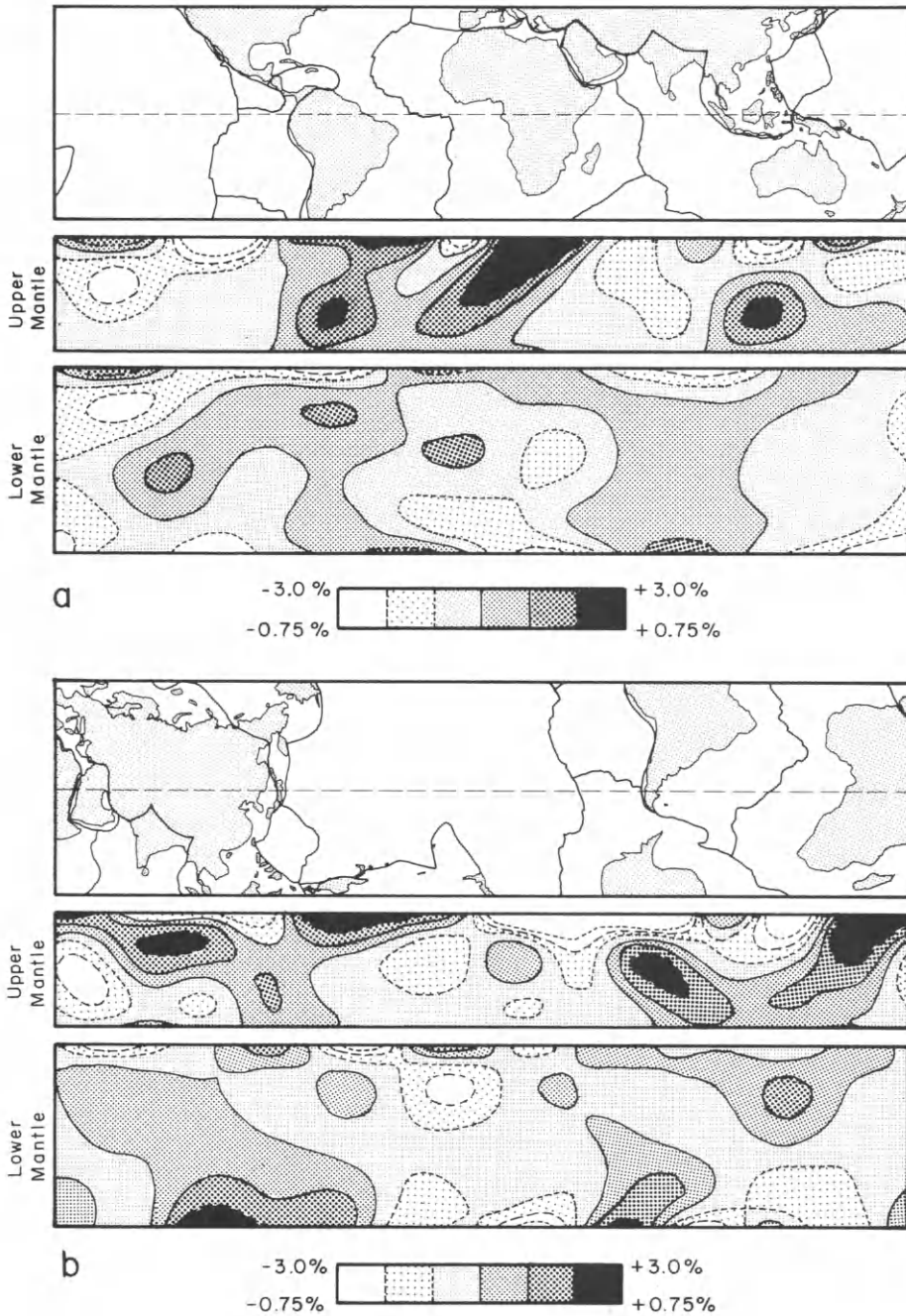


FIG 3.6 Two vertical seismic-wave velocity sections through the mantle. Sections are great circles shown on the maps. The upper mantle (25–670 km) has an 8:1 vertical exaggeration and the lower mantle (670–2900 km) a 4:1 vertical exaggeration. Upper-mantle distributions based on S-waves and lower mantle on P-waves. Deviations in percent from average velocities, + refers to fast and – to slow. After Woodhouse and Dziewonski (1984), copyright by the American Geophysical Union.

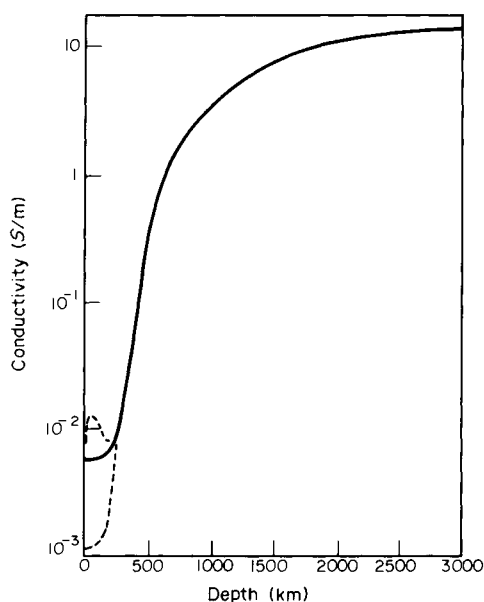


FIG 3.7 Idealized electrical conductivity distribution in the mantle.
From Rikitake (1973).

In both sections, low-velocity zones underlie ocean ridges, although their depth of penetration into the mantle varies. For instance, the root zone of the Mid-Atlantic ridge seems to end at the 400 km discontinuity, while that of the East Pacific rise may extend to the base of the mantle (offset by the Precambrian shield high-velocity root). The dipping high-velocity roots beneath Japan, the southern tip of South America and Indonesia appear to penetrate the 670 km discontinuity and probably reflect descending slabs of lithosphere. As with the slab penetration studies, the seismic tomography results indicate that large inhomogeneities exist in the mantle and that they extend to great depths. The origin and significance of these anomalies are not yet clear. The fact that many anomalies cross the 670-km discontinuity, however, tends to favor whole-mantle rather than layered-mantle convection. Perhaps the most intriguing question is that of whether continental plates carry their deep mantle roots with them as they move about on the surface.

Electrical Conductivity Distribution

Considerable uncertainty should be attached to estimates of electrical conductivity distribution in the mantle due to poor resolution of data. Especially large variations occur in conductivity models for depths less than 500 km. An idealized electrical conductivity distribution in the mantle is shown by the solid line in figure 3.7. The dashed lines indicate the combined range of uncertainty and real variations at depths less than 250 km. In most continental areas the conductivity of the lower crust is in the range of 10^{-2} to 10^{-3} S/m. Conductivity appears to increase rapidly to the vicinity of 1 S/m at the base of the transition zone and less rapidly to about 10 S/m at the core-mantle interface. The deep oceanic mantle may be more conductive than the mantle beneath continents at similar depths (Roberts, 1986). Such a pattern is

consistent with seismic tomographic data and supports the existence of large-scale lateral inhomogeneities in the mantle. In regions where the seismic LVZ occurs at shallow depths, such as beneath the Basin and Range Province, ocean ridges and young continent rift systems, a region of shallow high electrical conductivity ($\leq 10^{-2}$ S/m) also exists. This high conductivity zone is nearer the surface beneath oceans than beneath continents.

Significant lateral variations exist in terrestrial electrical conductivity. The best-known variations occur at ocean–continent interfaces and are produced by the large conductivity of seawater. Water-saturated sediments can also produce shallow conductivity variations. Regional variations such as these are referred to as *conductivity anomalies* (or CAs). Not all conductivity anomalies can be explained by seawater or water-saturated sediments, and they appear to have their sources deep in the crust or upper mantle. Two types of such anomalies related to short-period magnetic variations have been identified. One is characterized by variations in the ratio of the vertical to the horizontal components of the field ($\Delta Z/\Delta H$), and the other by variations in the vertical component only (ΔZ). The first type of CA is characteristic of convergent plate boundaries and has been described in Japan and in the Andes in South America (Honkura, 1978). The second type results from shallowing of highly conductive layers and is found beneath both oceanic and continental rift systems as well as within continental lithosphere. An example of the first type of CA in Japan is shown in figure 3.8. The highly conducting layer increases in depth beneath eastern Japan to greater than 200 km, and then decreases to depths less than 30 km beneath the Japan Sea. Examples of the second type of CA are shown in figure 3.9. In this figure, three major anomalies are shown (intense stippling): one extends south from the north end of the southern Rockies beneath the Rio Grande rift; another trends southwest along the Wasatch Front in Utah; and a third occurs beneath the southern Canadian Rockies. In general, upper-mantle conductivities are higher in the western United States, and especially beneath the Basin and Range Province, than in the midwestern and eastern United States, as shown by the less dense stippling in the figure. The depths of conducting layers in the mantle and crust vary inversely with surface heat flow (Adam, 1976). The main conducting layers cluster around three depth ranges of 300–700 km, 50–300 km and 10–40 km.

The origin of high-conductivity layers in the upper mantle and crust is related to partial melting, increases in water content, changes in composition, preferential orientation of minerals or a combination of these factors. Increased temperature and partial melting appear to be most important in producing upper-mantle anomalies. The increase in depth of high-conductivity zones beneath trenches appears to reflect depression of isotherms (see fig. 3.16) produced by subduction (Honkura, 1978). The shallow depth of high-conductivity layers above descending slabs, such as beneath the Philippine and Japan Seas, is related both to higher temperatures and to partial melting above such slabs. The shallowing of conductivity anomalies beneath ocean ridges and continental rift systems probably reflects upwelling asthenosphere from the LVZ or mantle plumes which have thinned the lithosphere. A relatively high conductivity zone at shallow depths beneath ocean ridges and continental rifts corresponds to a region of low V_P which, in turn, reflects a widespread shallow depth of partial melting (Tarits, 1986).

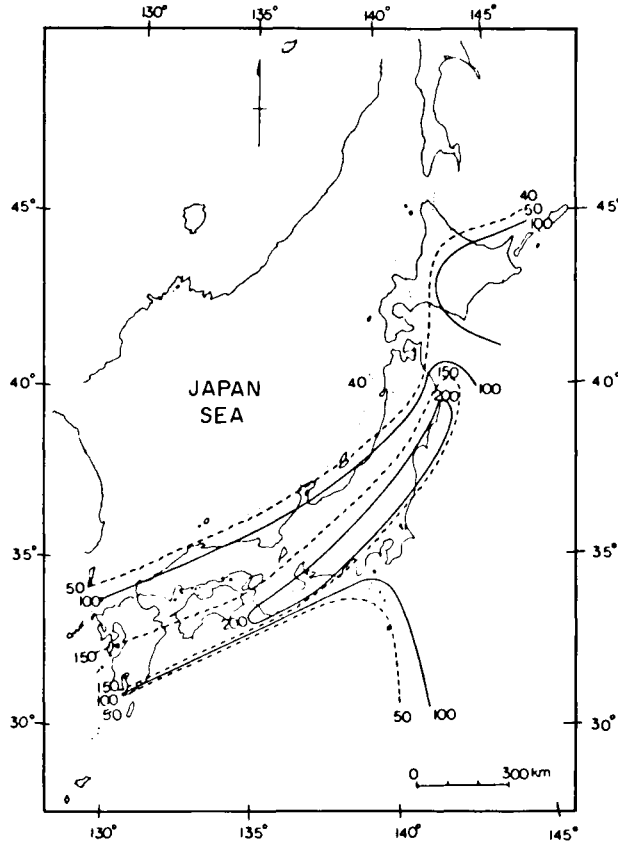


FIG 3.8 Contoured depth (in km) of a highly conducting zone beneath Japan as deduced from variations in the ratio of the vertical to the horizontal component of the Earth's magnetic field. From Rikitake (1969).

Global Gravity

Gravity Anomalies

Global free-air gravity anomalies calculated from spherical harmonic coefficients of the gravitational field of degrees 2 through 16 are shown in figure 3.10. Most of the large-scale gravity features of the Earth can be correlated with lithosphere-asthenosphere interactions or with mantle convection (Phillips and Lambeck, 1980). Positive anomalies in oceanic areas are located close to ocean ridges but the converse is not true. In fact, part of the Pacific–Antarctic ridge system exhibits a negative anomaly. Large positive anomalies in oceanic areas probably reflect upwelling mantle convection currents or plumes that bring dense mantle material close to the surface. Some negative anomalies also characterize plumes, however. This may be due to two competing processes of the same order of magnitude in the upper mantle: (a) dense material brought close to the surface (favoring a positive anomaly), and (b) higher temperatures in the rising current causing a mass deficit (favoring a negative anomaly). Either process may dominate, and Iceland is an example of the former case and the Galapagos Islands (anomaly not resolved in fig. 3.10) of the latter.

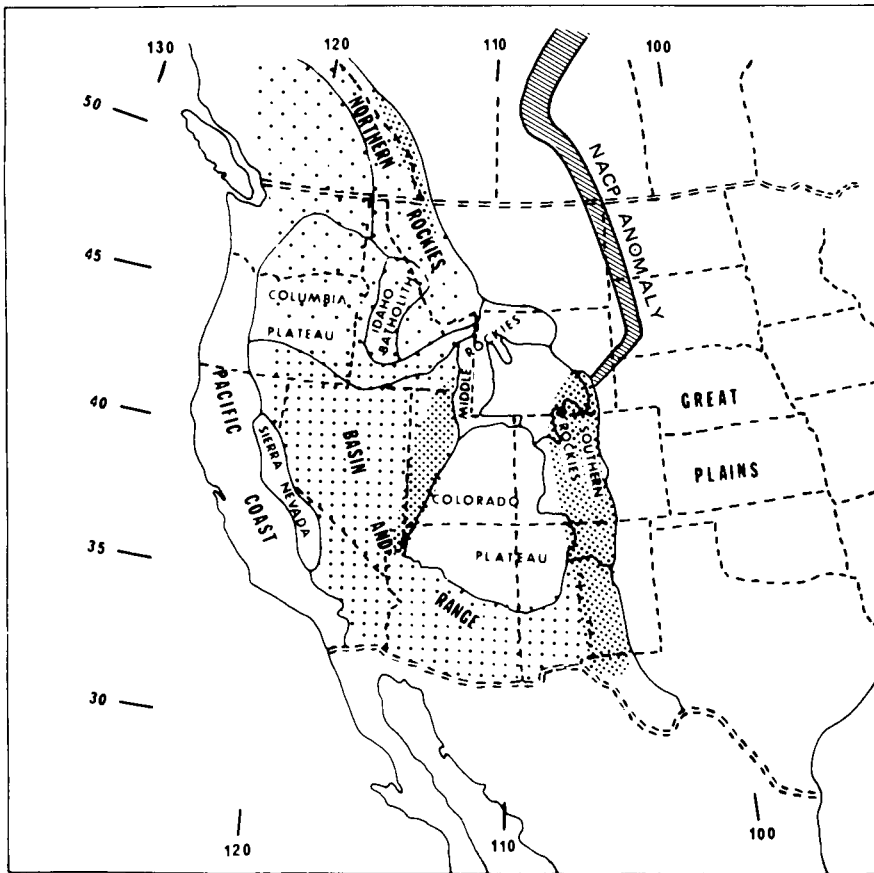


FIG 3.9 Conductivity distribution in the upper mantle in western North America. After Gough (1973) and Camfield and Gough (1976). Areas with no pattern are regions of very low conductivity; intensity of stippling in other areas is proportional to intensity of conductivity.

Major negative anomalies characterize ocean basins and appear to reflect maximum horizontal acceleration of lithosphere and asthenosphere away from ocean ridges. There is not, however, a systematic relationship between the distribution of these anomalies and ocean depth after the cooling lithosphere effect has been subtracted. Trench-arc systems also seem to be associated with positive anomalies, as is exemplified by the Indonesia area and Japan. Gravity model studies indicate that the downgoing slab contributes very little to these anomalies. Rather, they appear to result from a combination of elevated back-arc lithosphere and a low topographic rise on the seaward side of most trenches. This rise is produced by buckling of the oceanic lithosphere as it begins to descend into the mantle.

Major negative anomalies on continents are commonly associated with regions that are isostatically rising after the retreat of Pleistocene glaciers, as for example, northern Canada. In these areas relatively light lithosphere is depressed into the asthenosphere, producing negative anomalies. Rapid recovery rates will greatly decrease or remove these anomalies in a few thousand years. The large negative

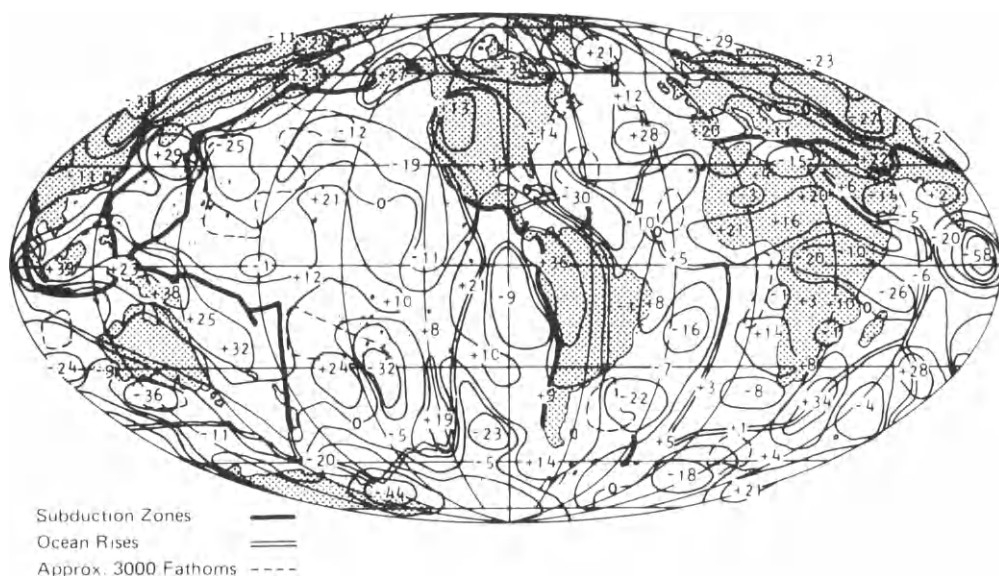


FIG 3.10 Global free-air anomalies (in mgal) referred to an ellipsoid of flattening of $1/299.8$. After Kaula (1972). Used with permission of McGraw-Hill Book Co.

anomalies in south-central Asia in the Hindu Kush–Tien Shan area and over Antarctica seem to have more complex explanations. Calculations indicate that thickening of the crust or lithosphere resulting from a continent–continent collision cannot alone account for the negative anomaly in south-central Asia (Kaula, 1972). The negative anomaly in Antarctica, which is almost completely surrounded by active ocean ridges, cannot readily be explained in terms of either Pleistocene downwarping or lithosphere thickening produced by insufficient asthenospheric flow. It may reflect convection in deeper parts of the mantle.

Geoid Anomalies

Recent developments in seismology and theoretical dynamics give new insight into the interpretation of long-wavelength components of the Earth's gravity field. Seismic velocity distributions in the Earth reflect the density distribution which, in turn, influences the *geoid* (the topographic shape of the Earth's surface). Geodynamic theories indicate that a negative correlation should exist between density variation and geoid elevation in a convecting Earth (Hager et al., 1985). Hot upwellings in the mantle should cause uplift of the Earth's surface and of the core–mantle boundary. This dynamically maintained topography has a strong effect on the gravity field which, in turn, reflects the viscosity distribution and presence or absence of stratification in the mantle. In principle, mantle-wide convection can be distinguished from layered convection by the distribution of large-scale topography at the Earth's surface.

The calculated geoid of Earth is shown in figure 3.11a. Geoid highs over Africa and the central Pacific reflect hotter (less dense) than average mantle. Hotspots, presumably reflecting rising mantle plumes, are also frequent in these regions. The

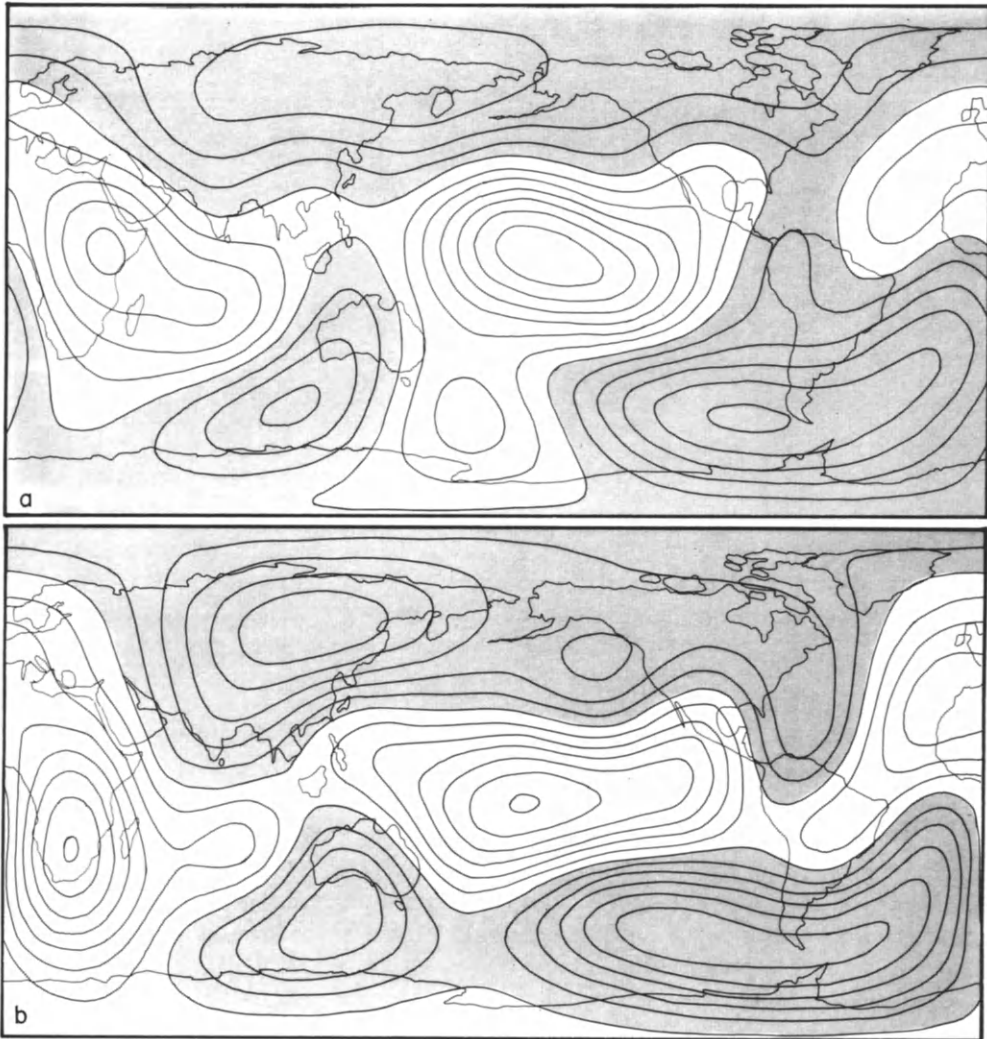


FIG 3.11 (a) The calculated geoid for a convecting Earth. Density contrast assumed equal to 2.3 gm/cm^3 , contour interval 200 m, and degree 1–6. (b) Calculated topography at the core–mantle boundary. Density contrast across the boundary assumed to be 4.5 gm/cm^3 , contour interval 400 m, and degree 1–6. Geoid lows are shaded. After Hager and Clayton (1987).

calculated topography of the core–mantle boundary is shown in figure 3.11b. Both the core–mantle boundary and Earth’s surface are upwarped in regions of low seismic velocities and downwarped in regions of high velocities (compare to fig. 3.6). These differences are probably caused by temperature differences of convective origin. Mantle-wide convection provides the simplest explanation of why surface highs tend to overlie hot lower mantle as reflected by topographic highs on the core–mantle boundary.

Composition of the Mantle

The mineralogical and chemical composition of the mantle can be approximated from the combined results of seismic velocity distributions in the Earth, high-

TABLE 3.1 *Ultramafic and Mafic Rocks*

Rock	Principal Minerals
1. Ultramafic	
Lherzolite	ol, cpx, opx
Harzburgite	ol, opx, } $\pm$ gar, $\pm$ sp, $\pm$ plg, $\pm$ amp
Wehrlite	ol, cpx
Dunite	ol
Pyroxenite	px
Serpentinite	serp
Kimberlite	ol, px, gar, serp, mica, ilm
2. Mafic	
Gabbro	plg, cpx, ol
Garnet granulite	plg, px, gar
Eclogite	cpx, gar

ol, olivine; px, pyroxene; cpx, clinopyroxene; opx, orthopyroxene; gar, garnet; sp, spinel; plg, plagioclase; amp, amphibole; serp, serpentine; ilm, ilmenite.

pressure-temperature experimental studies, shock-wave experiments, and geochemical and isotopic studies of meteorites and ultramafic rocks. It is possible to attain static pressures in the laboratory of 300 kbar to more than 1.5 Mbar, which allows direct investigation of the physical and chemical properties of minerals that may occur throughout most of the Earth. Shock-wave studies have also been important in defining stability fields of possible mantle mineral assemblages at depths greater than 600 km.

Several lines of evidence indicate that *ultramafic rocks* compose large parts of the upper mantle. These rocks contain more than 70% of iron- and magnesium-rich minerals such as pyroxenes, olivine and garnet. *Mafic rocks* are of basaltic composition and are composed of various mixtures of plagioclase, pyroxenes, olivine, amphibole, and garnet, and may also occur in parts of the upper mantle, although probably in small amounts. Principal minerals in important ultramafic and mafic rocks are summarized in table 3.1.

Geochemical, isotopic and seismic studies all agree that the mantle is heterogeneous. Scales and causes of heterogeneities, however, are poorly understood and published models range from a strongly stratified mantle to an unstratified "plum pudding" mantle. The study of radiogenic daughter isotopes in basalts that have been derived from the mantle clearly indicates chemical heterogeneity in the mantle (see Chapt. 7). The results suggest the existence of distinct mantle reservoirs that have been in existence for more than 10^9 years. The sizes, shapes and locations of these reservoirs, however, are not well constrained by the isotopic data and it is only through increased resolution of seismic velocity distributions that these reservoirs can be more accurately characterized.

Measured P-wave Velocities at High Pressures

Laboratory measurements of V_P at pressures up to 20 kbar provide valuable data regarding possible mineral assemblages in the upper mantle (Christensen, 1966). P-wave velocities in rocks increase with increasing pressure in response to (a) the intrinsic effect of pressure on individual minerals, (b) compression of minerals, and

TABLE 3.2 *P-wave Velocities and Densities at 10 kbar*

	Density (1 bar) gm/cm ³	V _P (10 kbar) km/sec
Dunite	3.2–3.3	8.0–8.3
Lherzolite	3.2–3.3	8.0–8.3
Garnet lherzolite	3.3–3.4	8.0–8.4
Eclogite	3.4–3.5	8.1–8.5
Garnet granulite	3.2–3.3	7.8–7.9
Amphibole lherzolite	3.0–3.2	7.7–7.9
Plagioclase lherzolite	3.0–3.2	7.7–7.9
Serpentinized lherzolite (13%)	3.2	7.8
Serpentinite (100% serpentine)	2.5	5.2

Note: Velocities calculated if direct measurements not available.

After Birch (1960); Christensen (1966); Kumazawa et al. (1971).

(c) progressive closing of pore spaces. An increase in temperature above room temperature will decrease measured velocities. These effects are approximately self-canceling, such that V_P measured at room temperature and 10 kbar pressure will be only <0.1 km/sec larger than actual velocities in the Earth at 10 kbar (about 35 km) (Kumazawa et al., 1971). Ranges of P-wave velocity at 10 kbar and rock densities at 1 bar for various mafic and ultramafic rocks are summarized in table 3.2. Dunite, herzolite, garnet/herzolite and eclogite or some mixture of these rock types are consistent with observed P_n velocities beneath the Moho in most areas. Higher than normal temperatures and/or some combination of garnet granulite, or amphibole, plagioclase, or serpentinized/herzolite, may account for regions with low P_n velocities. The high temperatures characteristic of most of these regions also exceed the stability of serpentine, thus greatly limiting the extent of even slightly serpentinized lherzolite in the mantle.

P-wave velocity measurements at various orientations to rock fabrics substantiate the idea that differences in mineral alignment can produce significant anisotropy (Ave'Lallemant and Carter, 1970; Kumazawa et al., 1971). Differences in V_P of more than 15% occur in some ultramafic samples and are related primarily to the orientation of olivine grains. Anisotropy in the upper mantle beneath ocean basins may be produced by recrystallization of olivine and other minerals accompanying seafloor spreading with the [100] axes of olivine and [001] axes of orthopyroxene statistically oriented normal to ridge axes (the higher V_P direction) (Estey and Douglas, 1986). Evidence for alignment of these minerals comes from studies of ophiolites (Chapt. 6) and upper-mantle xenoliths. The mechanism of mineral alignment requires upper-mantle shear flow which aligns minerals by dislocation glide. The crystallographic glide systems have a threshold temperature necessary for recrystallization of about 900°C, which yields a thermally defined lithosphere depth similar to that deduced from seismic data (~ 100 km). Mineral alignment is actively maintained below this boundary in the LVZ and is in a fossil state in the overlying lithosphere.

Experimental Data Related to the Moho

Two rival hypotheses exist for the origin of the density change that is responsible for the Moho. One is that it is produced by a phase change from a low- to a high-

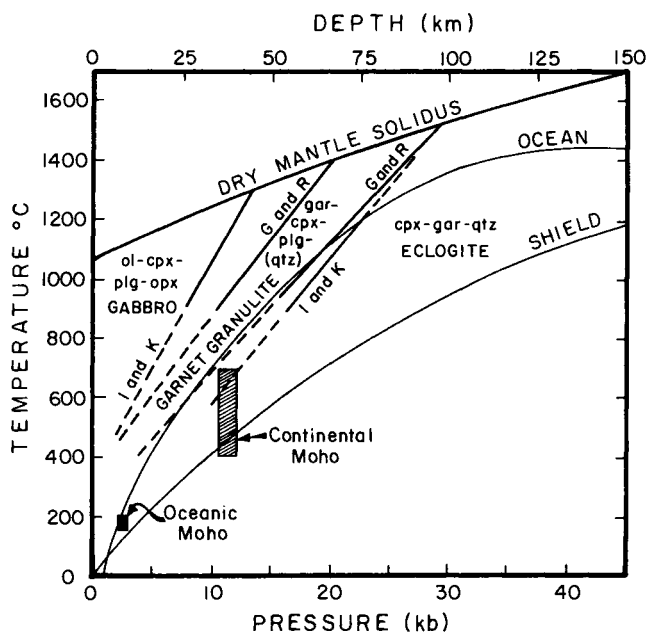


FIG 3.12 Pressure-temperature fields of mafic mineral assemblages. Also shown are oceanic and Precambrian shield geotherms and P-T ranges of the average oceanic and continental Moho. Oceanic geotherm is for a region 200 km from an ocean ridge and a spreading rate of 4 cm/a. Boundaries of the garnet granulite field are after Ito and Kennedy (1971) (I and K) and Green and Ringwood (1972) (G and R). Abbreviations: ol = olivine; cpx = clinopyroxene; opx = orthopyroxene; gar = garnet; plg = plagioclase; qtz = quartz.

pressure mineral assemblage, and the other is that it represents a compositional (and phase) change. The complexities in defining the Moho in some areas, as well as the differing temperatures and pressures of the oceanic and continental Moho (fig. 3.12), suggest that it does not have a single origin.

One origin proposed for the Moho is that it represents a phase change from serpentinite in the lower crust to lherzolite in the upper mantle (Hess, 1962). Experimental data, however, seem to eliminate this mechanism as a possible cause for either the oceanic or continental Moho. The experimentally determined univariant reaction curves for the formation of serpentine or serpentine plus brucite from Mg-rich olivine intersect oceanic geotherms at depths much greater than observed Moho depths. Also, average seismic-wave velocities in the lower crust are rather constant, and would necessitate a fortuitously constant degree of serpentinization over much of the Earth.

The question as to whether or not the continental Moho represents a phase change from gabbro in the lower crust to eclogite in the upper mantle has been a subject of considerable research. Experimental data indicate that a garnet granulite transition zone 5–10 kb wide exists between gabbro and eclogite at 1100°C (fig. 3.12). The upper and lower stability boundaries of the garnet granulite assemblage at continental Moho depths depend critically on the extrapolation of high-temperature data, since metastable plagioclase and garnet nucleation problems have, thus far, prevented accurate determination of these boundaries at temperatures lower than

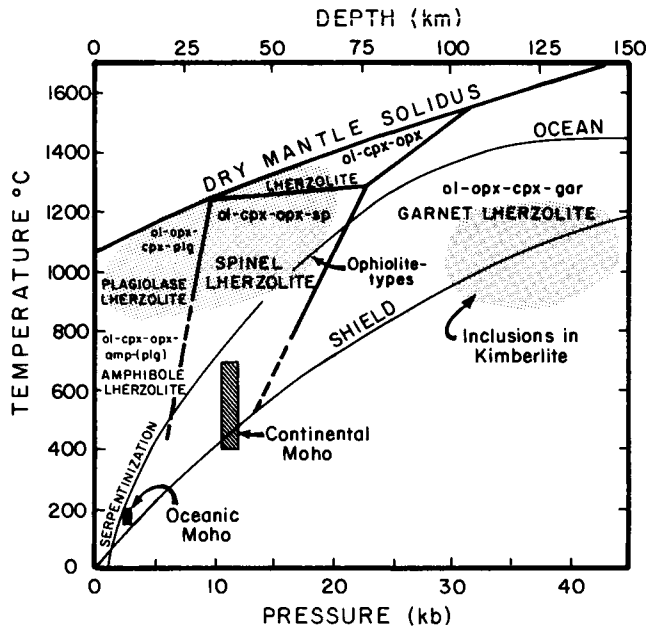


FIG 3.13 Pressure-temperature fields of ultramafic mineral assemblages (modified after Ringwood, 1969b). Inferred minimum depths of formation of ophiolite-type ultramafics and of garnet lherzolite inclusions in kimberlite (as deduced from clinopyroxene compositions) are also shown. Abbreviations: amp = amphibole; sp = spinel; other information as given in fig. 3.12.

about 1000°C. Estimates of the gradient for these boundaries range from 10 bar/°C (Ito and Kennedy, 1971) to 21 bar/°C (Green and Ringwood, 1972); extrapolations of each are shown in figure 3.12, together with temperature and pressure estimates of the continental and oceanic Moho. The lower gradient extrapolation suggests that the transition from garnet granulite to eclogite may occur beneath some parts of the continental crust, while extrapolation of the steeper gradient suggests that eclogite is stable throughout the continental crust. However, eclogite will not form below about 500°C due to very small reaction rates, and hence garnet granulite is the stable mafic mineral assemblage throughout the lower continental crust. The width of the extrapolated garnet granulite field at continental Moho temperatures using either gradient is too great to explain the sharpness of the Moho in most areas. Also, the high seismic-wave velocities in eclogite severely limit its abundance in the crust. The absence of a correlation between surface heat flow and crustal thickness does not favor a gabbro-garnet granulite-eclogite phase change at the Moho, since surface heat flow would reflect temperature distribution at depth and, hence, phase transition depths. Furthermore, it is difficult to see how eclogite could compose a major part of the near-Moho upper mantle, in that its average density (3.5) exceeds that of garnet lherzolite (3.4) upon which it probably rests (fig. 3.13); hence it would tend to sink. In subduction zones, however, phase transitions of mafic rocks to eclogite enhance the sinking rate of the consumed plate.

More experimental studies are needed to locate accurately the garnet granulite phase boundaries at lower temperatures. However, it seems clear that the Moho

cannot represent a serpentinite–lherzolite phase change, nor a single phase change from gabbro to eclogite. As will be discussed below and in the next chapter, a composition change from a lower crust of mafic or intermediate composition to an upper mantle composed dominantly of lherzolite is most consistent with existing data.

Upper-mantle Mineral Assemblages

Possible upper-mantle mineral assemblages of ultramafic compositions are shown in figure 3.13. This figure summarizes phase equilibria studies at high pressures and temperatures. The *mantle solidus* — i.e. the temperature and pressure at which melting begins in the mantle — is assumed to be that of lherzolite. Oceanic and shield geotherms and estimates of temperature and pressure for the oceanic and continental Moho are also shown.

The data in figure 3.13 are consistent with an upper mantle beneath the continents composed chiefly of garnet lherzolite. Stable ultramafic mineral assemblages beneath oceanic areas are dependent upon distance to active ocean ridge systems. Most estimates of the range of oceanic geotherms are, however, consistent with an oceanic upper mantle at shallow depths (25–75 km) composed dominantly of spinel lherzolite or plagioclase lherzolite and at greater depths, garnet lherzolite. Serpentinized lherzolites are produced at rather low temperatures and pressures (100–400°C; ≤ 25 km), probably during tectonic emplacement.

Ultramafic Rocks

There are five major occurrences of ultramafic rocks in the Earth's crust (after Wyllie, 1970):

1. ultramafic zones of stratiform intrusions
2. dredge samples from oceanic crust
3. ophiolites
4. inclusions in basalt and kimberlite
5. ultramafic flows and sills.

Stratiform intrusions are produced by fractional crystallization of basaltic magmas emplaced in the crust, and the ultramafic zones represent early differentiates. Little is known about the occurrence of lherzolites and serpentinites dredged from the ocean floors. Models for the oceanic crust, however, suggest that oceanic ultramafic rocks occur as mantle diapirs that have penetrated the oceanic crust, perhaps along transform faults. Most of the samples are altered and serpentinized in varying degrees, and it seems doubtful that many or perhaps any represent fresh samples of the upper mantle beneath oceans.

Ultramafics from ophiolites (see Chapt. 6) are a rather heterogeneous group and include bodies of varying origins and modes of emplacement. Most appear to represent fragments of oceanic upper mantle that have been emplaced on continental (or arc) crust by faulting or overthrusting associated with subduction or plate collisions. Some may represent tectonically emplaced diapirs from the mantle. A

large proportion of ophiolite ultramafics have been altered, metamorphosed, deformed, and serpentinized in varying degrees. For this reason, extreme care must be used in interpreting the composition of these bodies in terms of the composition of the upper mantle.

Ultramafic inclusions are common in alkali basalts and in kimberlite pipes. Geochemical and mineralogical data indicate that inclusions of more than one origin may occur in the same lava flow or pipe. Some represent early fractionates of the host rock, while most appear to represent xenoliths of the upper mantle that were picked up by magmas at or near their source or during their ascent through the mantle.

Ultramafic flows and sills are rare. They occur in Archean volcanic terranes (i.e., komatiites, Chapt. 9) and in some ophiolite complexes. Many of them appear to be Mg-rich basalts containing a large quantity of olivine crystals, and hence should not be considered as complete melts of upper-mantle material.

Experimental studies of clinopyroxenes coexisting with olivine, orthopyroxene, and an Al-rich phase (such as spinel or plagioclase) indicate that Al, Mg, and Ca content of clinopyroxene is sensitive to the temperature and pressure at which it crystallized (or recrystallized) (MacGregor, 1974). Hence, compositional data from clinopyroxenes in four-phase ultramafic rocks provide information on minimum depths of origin for these rocks. Results from a suite of ultramafic inclusions from a given locality may define a P-T curve which is thought to represent a *paleogeotherm* (Boyd, 1973). Such geotherms are similar to those estimated from geophysical and geochemical considerations. The depth ranges recorded by clinopyroxenes in ultramafics from ophiolites and by garnet lherzolite xenoliths from kimberlites are outlined in figure 3.13. Ophiolite ultramafic rocks record depths of at least 10–75 km (some of these are probably recrystallization depths) and those from garnet lherzolites record depths greater than 100 km. Ultramafic inclusions from alkali basalts fall between and significantly overlap these two groups.

Textures and structures in ultramafic rocks from ophiolites and mantle xenoliths, when interpreted together with geochemical and experimental petrologic data, suggest that specific mantle environments can be identified from the study of these ultramafic rocks. Flow patterns in the oceanic upper mantle can be directly studied by structural mapping of ophiolite ultramafics (Nicholas, 1986). Ophiolites are fragments of oceanic lithosphere created at ocean ridges and later tectonically emplaced in continental crust (Chapt. 6). Because they sample at most the first few kilometers of the mantle, they represent asthenosphere 'frozen' within the first few millions of years of seafloor spreading. Fast-spreading ocean ridges (>1 cm/a) are characterized by harzburgite ($\pm$ dunite) from which $\geq 20\%$ basaltic melt has been extracted (Bottinga and Allegre, 1978). Foliations and lineations in the ultramafic rocks are gently dipping and may actually reflect subhorizontal asthenosphere flow. Slow-spreading ridges (≤ 1 cm/a) are characterized by anomalously thin oceanic crust (≤ 5 km), less melt extraction ($>15\%$) and a residual mantle of plagioclase lherzolite. Foliation and lineations in the ultramafic rocks are steep and may actually map mantle diapirs that have risen beneath ocean ridges. Xenoliths from alkali basalts extruded in continental rifts are generally spinel lherzolites and the most common ultramafic xenoliths in kimberlites are garnet lherzolites. Both of these xenolith populations appear to have less than 10% basaltic melt extracted from them.

TABLE 3.3 *Estimates of the Composition of the Mantle*

	1	2	3	4	5	6
SiO ₂	44.5	44.2	45.2	46.0	48.1	43.2
MgO	41.7	41.3	37.5	38.1	37.1	38.1
FeO	8.7	8.3	8.4	8.6	12.7	9.3
Al ₂ O ₃	2.6	2.7	3.5	3.6	3.0	3.9
CaO	2.2	2.4	3.1	3.1	2.3	3.7
Na ₂ O	0.3	0.3	0.6	0.6	1.1	1.8

1. Average of 168 ultramafic rocks (White, 1967).
2. Average of oceanic serpentinites (calculated water-free) and high-Ca-high-Al olivine inclusions (Harris, Reay, and White, 1967).
3. Pyrolite—a 1:3 mixture of basalt (Hawaiian tholeiite) with a synthetic lherzolite (Ringwood, 1975).
4. Mixture of 40% Type I carbonaceous chondrites, 50% ordinary chondrites, and 10% iron meteorites such that mantle/core ratio equals 68/32; mantle recalculated to 100% (after Murthy and Hall, 1970).
5. Mixture of 32.4% iron meteorite (with 5.3% FeS) and 67.6% oxide portion of bronzite chondrites; mantle recalculated to 100% (Mason, 1966).
6. Nonvolatile portion of Type I carbonaceous chondrites with $\text{FeO}/\text{FeO} + \text{MnO} = 0.12$ (Ringwood, 1975).

Chemical Composition of the Mantle

Several approaches have been used to estimate the composition of the mantle: (1) using the compositions of various groups of ultramafic rocks; (2) using theoretical compositions calculated from geochemical considerations; (3) using compositions of various meteorite mixtures; and (4) using data from shock-wave studies and from ultra-high-pressure experimental studies. Two estimates of the major-element composition of the upper mantle based on various averages of ultramafic rocks are given in table 3.3 (columns 1 and 2) for comparison. Also shown is a theoretical upper-mantle composition (column 3) defined by the property that upon partial melting it will yield a basaltic magma and leave behind a refractory residue of harzburgite (Ringwood, 1975). Columns 4-6 show estimates of mantle composition produced by mixing various meteorite classes in appropriate ratios to give the core/mantle mass ratio of 32/68. Although variability exists between the estimates of mantle composition, it is clear that more than 90% of the mantle by weight can be expressed in terms of the system FeO–MgO–SiO₂ and that no other oxide exceeds 4%. It is also noteworthy that the two estimates of upper-mantle composition based on ultramafic rocks exhibit lower Al₂O₃ and Na₂O and higher MgO than the meteorite or theoretical estimates. This may reflect a real depletion in aluminum in the upper mantle today, since the meteorite and theoretical estimates are designed to approximate more closely the composition of undepleted, primitive mantle. However, changes in composition during emplacement of the ultramafic rocks may also contribute to these apparent element differences.

The Sr, Nd, and Pb isotopic compositions of basalts from ocean basins indicate the existence of at least three types of isotopically distinct mantle reservoirs. The most widespread source is the depleted source from which ocean ridge basalts (MORB) are derived. Isotopic data, furthermore, indicate that this source has existed for at least 2 Ga. In comparison to the MORB source, all other mantle reservoirs are enriched in incompatible elements. The geographic distribution of isotopic and trace element data support the existence of both small (a few km³) and

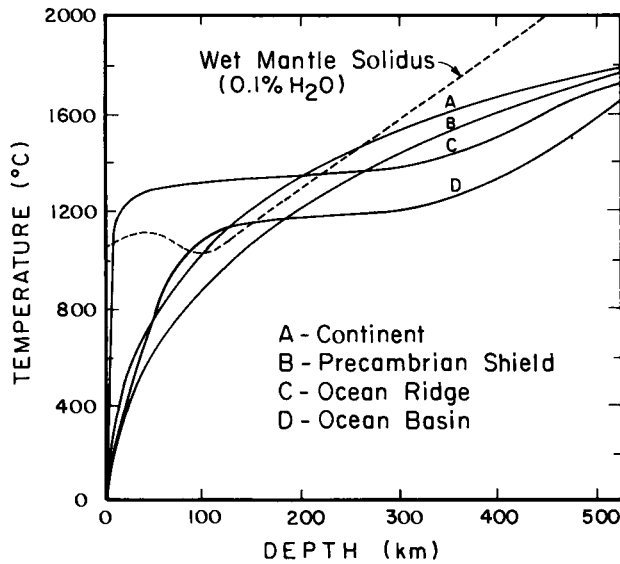


FIG 3.14 Pressure-temperature diagram showing the wet mantle solidus (0.1% H₂O) and average geotherms for different tectonic environments.

large ($\geq 10^9$ km³) heterogeneities in the mantle. Just how these heterogeneities formed and survived in a convecting mantle is a subject of much disagreement and debate, and will be further discussed in Chapters 7 and 10.

The Low-velocity Zone

The low-velocity zone (LVZ) is characterized by low seismic wave velocities, low Q and high electrical conductivity. Any hypothesis for the origin of the LVZ must explain these features. Among the causes considered for the low seismic wave velocities are a phase change, a composition change, alignment of olivine crystals, and incipient melting. No important phase changes are known to occur in this depth range, and compositional changes (such as an increase in the amount of garnet granulite) either do not seem probable geochemically or do not readily explain the low Q and high electrical conductivity. Experimental and theoretical studies indicate that both alignment of olivine crystals ([100] axes) and incipient melting can result in lower seismic-wave velocities, increased attenuation and high electrical conductivity (Wyllie, 1971b; Estey and Douglas, 1986). The rather sharp boundaries of the LVZ are also consistent with one or both of these mechanisms.

The probable importance of incipient melting in the LVZ is attested to by the high surface heat flow observed when the LVZ reaches shallow depths, such as beneath ocean ridges. Experimental data show that, to produce incipient melting in this region, a minor amount of water is required to depress silicate melting points (Wyllie, 1971b). Melting conditions for ultramafic mantle compositions under slightly hydrous conditions (0.1% water) are shown in figure 3.14. Various geotherms are also given for reference. It is clear that even with only 0.1% water, partial melting of garnet lherzolite occurs in the appropriate depth range for the

LVZ. The source of water in the upper mantle may be from the breakdown of minor mantle phases that contain water, such as hornblende, mica, titanoclinohumite, or hydrated silicates. The amount of melt produced is very small and is almost a direct function of the water content. The theory of elastic wave velocities in two-phase materials indicates that only 1% melt is required to produce the lowest velocities measured in the LVZ (Anderson et al., 1971). If, however, melt fractions are interconnected by a network of tubes along grain boundaries, the amount of melting may exceed 5% (Marko, 1980).

The downward termination of the LVZ may be caused by one or a combination of several effects:

1. rapid decrease in the amount of water available;
2. water enters high-pressure hydrous silicate minerals (such as hydrous pyroxenes and garnets);
3. rapid increase in the creep strength of olivine such that crystals are no longer aligned by asthenosphere flow; and
4. the geotherm passes through the mantle solidus, as illustrated in figure 3.14.

The LVZ plays a major role in plate tectonics, providing a relatively low-viscosity region upon which lithospheric plates can slide with very little friction. Lateral motion of plates away from ocean ridges and consumption of plates in subduction zones are offset by return flow deeper in the mantle. The fact that the LVZ is absent or poorly developed beneath Precambrian shields suggests that the roots of shield areas that may extend to 400 km depth, may cause drag on lithospheric plates. Calculations, however, indicate that this drag is small compared to the drag produced in subduction zones.

The Transition Zone

The 400 km Discontinuity

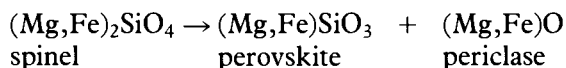
High-pressure experimental studies document the breakdown of Mg-olivine to a high-pressure phase known as β -phase at about 118 kb which is equivalent to 400 km burial depth in the Earth. Mantle olivine (Fo_{90}) completely transforms to β -phase over about a 2 kb pressure range at appropriate temperatures for the 400 km discontinuity ($\sim 1000^\circ\text{C}$) (Bina and Wood, 1987). This pressure range, which is equivalent to about 7 km, is in excellent agreement with the width of the 400 km discontinuity deduced from seismic data. Because olivine is likely the most abundant mineral in the upper mantle, the olivine- β -phase transition is the probable explanation for the 400-km discontinuity. Comparison of the observed magnitude of the velocity jump at 400 km with that computed for pure olivine (Fo_{90}) suggests, furthermore, that the upper mantle is composed of about 70% olivine, an amount greater than previous estimates. Experimental data indicate that β -phase should transform to a more densely packed spinel structure at equivalent burial depths of 500–550 km. Only a small density change ($\sim 2.5\%$) accompanies this transition, and hence a seismic discontinuity is rarely resolved in this depth range.

High-pressure experimental data also indicate that at depths of 350–400 km, aluminous orthopyroxenes are transformed into garnet, involving a density increase

of about 10% (Ringwood, 1975). It is possible that the increase in velocity gradient commonly observed beginning at 350 km (fig. 3.2a) and leading up to the 400 km discontinuity is caused by such pyroxene transformations.

The 670 km Discontinuity

One of the most important questions related to the style of mantle convection in the Earth is whether the 670-km discontinuity represents a compositional change or a phase change. If the boundary represents a compositional change, two-layer mantle convection is favored with the 670-km discontinuity representing the base of the upper layer. Large increases in both seismic-wave velocity (4–7%) and density (6–9%) occur at this boundary. Another feature of the boundary is that reflections occur in some geographic regions and not in others, suggesting variable thickness or significant topographic variation of the boundary. The ultra-high-pressure experimental equipment of the last decade has made it possible to study phase transformations that may be responsible for the 670-km discontinuity. Data indicate that $(\text{Mg,Fe})_2\text{SiO}_4$ spinel (Fo_{90}) undergoes a transition at about 260 kbar and 1000°C to a mixture of orthorhombic perovskite and periclase as follows (Liu, 1976):



The reaction involves an increase in zero-pressure density of about 11% and corresponds closely with the observed depth of the 670-km discontinuity. At about the same pressure and temperature, pyrope garnet transforms to an ilmenite structure, which is followed by a transformation to an orthorhombic perovskite structure at about 300 kbar (Liu, 1979). These transformations involve increases in density of 7.9 and 7.7%, respectively. A combination of the spinel and garnet transformations to perovskite structural states could explain the 670-km discontinuity. The spinel transformation, however, should dominate due to the large amount of spinel (~70%). However, experimental data suggest that the spinel-perovskite phase region is too wide (equivalent to about 30 km) to explain the seismic thickness of only a few kilometers (Lees et al., 1983). This led to the idea that the transition represents a compositional (and phase) change in which the lower mantle beneath the discontinuity is enriched in iron and perhaps in silicon (Knittle et al., 1986). However, problems exist in matching changes in chemical composition to seismic data. Also, because the spinel-perovskite phase diagram is not precisely known, results may not accurately constrain the width of the transition. Although the spinel-perovskite transition must at least partially account for the 670-km discontinuity, until we have more accurate high-pressure experimental data available, it is not possible to resolve the question of a chemical compositional change also occurring at this boundary.

Summary of Mineralogical Changes to 1000 km

Possible mineral assemblages in the mantle to 1000 km depth and mineral transformations as determined largely from experimental data are summarized in table 3.4. The upper mantle to 400 km is composed chiefly of olivine, pyroxenes,

TABLE 3.4 *Summary of Possible Mineral Assemblages of the Upper Mantle and Transition Zone*

	Depth Range (km)	Mineralogy	%*	Density (zero pressure)
Upper Mantle	up to 350	Olivine, $\text{Fo}_{90}(\text{ol})(\text{Mg,Fe})_2\text{SiO}_4$ Orthopyroxene $(\text{opx})(\text{Mg,Fe})\text{SiO}_3$ Clinopyroxene $(\text{cpx})(\text{Ca,Mg,Fe})_2(\text{NaAl})(\text{Si}_2\text{O}_6)$ Pyrope garnet $(\text{gar})(\text{Mg,Fe,Ca})_3(\text{Al,Cr})_2\text{Si}_3\text{O}_{12}$	70 15 5 10	3.4
400-km Discontinuity	350-400 400-430	$(\text{Mg,Fe,Al})\text{opx} \rightarrow (\text{Mg,Fe})\text{gar}$ $(\text{Mg,Fe})\text{ol} \rightarrow (\text{Mg,Fe})\beta\text{-phase}$		
Upper Transition Zone	400-550	$(\text{Mg,Fe})\beta\text{-phase}$ $(\text{Mg,Fe,Ca})\text{gar}$ $(\text{NaAl})\text{cpx}$ $(\text{Ca})\text{gar} \rightarrow \text{perovskite}$	70 27 3	3.6
	500-550	$(\text{Mg,Fe})\beta\text{-phase} \rightarrow (\text{Mg,Fe})\text{spinel}$		
	550-670	$(\text{Mg,Fe})\text{spinel}$ $(\text{Mg,Fe})\text{gar}$ $(\text{Ca})\text{perovskite}$ $(\text{NaAl})\text{cpx}$	70 25 2 3	3.7
670-km Discontinuity		$(\text{Mg,Fe})\text{spinel} \rightarrow (\text{Mg,Fe})\text{perovskite} + (\text{Mg,Fe})\text{periclase}$ $(\text{Mg,Fe})\text{gar} \rightarrow (\text{Mg,Fe,Al})\text{ilmenite}$ $(\text{NaAl})\text{cpx} \rightarrow (\text{NaAl})\text{Ca-ferrite} + (\text{SiO}_2)\text{stishovite}$		
Lower Transition Zone	700-1050	$(\text{Mg,Fe})\text{perovskite}$ $(\text{Mg,Fe})\text{periclase}$ $(\text{Mg,Fe})\text{ilmenite}$ $(\text{Ca})\text{gar}$ $(\text{NaAl})\text{Ca-ferrite}$ $(\text{SiO}_2)\text{stishovite}$	70 20 10	4.0

*Estimated weight percent of each mineral. After Ringwood (1975), Liu (1979), Lees et al. (1983), and Bina and Wood (1987).

and garnet. Incipient melting of these minerals, probably caused by the presence of a small amount of water, produces the LVZ underlying the lithosphere except in Precambrian shield areas. The change in velocity gradient beginning at about 350 km may reflect transformation of Al-rich orthopyroxene to garnet. The 400-km discontinuity, recording a density change from 3.4 to 3.6 gm/cm³, is probably caused by the olivine $\rightarrow$ β -phase transition. The uppermost part of the transition zone is a rather homogeneous region composed chiefly of β -phase, garnet and minor Na-clinopyroxene. If excess Al₂O₃ is available, the Mg-Fe-Ca component of the clinopyroxene may be incorporated in the garnet solid solution; if not, it would transform to a high-pressure phase as yet unidentified. Ca garnet may also invert to cubic perovskite (CaSiO₃), which is stable in this depth range. At 500–550 km, β -phase will transform to spinel, involving only a 2.5% increase in density. The lower part of the transition zone is probably composed chiefly of Mg-Fe spinel and garnet, and has a density of about 3.7 gm/cm³.

The 670-km discontinuity, in part or entirely, reflects the transition from spinel ($\pm$ garnet) to perovskite and periclase structures. Na-clinopyroxene may also break down to a phase with a Ca-ferrite structure plus stishovite at about this depth, although contributing less than 1% to the density increase. In addition, an increase in $\text{Fe} \pm \text{Si}$ may occur at this boundary. (Mg,Fe) ilmenite should invert to an orthorhombic perovskite structure at about 800 km. Hence, it would appear that the lower part of the transition zone in the mantle is composed largely of Mg–Fe-rich minerals with perovskite and periclase structures.

The Lower Mantle

High-pressure data and studies of density distribution in the lower mantle indicate that this region is about 5% more dense than an oxide mixture with the approximate bulk composition of garnet lherzolite. Two explanations have been offered for this increase in density:

1. further mineral transformations in the lower mantle to assemblages with a density greater than an isochemical oxide mixture, or
2. an increase in the Fe/Mg ratio in the lower mantle.

In principle, a choice between these alternatives can be made by comparing seismic elastic ratios obtained from seismic velocities with those estimated from shock-wave studies. However, uncertainties in estimating elastic-ratio temperature corrections for the lower mantle allow either (or both) of the above interpretations. Recent studies suggest, however, that only phase assemblages dominated by $(\text{Mg},\text{Fe})\text{SiO}_3$ perovskite have sufficiently high densities and seismic parameters to be compatible with the observed density and seismic-wave velocity distribution in the lower mantle. One or more seismic discontinuities occur at about 1000 km and several velocity gradients are observed at greater depths (fig. 3.5). In addition, one or two discontinuities occur in the D'' region at 2600–2900 km. These changes in the lower mantle may reflect further transformations to dense assemblages or/and changes in chemical composition. Evaluation of these alternatives awaits experimental studies with ultra-high-pressure equipment.

Temperature Distribution in and Beneath Plates

Although it is not possible to determine temperature gradients in the upper mantle directly, several sources of information allow estimates to be made, as shown in figure 3.14 for two oceanic and two continental areas. These gradients are based on surface heat flow measurements and models of heat production and thermal conductivity distributions with depth. Convection is assumed to be the dominant mode of heat transfer in the asthenosphere beneath oceanic areas with conduction dominating in the lithosphere.

Although major differences in temperature distribution exist in the upper mantle, it is necessary that all temperature gradients converge at depths of a few hundred kilometers, or large unobserved gravity differences would exist between continental and oceanic areas. Most results indicate that temperatures beneath ocean basins and

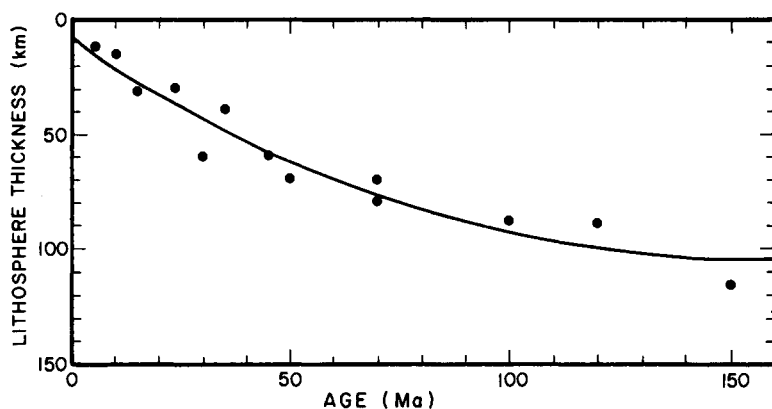


FIG 3.15 Thicknesses of the oceanic lithosphere estimated from surface wave dispersion studies and plotted against age of the lithosphere. Each point represents a different geographic location. Modified after Sacks (1983).

Precambrian shields converge between 100 and 200 km depth. Heat flow distribution, heat production models, and seafloor spreading models suggest that temperature gradients to about 50 km depth range from 10° to 30°C/km . Data suggest that the mantle is partially molten beneath the axes of ocean ridges (curve C, fig. 3.14) and that temperature gradients decrease with distance from ridges at a rate inversely proportional to lithosphere spreading rates (Bottinga and Allegre, 1973). With the exception of beneath ridge axes, temperature gradient must decrease significantly in the vicinity of 50–200 km in order to avoid large amounts of melting in the upper mantle, which are not allowed by seismic data. It is probable that the temperature gradients at depths of 100–200 km are close to the melting gradient of basalt, which is about 3°C/km at these depths.

The temperature gradient beneath ocean ridges decreases as the oceanic lithosphere spreads and cools, rapidly approaching the gradient typical of ocean basins (curve D). This decrease in geothermal gradient results in progressively greater depths of intersection of the geotherm with the mantle solidus, and hence to a thickening of the oceanic lithosphere as it ages. Cooling leads to oceanic lithosphere thicknesses of about 100 km in 100 Ma, which agrees well with thicknesses estimated from surface wave studies (fig. 3.15). The effect of continued cooling is relatively small such that, at the time of subduction, oceanic lithosphere is rarely greater than 150 km thick.

As relatively cool lithospheric slabs descend into the mantle at subduction zones, they begin to warm up. Possible heat sources for warming include conductive heat transfer from surrounding mantle, adiabatic compression of the slab, frictional heating along the upper surface of the slab, and exothermic phase changes in the slab (Toksoz et al., 1971). A typical calculated temperature distribution in a descending slab 10 Ma after initiation of subduction is shown in figure 3.16. Note that isotherms are bent downward and that phase changes occur at shallower depths in the slab than in surrounding mantle, in response to cooler temperatures in the slab. The calculated heat flow distribution shows a minimum over the trench, as shown also by measured heat flows. This minimum results chiefly from conductive heat transport from surrounding mantle into the slab at shallow depths.

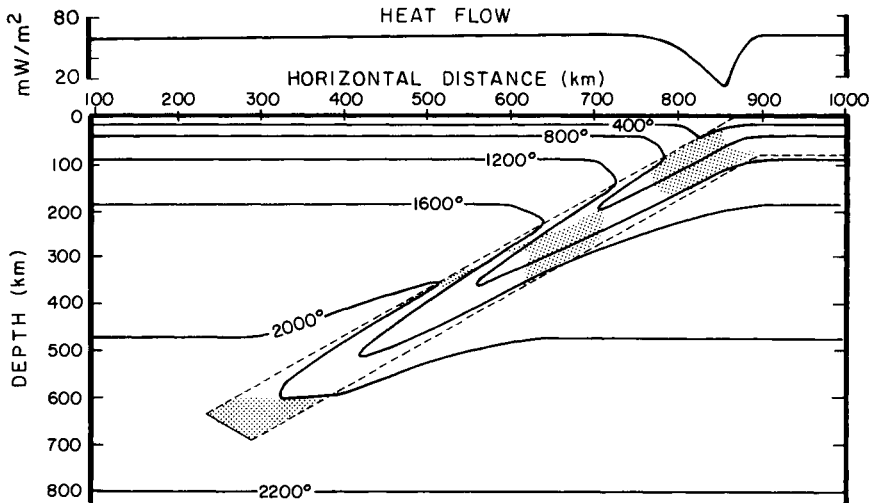


FIG 3.16 Temperature distribution in a descending lithospheric slab 10 Ma after initiation of subduction for a spreading rate of 8 cm/a. Shaded zones indicate the gabbro-eclogite, olivine- β -phase, and the 670-km phase changes, in order of increasing depth. Modified after Toksoz et al. (1971).

Forces Driving Plates

Although plate motions are clearly related to thermal convection in the mantle, a quantitative model relating the two processes remains elusive. The shapes and sizes of plates and their velocities exhibit large variations, and do not show simple geometric relationships to convective flow patterns. Models indicate that plates move in response to ridge-push and slab-pull forces and that stresses transmitted from the asthenosphere to the lithosphere are very small (less than a few tens of bars). In effect, stress distributions are consistent with the idea that at least oceanic plates are decoupled from underlying asthenosphere (Wiens and Stein, 1985). Ridge-push forces are caused by two factors: (1) horizontal density contrasts resulting from cooling and thickening of the oceanic lithosphere as it moves away from ridges, and (2) the elevation of the ocean ridge above the surrounding seafloor (Richardson and Cox 1984; Spence, 1987). The slab-pull forces in subduction zones reflect the cooling and negative buoyancy of the oceanic lithosphere as it ages. The gabbro-eclogite and olivine- β -phase transitions that occur in descending slabs also contribute to slab-pull by increasing the density of the slab.

The relationship between average plate velocity and percentage of plate edge that is a subduction zone suggests that the slab-pull effect dominates in driving plates (fig. 3.17). Plates surrounded by <5% of subduction zones have velocities <5 cm/a, while those with $\geq 20\%$ of their margins subducting have velocities >5 cm/a. Calculations of stress distributions in plates also indicate that slab-pull is the major force moving plates. Although slab-pull forces are more important in driving plates than ridge-push forces, the difference in the magnitude of these forces is not well established. Estimates suggest that slab-pull forces exceed ridge-push forces by a factor of at least two, and perhaps as high as 10 (Spence, 1987). Although slab-pull cannot initiate subduction, once a slab begins to sink the slab-pull force rapidly becomes the dominant force for continued subduction.

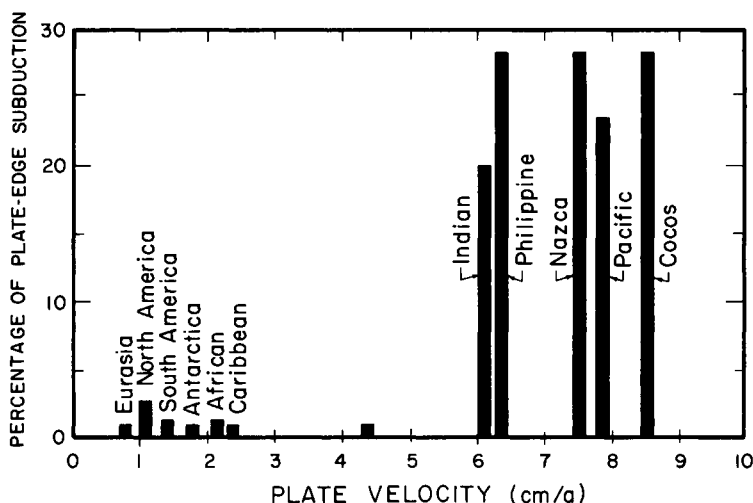


FIG 3.17 Mean plate velocity as a function of percentage of subduction around plate edge. After Forsythe and Uyeda (1975).

Convection in the Mantle

It is generally agreed that convection is responsible for driving plate tectonics. Convection arises because of buoyancy differences, with lighter material rising and denser material sinking. In terms of quantitative laboratory models, *Rayleigh–Bernard convection* is perhaps best understood. This type of convection arises because of heating at the base and cooling at the surface. Lord Rayleigh was the first to show that convective behavior of a substance is dependent on a dimensionless number, now known as the *Rayleigh number*. For a simple homogeneous liquid to convect as it is heated at the base, the Rayleigh number must exceed 2000, and for the convection to be vigorous it must be the order of 10^5 . Irregular turbulent convection begins when the Rayleigh number reaches about 10^6 , and such convection probably exists in the Earth where the Rayleigh number is $>10^6$. Another factor contributing to mantle convection is lateral motion of subducted slabs. Absolute plate motions indicate that slab migration is generally opposite to direction of subduction at rates of 10–25 mm/a. As a result, the downward motions of slabs are generally steeper than their dips (Garfunkel et al., 1986). Calculations indicate that the mass flux in the upper mantle caused by lateral slab motions may be an important contribution to large-scale mantle convection. Migrating slabs do not appear to separate mantle convection cells. An important consequence of lateral slab migration is mixing in the upper mantle.

A voluminous literature exists on models for convection in the Earth. In general, models fall into two categories: (1) layered convection models and (2) whole-mantle convection models. In the *layered convection models*, convection occurs in two or more isolated zones within the mantle, while *whole-mantle convection models* involve the entire mantle. Of the layered models, only the two-layer model, which calls upon separate convection above and below the 670-km discontinuity, seems possible in terms of available data. Whether the upper and lower mantle convect separately

depends on the nature of the transition zone at 670 km — a change in chemical composition would seem to necessitate separate convection. Evidence cited to support two-layer convection is largely geochemical and isotopic. The depleted reservoir for ocean-ridge basalts (see Chapt. 7), which must have been isolated from other reservoirs for at least 2 Ga, is assigned to the mantle above 670 km and the lower mantle to either undepleted (primordial) or mixed geochemical reservoirs. However, the geochemical arguments are not definitive in terms of the locations of mantle reservoirs and several lines of evidence seem to support whole-mantle convection models (Davies, 1984; Loper, 1985).

Of the difficulties recognized for two-layer convection models, the following are most significant.

1. Basalts derived from both layers occur at the Earth's surface. If the mantle were statically layered, uncontaminated material from the lower mantle should not serve as a magma source.
2. Two-layer models involve lower mantle that is too hot and upper mantle that is too cool. Thus, heat from the lower mantle must be convected through a thermal boundary layer into the upper mantle and this, in turn, requires a viscosity contrast at this boundary that is not observed. In fact, recent data suggest the lower mantle is more, rather than less, viscous than the upper mantle (Gurnis and Davies, 1986a).
3. The aspect ratio (width/depth) of convection cells should be close to unity. This is not consistent with layered convection in which convection cells bottom out at 670 km, yet horizontal measurements probably exceed 10^4 km as reflected by plate sizes.
4. Sr and Pb isotopic data from oceanic volcanics require not just two, but many ancient mantle sources (see Chapt. 7).
5. Seismic-velocity studies indicate that descending slabs of lithosphere sink into the lower mantle. This would seem to lead to mixing of the upper and lower mantle.
6. Seismic tomographic and geoid studies indicate the existence of large heterogeneities in the mantle which cross the 670-km discontinuity. Furthermore, they suggest that hot upwellings of mantle extend from the core to the lithosphere.
7. Subduction zone dips can be accommodated in whole-mantle convection models but not in layered-mantle convection models. Although whole-mantle convection seems necessary in terms of the above constraints, it is not yet clear how geochemical heterogeneities are preserved. Recent theoretical models suggest that it is possible to preserve heterogeneities of a variety of sizes in times of 1–2 Ga (Gurnis and Davies, 1986b), a subject which will be returned to in Chapter 10.

The Core

Introduction

Seismologic data indicate that the radius of the core is 3485 ± 3 km and that the outer core does not transmit seismic waves. This latter observation is interpreted to mean that the outer core is in a liquid state (Lilley, 1979). The inner core, with a

radius of 1220–1230 km, transmits S-waves at very low velocities, suggesting that it is a solid near the melting point or partly molten. Detailed analysis of travel-times of seismic waves reflected from and transmitted through the core indicate that the outer liquid core is relatively homogeneous and well mixed, probably by convection currents. The relief on the core–mantle interface is limited to about 5 km. The viscosity of the outer core is poorly known with an estimated lower limit of about 10^{-5} m²/sec and an upper limit of 0.5 m²/sec. Measurements of the melting point of iron, the primary constituent in the core, at very high pressures (up to 2.5 Mbar) indicate that at the core–mantle boundary (~ 1.4 Mbar) the melting point is about 4500°C and at the inner–outer core boundary (~ 3.3 Mbar) it is about 7300°C. Corrected for melting point depression of impurities (probably chiefly sulfur), these temperatures are reduced to 3500°C and 6300°C, respectively (Williams et al., 1987). These data constrain the maximum temperature at the Earth's center to about 6600°C.

Most evidence indicates that the Earth's magnetic field is generated in the core. To a first approximation, the Earth's magnetic field is similar to that produced by a bar magnet at the center of the Earth. It deviates from such a simple model, however, in that it contains a small non-dipole component characterized by random highs and lows in the field and it changes in position and strength with time. The field also can reverse its polarity and such reversals are recorded in the geologic record by rock magnetization (see Chapt. 5). The Earth's magnetic field is generally thought to be generated in the outer core by dynamo action of convecting currents. Changes in position, polarity, and strength of the magnetic field with time are consistent with such an origin. On short time scales (hundreds of years), static features in the magnetic field and absence of a westward drift suggest that the core may be coupled to the mantle (Bloxham and Gubbins, 1987). Flow patterns in the outer core that are responsible for the geodynamo are poorly understood. The near coincidence of the Earth's magnetic and rotational poles suggests that the daily rotation of the core strongly influences fluid motions. Proposed flows range from oscillations of a roughly stratified fluid to turbulent convection. There appears to be a tremendous number of interconnecting current loops in the outer core which have different shapes and can grow, fade and reverse polarity on a variety of time scales.

Generation of the Earth's magnetic field requires a large input of energy (10^{10} – 10^{11} Watts). Both radioactive decay and latent heat from solidification of the inner core have been suggested as the required energy sources. The only long-lived radioactive isotope capable of providing this energy is ⁴⁰K. Thermodynamics and experimental data suggest that, at very high pressures, potassium may follow sulfur and thus be incorporated in the outer core. However, most of the potassium in the Earth appears to reside in the continents and the upper limit of the amount that may be in the core does not provide enough energy to produce the magnetic field. An alternate and preferred energy source is the latent heat released by crystallization of Fe–Ni on the surface of the inner core. For a cooling rate of 12°C per 10^9 a, 10^{12} Watts of energy are released, more than enough to generate the magnetic field. Just how this heat generates convective motions in the outer core is not well understood. It may be achieved by gravitational convection (Loper, 1978). In this model the material crystallizing on the surface of the inner core contains more nickel than the average outer core. Thus, a nickel-depleted low-density layer is developed adjacent to the

TABLE 3.5 *Estimates of the Composition of the Earth's Outer Core*

	1	2	3
Fe	85.5	80	85
Ni	7.5	5	5
S	6.0	15	—
O	—	—	10

1. Iron meteorite composition (6% sulfur).
2. Iron meteorite composition (15% sulfur).
3. FeO model (Ringwood, 1979).

inner core. This layer is gravitationally unstable and will rise, producing convection in the outer core.

Composition of the Core

Shock-wave experimental results limit the major constituents in both the outer and inner core to a mixture of such elements as Fe, Ni, V, and Co. Cosmochemical and nucleosynthesis considerations show that, of these elements, only Fe is sufficiently abundant in the solar system to form a major part of the core. By analogy with meteorites it would appear that an Fe–Ni mixture, dominated by Fe, comprises most of the Earth's core (table 3.5). Refinement of seismic velocity and shock-wave data, however, indicates that the outer core is 8–15% less dense than an Fe–Ni mixture, and requires the presence of one or more elements with low atomic weight. The inner core, which is near its melting point or partially molten, does not contain light elements and may contain as much as 50% Ni. Shock-wave data indicate that the outer core must contain 5–15% of one or more of the elements O, Si, or S (Brett, 1976). Experimental data are consistent with either sulfur or oxygen being the principal light element in the outer core.

The case for sulfur being the most important diluent in the outer core has gained support in recent years. The geochemical behavior of sulfur during the non-oxidizing environment that must have existed during core formation suggests that it should have a strong affinity for iron. Under the same conditions, silicon should enter silicates and remain in the mantle. Also, the fact that iron meteorites contain sulfides (as troilite), yet most do not contain silicates or oxides, supports the existence of sulfur in planetary cores. Experimental data indicate, however, that FeO should also be soluble in iron at high temperatures (Ringwood, 1979).

Origin of the Evolution of the Core

The origin of the core can be considered in terms of two questions:

1. By what mechanism did Fe–Ni separate from the mantle and move to the center of the Earth?
2. When did the core form and is it still growing in size?

Two models for core formation have been proposed. The inhomogeneous accretion model presents many geochemical problems (see Chapt. 2) and most data favor a

model in which iron melts in the early Earth and gravitationally sinks to the center. The amount of gravitational energy released by such sinking is enough to heat the Earth by about 2000°C and, if retained, to completely melt the Earth. The fact that gaseous elements are retained in the mantle does not favor complete melting of the Earth, and thus much of this early gravitational energy must have been lost (Chapt. 10). Several lines of geochemical evidence indicate that the core formed no later than 50 Ma after planetary accretion and perhaps during late stages of accretion. An increase in the U/Pb ratio in the Earth within 100–200 Ma of accretion as implied by Pb isotope distributions may record core formation (Vollmer, 1977). The distribution and ratios of siderophile elements such as Mo, W, Co, and Ni in the mantle suggests that core formation was essentially complete after 85–95% of the Earth had accreted (Newsom and Palme, 1984). The high contents of these elements in the mantle indicate they must have accreted to the Earth largely after core formation, since core formation would have depleted these elements below their observed concentrations in the mantle.

Several models have been proposed for the sinking of Fe–Ni, or more likely an Fe–Ni–S mixture, to the Earth's center (Stevenson, 1981). One model involves the sinking of large blobs of molten Fe–Ni which grow and migrate downwards through less dense mantle. Another involves the downward migration of a spherical molten layer of Fe–Ni. Silicates, which have higher melting points than the Fe–Ni (+sulfur) layer would float upwards through the layer as it sinks. Both of these models require slow thermal diffusion and may not be feasible in the short period of time during which the core formed (<50 Ma). Another model involves asymmetry in the molten iron layer which results in large stresses that weaken and fracture the underlying silicates. Molten iron can then catastrophically migrate down the fractures, displacing the silicates.

Another line of evidence that favors the presence of sulfur or/and oxygen in the outer core is the fact that core formation seems to require such diluents. However, because pure iron and mantle silicates have similar melting points to great depths, it is difficult to melt iron without extensively melting the mantle, which seems unlikely. The addition of either sulfur or oxygen to iron lowers the melting point substantially because these systems have eutectics, as illustrated for sulfur in figure 3.18. In the present outer core, the geotherm is well above the Fe–FeS eutectic, whereas it is below the Fe melting curve for the inner core. Note that, as the outer core cools, iron will crystallize and sink to the surface of the inner core. This indicates that as the Earth cools the inner core should grow outward at the expense of the outer core until the core is completely solidified.

The beginning of core formation is illustrated diagrammatically in terms of thermal and melting constraints in figure 3.18. Steepening geotherms during accretion intersect the Fe–FeS eutectic in the upper mantle at point A, and a thin discontinuous layer of molten Fe–FeS begins to form at this depth. Continued heating of the Earth causes geotherms to rise above the eutectic melting curve (points B and C), producing a widespread spherically distributed molten layer of Fe–FeS which, because of gravitational instability, begins to sink to the center of Earth. Continued rising temperatures 'clean out' the entire mantle of reduced Fe (and Ni) and much of its sulfur as the core continues to grow. Most or all of core growth occurs during the late stages of planetary accretion.

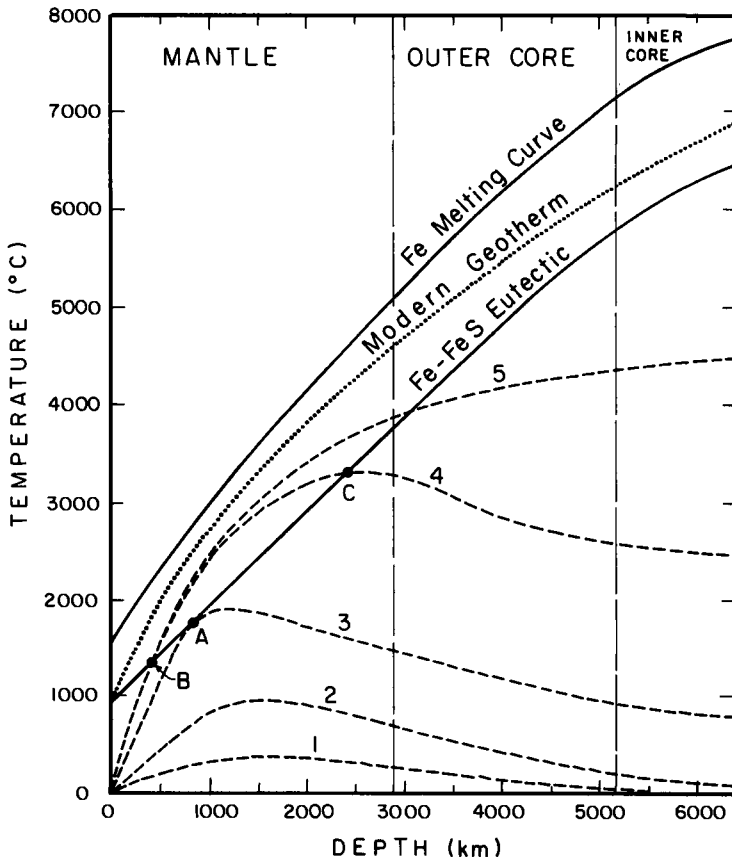


FIG 3.18 Variation in melting points of iron and an Fe-FeS mixture with a eutectic. Temperature distributions based on high-pressure measurements (Williams et al., 1987). Also shown are progressively steepening geotherms (1-5) which intersect the Fe-FeS eutectic melting curve in the mantle.

Summary Statements

1. The Moho occurs from 10 to 12 km beneath oceans and generally from 30 to 50 km beneath continents. Mantle seismic-wave velocities beneath the Moho vary laterally with crustal type.
2. Major seismic discontinuities are recognized in the mantle at 400 and 670 km and minor ones at 550 and 1050 km. In addition, a seismic low-velocity zone (the LVZ) occurs between 50 and 300 km depth except beneath Precambrian shields.
3. Seismic travel-time studies indicate that subducted lithospheric slabs extend at least several hundred kilometers beneath the 670-km discontinuity.
4. Seismic velocity gradients decrease in the D'' region at the base of the mantle. This region, which is likely stratified, differs from overlying mantle by either a phase change or a compositional change.
5. Seismic tomography results indicate the presence of large heterogeneities in the mantle, many of which extend across the 670-km discontinuity and some may extend to the core-mantle boundary. Both the Earth's surface and the

core–mantle interface are upwarped in regions of low seismic velocities in the mantle.

6. Variations in electrical conductivity in the mantle are related to one or a combination of partial melting, increases in water content, changes in composition, or preferential alignment of minerals.
7. Large-scale gravity anomalies reflect lithosphere–asthenosphere interactions or mantle convection. Geoid highs appear to reflect rising convection currents.
8. Seismic anisotropy in the oceanic upper mantle (and crust) appears to reflect alignment of olivine and pyroxene crystals caused by asthenospheric flow.
9. Experimental data indicate that the Moho is not caused by a phase change from gabbro to eclogite. The upper mantle is probably composed chiefly of olivine (70%), pyroxenes, and garnet. It is composed of more than 90% of four elements: Mg, O, Fe, and Si.
10. Geochemical and isotopic data indicate the presence of long-lived ($\geq 10^9$ a) mantle reservoirs that are geochemically undepleted, depleted, and enriched on the basis of their magma and fluid interaction history.
11. The LVZ is probably caused chiefly by incipient melting of hydrous ultramafic rock with some contribution from alignment of olivine and pyroxene crystals.
12. The 400-km discontinuity probably reflects a phase change from Mg-rich olivine to Mg-rich β -phase, a more densely packed mineral. β -phase, in turn, inverts to Mg-rich spinel with only a minor density change at about 500–550 km.
13. The 670-km discontinuity may reflect either or both (a) a breakdown of spinel to perovskite plus periclase, or (b) a compositional change involving enrichment in iron and perhaps silicon.
14. As geotherms drop from ocean ridges to ocean basins, the lithosphere thickens to 100–150 km. Isotherms are depressed at subduction zones.
15. Although both ridge-push and slab-pull forces are responsible for driving lithospheric plates, a positive correlation between percentage of subduction along plate edges and plate velocity, and models of stress distribution in plates indicate that slab-pull forces dominate.
16. Of the two popular models for mantle convection (the two-layer and whole-mantle models), only the whole-mantle model accommodates a wide array of geochemical, isotopic, and geophysical constraints.
17. High-pressure experimental data indicate a temperature at the core–mantle boundary of about 3500°C, and a maximum temperature at the Earth's center of about 6600°C.
18. The Earth's magnetic field is generated by a dynamo action caused by convection currents in the outer core. Energy for this convection is probably supplied by latent heat associated with crystallization of iron on the surface of the inner core.
19. The outer core must contain 5–15% of one or a combination of sulfur, oxygen or silicon (least likely) to explain its low seismic velocities and the inner core, in addition to iron, may contain as much as 50% Ni.
20. The core probably formed during the late stages of planetary accretion by melting of an Fe–FeS eutectic mixture in the mantle, which sank to the Earth's center as it formed.

Suggestions for Further Reading

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CHAPTER 4

The Crust

Crustal Types

The crust is that region of the Earth above the Mohorovicic discontinuity. There are three major crustal divisions—oceanic, transitional, and continental—of which only oceanic and continental divisions are of major importance. Typically, *oceanic crust* ranges from 5 to 15 km thick and comprises 59% of the total crust by area. Islands, island-arcs, and continental margins are examples of *transitional crust* that have thicknesses of 15–30 km. *Continental crust* ranges from 30 to 80 km thick and comprises 79% of the total crust by volume. The average heights of the large continents are 100–300 m above sea level, whereas small continents ($<2 \times 10^6 \text{ km}^2$) are largely or completely submerged (such as New Zealand and the Seychelles Plateau). Both mean elevation and thickness of continents increase with their area (Cogley, 1984). The greater mean crustal thickness of large continents may reflect a greater importance of accretionary growth by collision. Most of the small continents appear to represent fragments rifted from the thinner edges of larger continents.

The Earth's crust can be further subdivided into crustal types (Brune, 1969). A *crustal type* is defined as a segment of the crust exhibiting similar geological and geophysical characteristics. The crust can be adequately described in terms of 12 major crustal types, listed in table 4.1 with some of their physical properties. The first two columns of the table, together with figure 4.1, summarize the areal and volume abundances of various crustal types. Column 3 describes tectonic stability and reflects chiefly the prevalence of earthquake activity and recent deformation. Crustal types are categorized into stable (S) (little or no earthquake activity), intermediate or variable stability (I), or unstable (U). When crustal types have more than one stability entry in the table, the first entry is most common. Brief geological descriptions of each major crustal type are given below. Reference should be made to the tectonic map in the pocket (Plate I).

Precambrian Shields

Precambrian shields are stable parts of the continents composed of Precambrian rocks with little or no sediment cover. Rocks comprising shields may range in age from 0.5 to $>3.5 \text{ Ga}$. Metamorphic and plutonic rock types dominate, and temperature–pressure regimes recorded in the exposed rocks suggest burial depths ranging from as shallow as 5 km to as deep as 35 km or more. Shield areas, in general, exhibit very

TABLE 4.1 *Physical Properties of the Earth's Crust*

Major Crustal Types	1	2	3	4	5	6
	Percentage Abundance		Tectonic Stability ¹	Heat Flow (mW/m ²)	Bouguer Anomaly (mgal)	Total Thickness ³ (km)
	Area	Volume				
1. Shield	6	12	S	38	-10 to -30	35
2. Platform	18	35	S,I	49	-10 to -50	41
3. Paleozoic orogenic belt	8	14	I,S	57		43
4. Mesozoic-Cenozoic orogenic belt ²	6	13	U,F	68	-200 to -300	40
5. Continental rift system ²	<1	<1	U	≥80		28
6. Volcanic island ²	≤1	≤1	I,U,S	≥80		
(a) Hawaii					+250	14
(b) Iceland					-30 to +45	12
7. Arc ²	3	3	U	≥75	-50 to +100	22
8. Trench ²	3	2	U	45	-100 to -150	8 (14)
9. Ocean basin	41	11	S	50	+250 to +350	7 (11)
10. Ocean ridge ^{2,5}	10	5	U	>100	+200 to +250	5 (6)
11. Marginal-sea basin ²	4	3	U,I	≥60	+50 to +100	9 (13)
12. Inland-sea basin ²	1	2	I,S	50	0 to +200	22 (25)
Average Continent	41	79		55	-100	36
Average Ocean	59	21		67 (95)	+250	7 (10)

Major Crustal Types	7	8	9	10	11	12
	V _P (km/sec)	V _P Upper Layer (km/sec)	Thickness Upper Layer (km)	V _P Lower Layer (km/sec)	Thickness Lower Layer (km)	Average Crustal V _P (km/sec)
1. Shield	8.1	6.1	16	6.6	19	6.3
2. Platform	8.1	6.1	17	6.8	21	6.5
3. Paleozoic orogenic belt ²	8.1	6.1	16	6.7	23	6.4
4. Mesozoic-Cenozoic orogenic belt ²	≤8.2	6.0	17	6.8	19	6.4
5. Continental rift system ²	≤7.8	6.1	17	6.9	11	6.4
6. Volcanic island ²						
(a) Hawaii	8.2	5.1	6	7.1	6	6.1
(b) Iceland	7.2	5.0	5	6.5	7	6.1
7. Arc ²	≤8.0	5.5	12	6.6	12	6.1
8. Trench ²	8.0	5.0	3	6.6	5	6.4
9. Ocean basin	8.2	5.1	1.5	6.8	5	6.4
10. Ocean ridge ^{2,5}	≤7.5	5.0	1	6.5	4	6.4
11. Marginal-sea basin ²	≤7.9	5.0	2	6.6	5	6.2
12. Inland-sea basin ²	8.0	4.0 ⁶	12 ⁶	6.5	10	6.0
Average Continent	8.0	6.1	17	6.8	21	6.4
Average Ocean	8.1	5.1	1.5	6.6	5	6.4

¹S = stable²Properties vary considerably³Depth to Moho from sea level given in parentheses⁴Excluding sedimentary layer except for inland-sea⁵Excluding Iceland⁶Sedimentary layer

I = intermediate

U = unstable

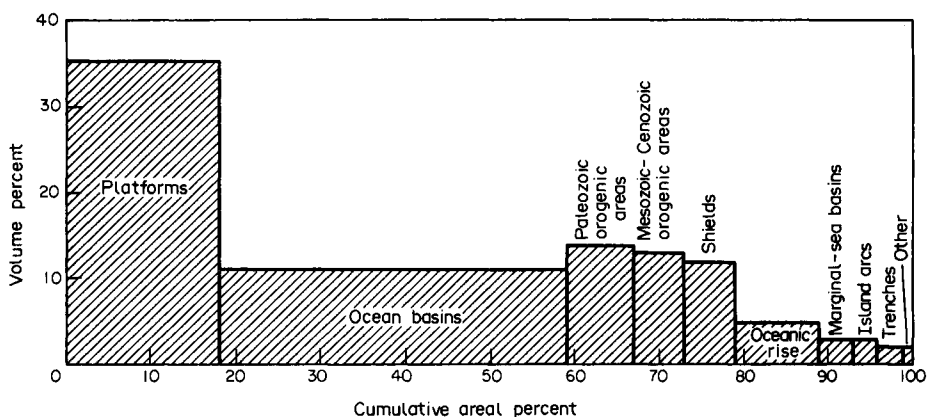


FIG 4.1 Areal and volume distribution of major crustal types.

little relief and have remained tectonically stable for long periods of time. They comprise about 12% of the total crust by volume, with the largest shields occurring in Africa, Canada and Antarctica.

Platforms

Platforms are also stable parts of the crust with little relief. They are composed of Precambrian basement similar to that exposed in shields, and are overlain by 1–3 km of relatively undeformed sedimentary rocks. Shields and the Precambrian basement of platforms are collectively referred to as cratons. A *craton* is generally considered as an isostatically positive portion of the continent that is tectonically stable relative to adjacent mobile belts. Sedimentary rocks on platforms may range in age from Precambrian to Cenozoic and reach thicknesses up to 5 km, as for instance found in the Williston basin in the north-central United States. Platforms comprise most of the crust in terms of volume (35%) and most of the continental crust in terms of both area and volume (fig. 4.1 and Plate I).

Paleozoic Orogenic Belts

Paleozoic orogenic belts are long, curvilinear belts of Paleozoic deformation and igneous activity. They range from several hundred to several thousand kilometers in length (Plate I), and are composed of a variety of rock types. Their surface expression is typically eroded mountain ranges with little relief. They are generally regions of intermediate or stable tectonic activity.

Mesozoic–Cenozoic Orogenic Belts

Mesozoic and Cenozoic orogenic belts are similar in size and geologic character to Paleozoic orogenic belts. They differ from the latter in that they are tectonically unstable. The most extensive Mesozoic belts are the Cordilleran in western North and South America and similar belts in eastern Asia (Plate I). The two most extensive orogenic belts of Cenozoic age are the Cordilleran belt along the western

edge of North and South America and the Alpine-Himalayan belt crossing southern Eurasia. *Plateaus* occur in both Paleozoic and Mesozoic-Cenozoic belts and are characterized by uplifted crustal areas that have escaped the deformation characterizing orogenic belts. Examples are the Colorado and Tibet Plateaus.

Continental Rift Systems

Continental rift systems are fault-bounded valleys ranging in width from 30 to 75 km and in length from a few tens to thousands of kilometers. They are characterized by a tensional tectonic setting in which the rate of extension is less than a few millimeters a year. Magma bodies have been detected by seismic studies beneath some rifts such as the Rio Grande rift in New Mexico. The longest modern rift system is the East African system, which extends over 6500 km from the western part of Asia Minor to southeastern Africa (Plate I). Rift systems range from simple *grabens* (a down-dropped block between two normal faults) or *half-grabens* (a tilted block with one or more normal faults on one side) to a *complex graben system* (a down-dropped or tilted block cut by many normal faults, many or most of which are buried by sediments). The *Basin and Range Province* in western North America is a multiple rift system composed of a complex series of alternating grabens and horsts. *Aulacogens* are rifts that open into ocean basins and die out towards the interior of continents. Many appear to have formed at triple junctions during fragmentation of a continent and represent "failed arms". Rift systems may occur in rocks of any age, and young rifts (≤ 30 Ma) are characterized by unstable tectonic conditions. Some large volcanic plateaus such as the Columbia River Plateau in the northwestern United States and the Deccan Traps in India also appear to be related to extensive tensional faulting.

Volcanic Islands

Volcanic islands occur on or near ocean ridges (e.g. St Paul Rocks and Ascension Island in the Atlantic Ocean) and in ocean basins (such as the Hawaiian Islands). They range in tectonic stability from intermediate or unstable in active volcanic islands to stable (such as Easter Island in the South Pacific). Their size ranges from <1 to about 10^4 km².

Arc Systems

Arc systems occur above active subduction zones and can be subdivided into island-arcs and continental-margin arcs. *Island-arcs* are arcuate chains of volcanic islands formed on oceanic crust such as the Mariana, Kermadec and Lesser Antilles arcs. *Continental-margin arcs* are produced above subduction zones that plunge beneath continental margins, such as the Japan and Andean arcs. Some arcs, such as the Aleutian Islands, continue from continental margins into oceanic crust. Arcs are composed dominantly of young volcanic and plutonic rocks and derivative sediments. Modern arcs are characterized by intense earthquake activity and volcanism, and by variable physical properties.

Trenches

Oceanic *trenches* mark the beginning of subduction zones and are associated with intense earthquake activity. Trenches parallel arc systems and range in depth from 5 to 8 km, representing the deepest parts of the oceans. They contain relatively small amounts of sediment derived chiefly from nearby arcs or continental areas.

Ocean Basins

In terms of area, ocean basins comprise most of the Earth's surface (fig. 4.1). They are tectonically stable and characterized by a thin, deep-sea sediment cover (approximately 0.3 km thick) and linear magnetic anomalies. The sediment layer thickens along continental slopes where detrital sediments are supplied from the continents. Although ocean basins are rather flat, they contain abyssal hills, seamounts, guyots, and oceanic islands. *Abyssal hills*, which are the most abundant topographic feature on the ocean floor, are sediment-covered hills ranging in width from 1 to 10 km and in relief from 500 to 1000 m. *Seamounts* are submarine volcanoes (both active and extinct), and *guyots* are flat-topped submarine volcanoes produced by erosion at sea level followed by submersion, probably due to sinking of the ocean floor. Coral reefs grow on some guyots as they sink, producing *atolls*.

Ocean Ridges

Ocean ridges are widespread linear rift systems in oceanic crust (Plate I), that are topographic highs and tectonically unstable. A medial rift valley generally occurs near their crests in which new oceanic lithosphere is produced by intrusion and extrusion of basaltic magmas as seafloor spreading moves the lithosphere away from the ridge axis. The relief in medial rift zones varies from rough, mountainous terranes such as that characterizing the Mid-Atlantic ridge, to more subdued relief, such as that characterizing the East Pacific rise. In general, rough topography occurs on slow-spreading ridges, while smooth topography is more common on fast-spreading ridges. Fast-spreading ridges also commonly lack well-developed axial rifts. Iceland is the only known example of a surface exposure of a modern ocean ridge system. The worldwide ocean ridge system is interconnected from ocean to ocean and is more than 80,000 km long. Ridge crests are cut by numerous transform faults (see Chapt. 5), which may offset ridge segments by thousands of kilometers.

Marginal-sea Basins

Marginal-sea basins are segments of oceanic crust between island-arcs (as the Philippine Sea) or between island-arcs and continents (as the Japan Sea and Sea of Okhotsk). They are most abundant in the western Pacific (fig. 4.2), and are characterized by a horst-graben topography (similar to the Basin and Range Province) with major faults subparalleling the adjoining arc systems. The thickness of sediment cover is variable, and sediments appear to be derived chiefly from continental or arc areas. Marginal-sea basins are classified into tectonically *active* and *inactive* types (fig. 4.2). Active basins have thin sedimentary cover, rugged horst-graben topography and high heat flow. Examples of active basins are the Lau-Havre

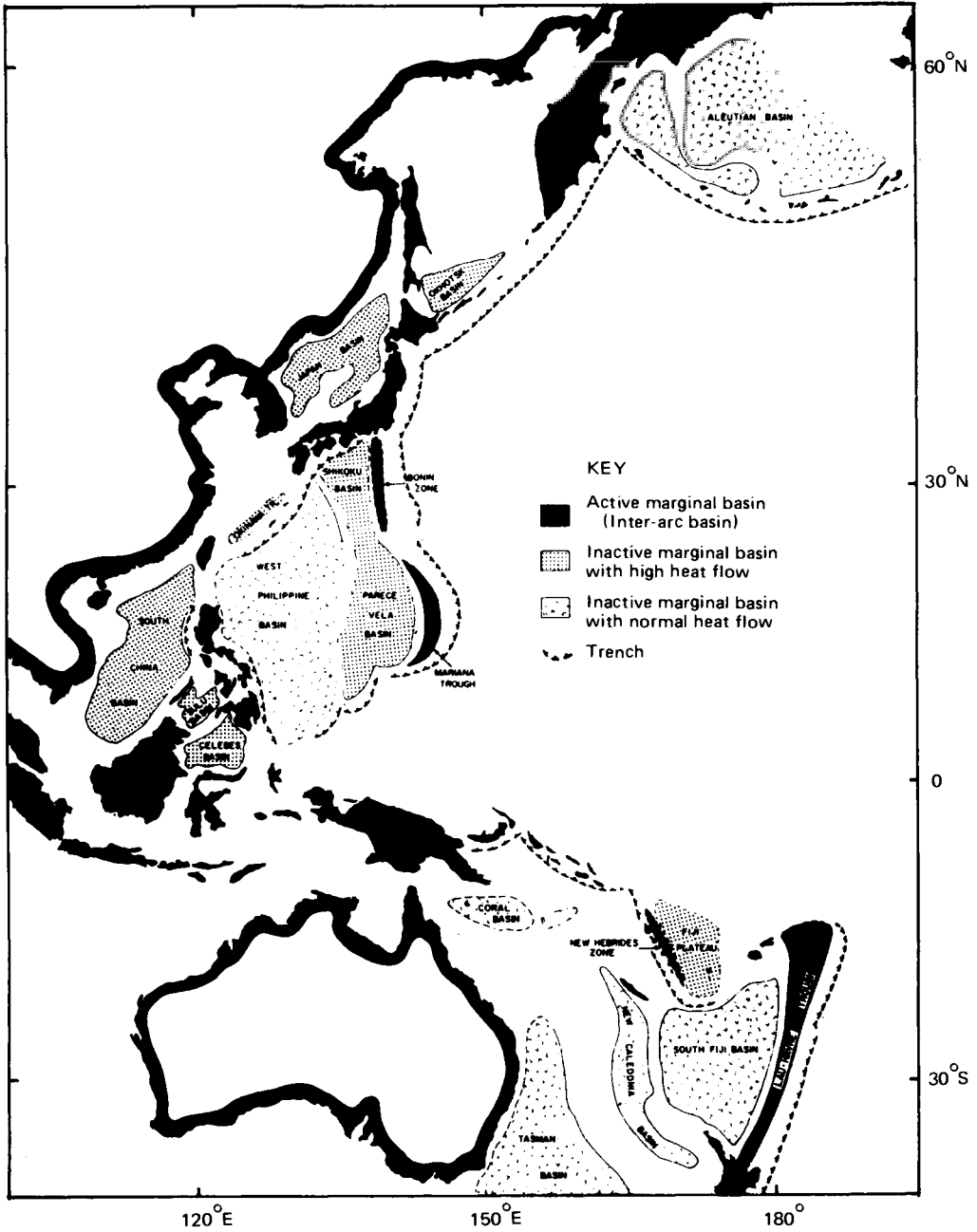


FIG 4.2 Marginal-sea basins in the western Pacific. From Karig (1971), copyright by the American Geophysical Union.

and Mariana troughs and typical inactive basins are the Tasman and West Philippine basins.

Inland-sea Basins

Inland-sea basins are partially to completely surrounded by tectonically stable continental crust. Examples include the Caspian and Black Seas and the Gulf of Mexico. Earthquake activity is negligible or absent. Inland-sea basins contain thick successions (10–20 km) of clastic sediments and both mud and salt diapirs are common.

Seismic Features

General Features

The structure of the crust is determined from body-wave refraction studies, surface-wave dispersion studies and reflection profiles. Crustal models based on seismic data indicate that oceanic crust can be broadly divided into three layers which are, in order of increasing depth: the *sediment layer* (0–1 km thick), the *basement layer* (0.7–2.0 km thick), and the *oceanic layer* (3–7 km thick). Models for the continental crust show a greater range in both number and thicknesses of layers. Although two- or three-layer models for continental crust are most common, one-layer models and models with more than three layers are proposed in some regions. Generally, the continental crust can be considered to be composed of a *sediment layer* (0–5 km thick) underlain by an *upper layer* (10–20 km thick) and a *lower layer* (15–25 km thick), respectively. The boundary between the upper and lower layers in the continental crust has been referred to as the *Conrad discontinuity*. This discontinuity is poorly defined in some regions and is gradational over several kilometers in other regions. In general, the Conrad discontinuity does not appear in seismic reflection data.

P-wave velocities in crustal sediment layers range from 2 to 4 km/sec, depending on such factors as degree of compaction, water content and lithologic types. P-wave velocities in the upper oceanic crust are about 5 km/sec, while those in the upper continental crust are about 6 km/sec (table 4.1). Lower crustal layers in both oceans and continents are characterized by P-wave velocities of 6.5 to 6.9 km/sec. The average crustal P-wave velocity (excluding the contribution of the sediment layer) in both oceanic and continental areas is 6.4 km/sec. With exception of continental borderlands and island-arcs, the average thickness of continents is between 35 and 40 km, while the average thickness of the oceanic crust is only 5 to 7 km. Mean P_n velocities roughly increase, both with mean crustal V_p and depth to Moho in continental areas (Woollard, 1968).

A generalized contour map of the thickness of North America is given in figure 4.3. The thickness distribution varies with crustal type. The central and eastern United States are characterized by rather thick crust (40–50 km), the Canadian Shield by crust of average thickness (30–40 km), and the western part of the continent by crust of variable thickness (25–50 km). The crust thickens beneath Lake Superior, along the Grenville Front in Quebec, beneath the Sierra Nevada Range in California, beneath eastern Montana and beneath the southern Appalachian Mountains. Thin crust (<30 km) occurs in continental borderlands and beneath the Basin and Range Province. Crust of variable thickness, variable P_n velocities (fig. 3.1), and positive upper-mantle travel-time anomalies (fig. 3.4) characterizes Mesozoic–Cenozoic orogenic areas, while moderate to thick crust,

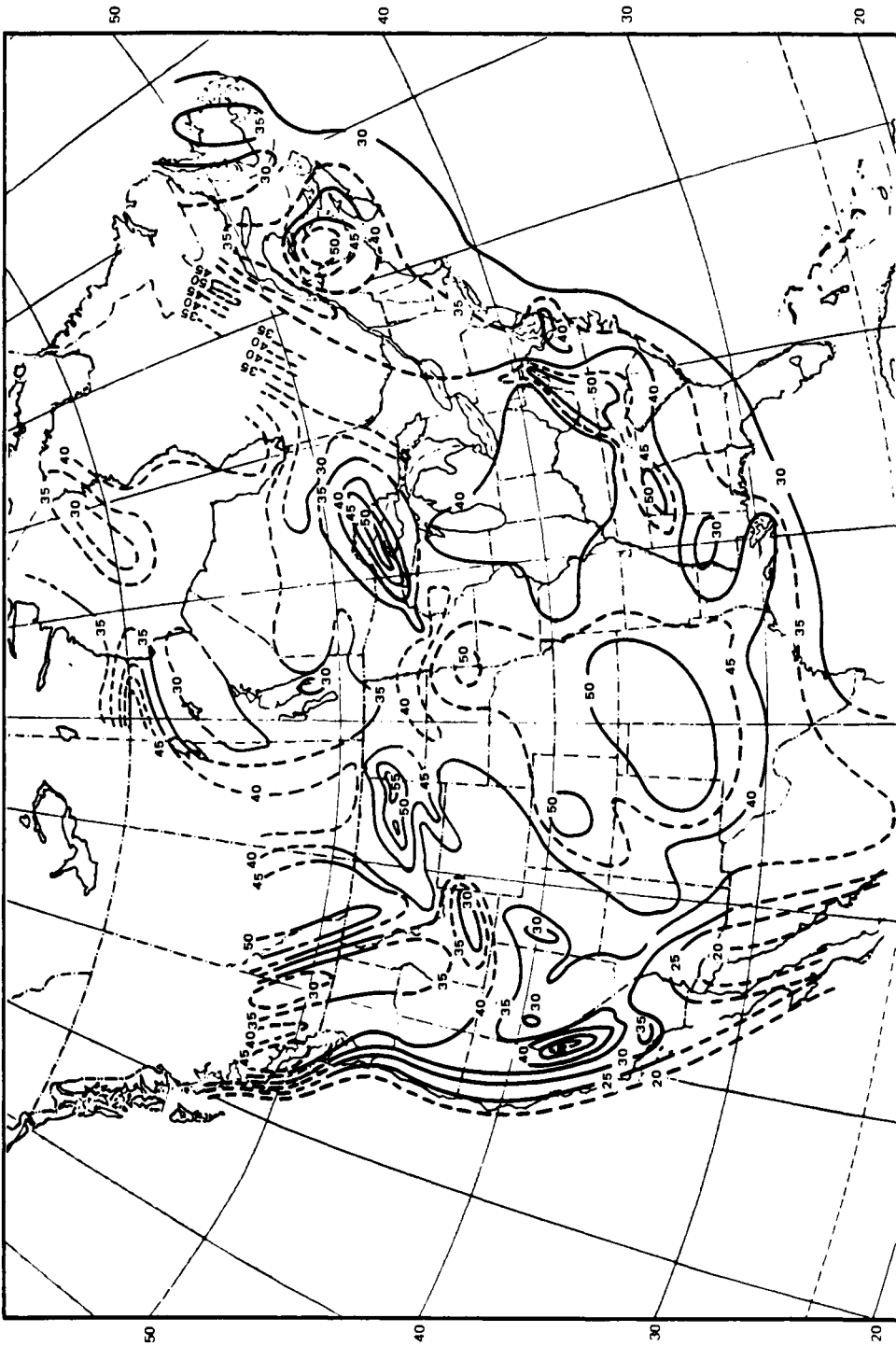


FIG 4.3 Contour map of the thickness (in km) of the North American crust (data chiefly from McConnell et al., 1966; Goodwin et al., 1972; Warren and Healy, 1973).

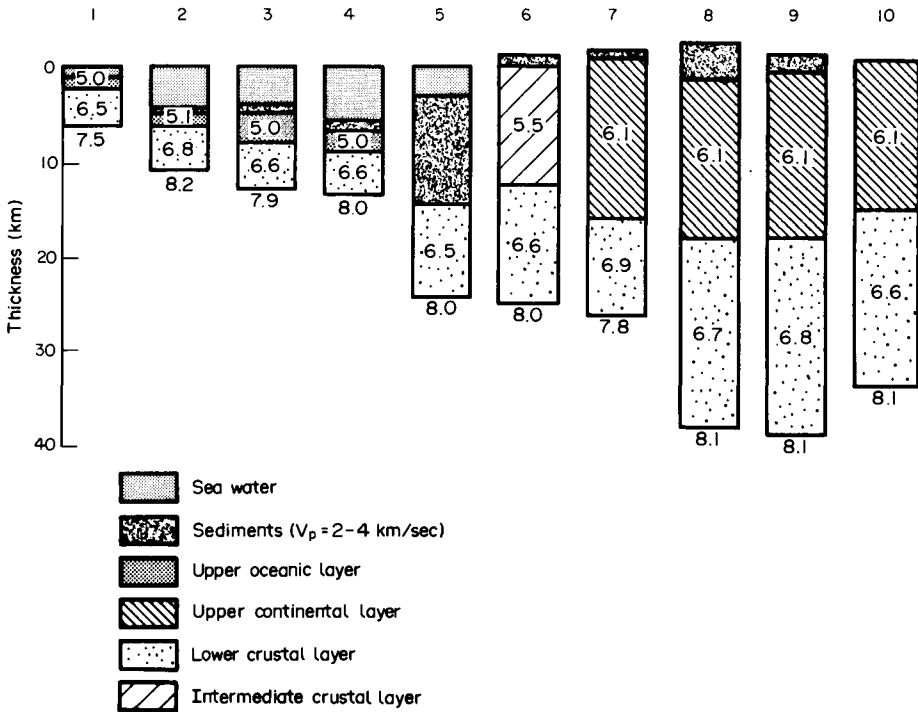


FIG 4.4 Seismic refraction sections of various crustal types. P-wave velocities in km/sec. Key: (1) ocean ridge, (2) ocean basin, (3) marginal-sea basin, (4) trench, (5) inland-sea basin, (6) arc, (7) continental rift, (8) Phanerozoic orogenic area, (9) platform, and (10) Precambrian shield.

average P_n velocities (8.1–8.2 km/sec), and negative upper-mantle travel-time anomalies characterizes pre-Mesozoic crustal types.

Crustal Types

Compilation of Data

The average seismic structure and velocity distribution of various crustal types are compiled in table 4.1 and illustrated diagrammatically in figure 4.4. Results represent median values of total crustal thicknesses, layer thicknesses, and P-wave velocities as compiled from many published seismic refraction sections. The majority of the models used are for a two-layer crust (excluding the sediment layer). In those few cases where more-than-two-layer models are presented, layers are combined (if velocities are close) or the models are rejected (only three out of more than 150 sections were rejected on this basis).

Precambrian Shields and Platforms

Shields and platforms have similar upper- and lower-layer thicknesses and velocities. The difference in their median thickness (column 6, table 4.1) reflects primarily the presence of the sediment layer in the platforms. Upper-layer thicknesses range from about 10 to 25 km and lower-layer thicknesses from 16 to 30 km. Velocities in both

layers are rather uniform, generally ranging from 6.0 to 6.2 km/sec in the upper layer and 6.5 to 6.9 km/sec in the lower layer. P_n velocities are typically in the range of 8.1–8.2 km/sec, rarely reaching 8.6 km/sec. Shield areas with high-grade metamorphic rocks dominating at the surface may be characterized by a one-layer crust (with $V_p = 6.5\text{--}6.6$ km/sec) as exemplified, for instance, by the Grenville Province in eastern North America (Taylor et al., 1980). Relief up to 10 km per 100 km distance has been reported on the Moho in some platform and shield areas, and refinements in the interpretation of seismic and gravity data indicate the presence of significant lateral variations in crustal layer and velocity distributions. An example of such lateral variation is given in figure 4.5, which is a seismic cross-section extending from west of the Rocky Mountains in British Columbia into the platform of Alberta and Saskatchewan. Data suggest that this region is composed of juxtaposed blocks with varying velocities and densities. Some of the near-vertical contacts may represent large fault zones that penetrate the entire crust and part of the upper mantle. Also, results suggest the existence of a high-velocity zone (~ 7.2 km/sec) in the lower crust. Similar complex lateral variations have been proposed for platform areas in Eurasia.

Phanerozoic Orogenic Areas

Crustal thickness in young, post-Paleozoic orogenic areas is extremely variable, ranging from about 20 km in island-arc and continental borderland areas to 80 km beneath the Himalayas. The average layer thicknesses and velocities of Phanerozoic orogenic areas are about the same as platforms (table 4.1, fig. 4.4). In areas of very thick crust, such as beneath the Andes (70 km) and Himalayas (80 km), the thickening occurs primarily in the lower crustal layer. P_n velocities in post-Paleozoic orogenic areas range from normal (8.1–8.2 km/sec) to low (<7.8 km/sec). Lower crust high-velocity zones (7.0–7.4 km/sec) are more frequent in post-Paleozoic orogenic areas than in older crust. The velocity contrast between the lower crust and upper mantle is also commonly less beneath young orogenic areas (0.5–1.5 km/sec) than beneath platforms and shields (1–2 km/sec) (Oxburgh, 1969).

Continental Rifts

Continental rift systems are characterized by thin crust (<30 km) and low P_n velocities (<7.8 km/sec). It is of interest in regard to the origin of continental rifts that thinning of the crust in these regions is accomplished by thinning of the lower crustal layer (fig. 4.4), which ranges from only 4 to 14 km thick. Although most earthquake foci in rifts are <20 km deep, some occur as deep as 25–30 km.

Arcs

Resolution of seismic data is poor in arc systems, and considerable uncertainty exists regarding arc crustal structure. Seismic sections of several arc systems are given in figure 4.6 to illustrate the large range in velocity and layer distributions. Crustal thickness ranges from 5 km in the Lesser Antilles to 35 km in Japan, averaging about 22 km, and P_n velocities range from normal (8.0–8.2 km/sec) to low (<7.8 km/sec). All values given in table 4.1 have large standard deviations and, as is

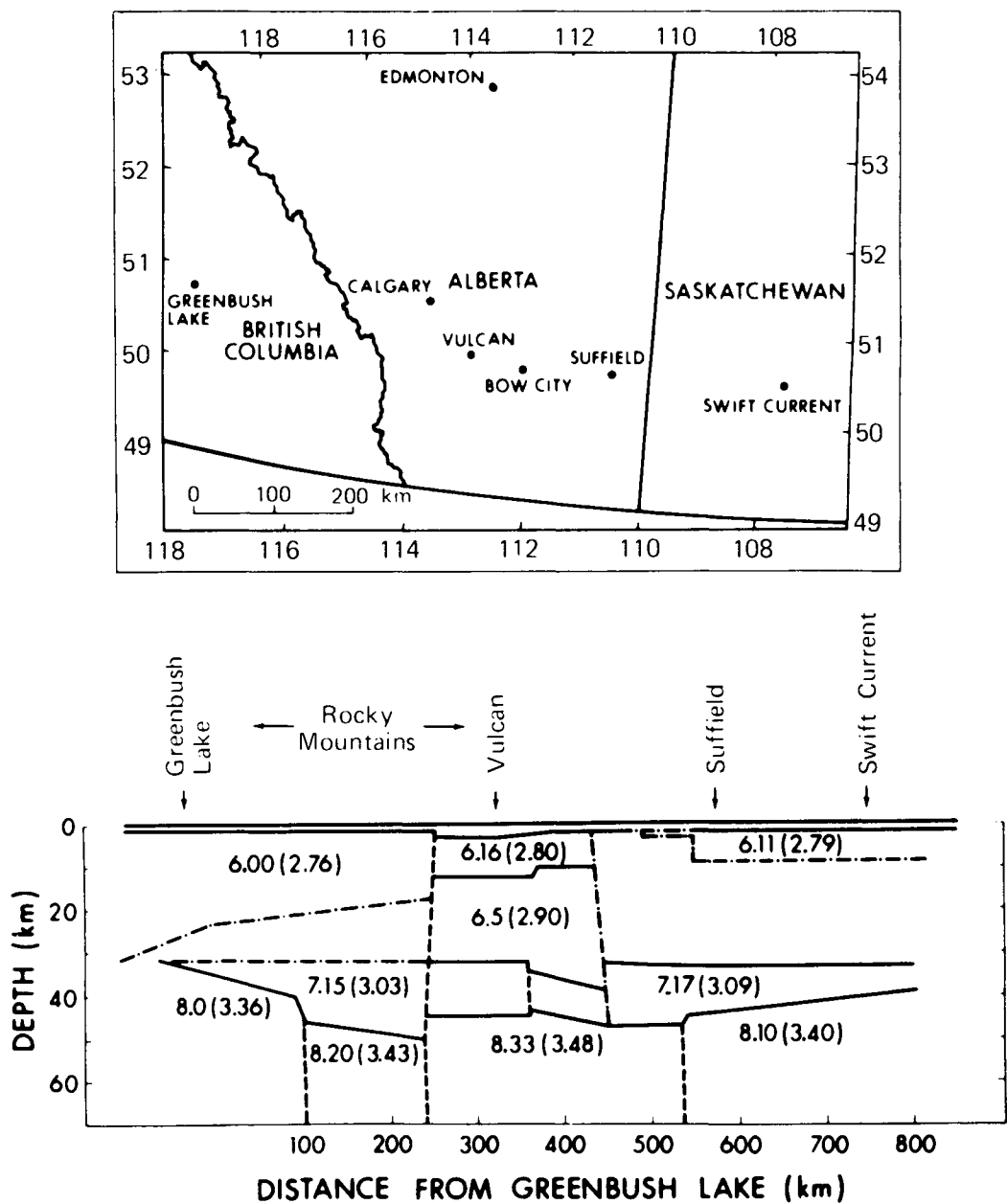


FIG 4.5 Crustal model of southwestern Canada based on seismic and gravity data. After Chandra and Cumming (1972). Number outside the parentheses gives P-wave velocity (km/sec) and that inside, the calculated density (gm/cm³). Reproduced by permission of the National Research Council of Canada.

indicated by the sections in figure 4.6, arc crustal structure cannot be fitted to a simple two-layer crustal model. There is also evidence in some arc systems of an intermediate-velocity layer (5.0–6.0 km/sec) of varying thickness.

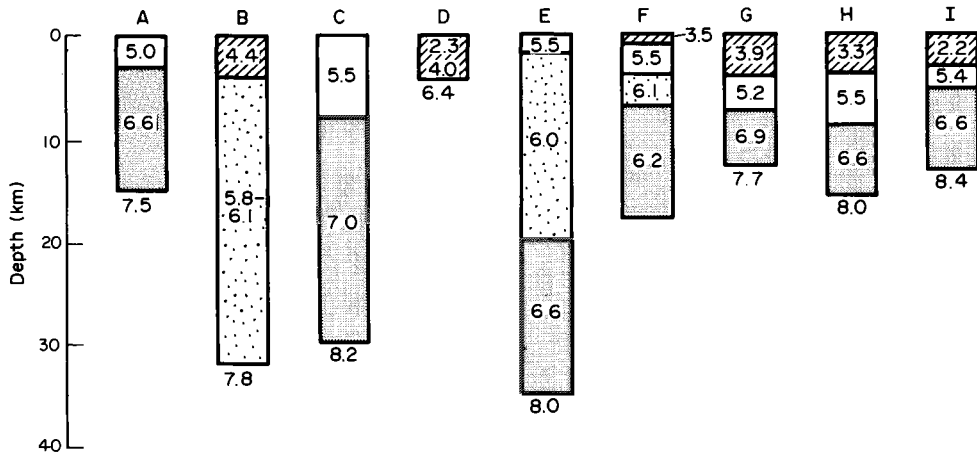


FIG 4.6 Seismic sections of arcs. Key: (A) Southern Kurile Islands (Gainanov et al., 1968); (B) Sakhalin (Gainanov et al., 1968); (C) Puerto Rico (Talwani et al., 1959); (D) Lesser Antilles (Fisher and Hess, 1963); (E) Northern Honshu, Japan (Rikitake et al., 1968); (F) New Zealand, near Wellington (McConnell et al., 1966); (G) Tonga Islands (Shor et al., 1971); (H) Mariana Islands (Murauchi et al., 1968); (I) Fiji Islands (Shor et al., 1971). Average velocities given in km/sec.

Volcanic Islands

Crustal thickness in volcanic islands ranges from 10 to 20 km, with upper crustal velocities ranging from 4.7 to 5.3 km/sec and lower crustal velocities from 6.4 to 7.2 km/sec (Furumoto et al., 1973). V_P ranges from 7.8 to 8.5 km/sec.

Trenches

As with arcs, the crustal structure of trenches is poorly known. Data suggest a wide range in crustal thickness beneath trenches (3–12 km), with most thickening and thinning occurring in the lower layer, which ranges from 4 to 8 km thick. The upper layer in trenches ranges from 2 to 4 km thick and exhibits variable V_P in the range of 3.8–5.5 km/sec (averaging about 5 km/sec). Most trenches have a sediment layer <3 km thick.

Ocean Basins

Crustal structure in ocean basins is rather uniform, not deviating greatly in either velocity or layer thickness distribution from that shown in figure 4.4. Total crustal thickness ranges from 6 to 8 km, and P_n velocities are generally uniform in the range of 8.1–8.2 km/sec. Oceanic crust does not appear to thicken with age (McClain and Atallah, 1986) and the width of the Moho may vary between 0 and 2 km. The sediment layer averages about 0.3 km in thickness and exhibits several strong reflecting zones, some of which are probably produced by cherty layers, as suggested by cores retrieved by the Deep Sea Drilling Project. The most prominent reflecting horizon is the interface between the sediment layer and basement, which is characterized by a rough topography. The thickness of the basement layer, which averages about 1.5 km, decreases toward ocean ridges and, in general, is inversely proportional to the calculated spreading rates for a given ocean basin or segment

thereof. Velocity increases rapidly with depth in the basement layer and significant seismic anisotropy has been described in some areas (Stephen, 1985). Such anisotropy is probably caused by preferred orientation of large fractures in the basement layer that were formed during the early stages of crustal development by near-axis extensional deformation. The oceanic layer is generally rather uniform in both thickness (4–6 km) and velocity (6.7–6.9 km/sec).

Ocean Ridges

Crustal thickness ranges from 3 to 6 km beneath ocean ridges, most of which is accounted for by the lower crust. The sediment and basement layers thin or disappear entirely on most ridges. Unlike other oceanic areas, the velocities in the oceanic layer are quite variable, ranging from 4.4 to 6.9 km/sec. Such a range of velocities is probably produced by seismic anisotropy and by varying but, on the whole, high temperatures beneath ridges. Anomalous mantle ($V_P \leq 7.8$ km/sec) occurs beneath ridge axes. Surface-wave data indicate that the lithosphere increases in thickness from ≤ 35 km beneath ocean ridges to 65 km at a crustal age of 40 Ma.

Marginal-sea Basins

The major difference between crustal structure of marginal-sea basins and ocean basins is the slightly greater total thickness of the former (10–15 km). This thickening results from a thicker sediment layer in marginal seas. Most seismic models show two to four layers above the oceanic layer, which exhibits velocities of 6.6–6.9 km/sec. The upper crustal layers commonly range in velocity from about 2 km/sec at the top (unconsolidated sediments) to about 5 km/sec above the oceanic layer. An intermediate velocity layer (6.0–6.2 km/sec) occurs just above the lower crustal layer in the Caribbean and New Caledonia basins.

Inland-sea Basins

Inland-sea basins show a considerable range in crustal thickness and layer distributions (table 4.1, fig. 4.7). Crustal thickness ranges from about 15 km, as exhibited by the Gulf of Mexico, to 45 km in the Caspian Sea basin. In general, the sedimentary layer or layers ($V_P = 2$ –5 km/sec) rest directly on the lower crust ($V_P = 6.3$ –6.7 km/sec), with little or no upper crustal layer. Differences in crustal thickness among inland-sea basins are accounted for by differences in thickness of both sedimentary and lower crustal layers. Increasing velocities in sediment layers with depth, as for instance shown by the Gulf of Mexico (fig 4.7A), appear to reflect an increasing degree of compaction and diagenesis.

Crustal Low-velocity Zones

Crustal low-velocity zones are reported in some young orogenic provinces and platform areas. Two examples are shown in figure 4.8, one for the Basin and Range Province in northwestern Utah and one for the platform in eastern New Mexico. The latter also includes an electrical conductivity profile. A low-velocity zone is detected at about 10 km in northwestern Utah and another at 20–25 km in both areas. In other parts of the Basin and Range Province, seismic data suggest the

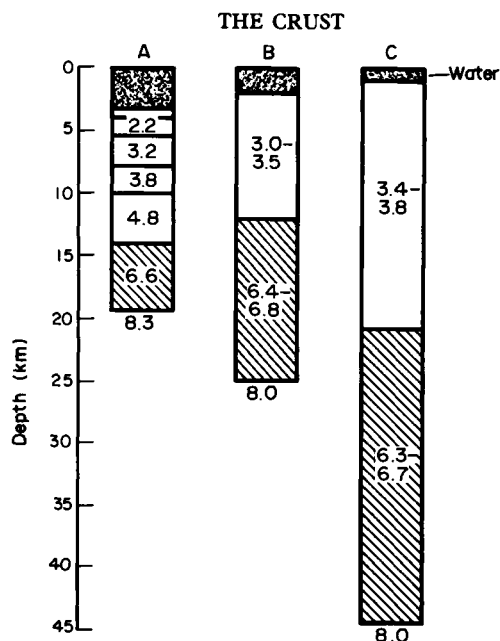


FIG 4.7 Seismic sections of inland-sea basins. After Menard (1967). P-wave velocities in km/sec. Key: (A) Gulf of Mexico; (B) Black Sea; (C) Caspian Sea. Copyright by the American Geophysical Union.

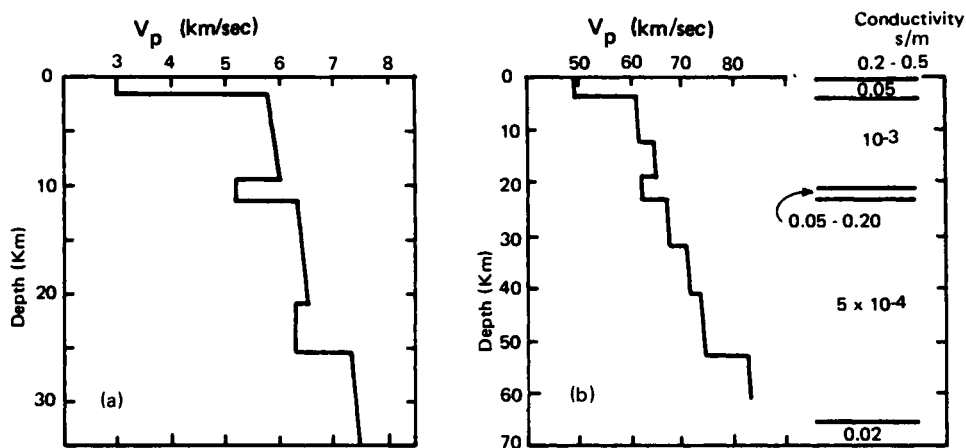


FIG 4.8 Crustal seismic and electrical conductivity profiles. (a) Northwestern Utah. From Landisman et al. (1971), copyright by the American Geophysical Union; (b) eastern New Mexico. After Mitchell and Landisman (1971).

existence of a widespread low-velocity layer (>3.5 km thick) beginning at a depth of 5–6 km. One or two crustal low-velocity layers have also been reported beneath the Rhine graben in Germany and beneath the Alps (Landisman et al., 1971). It is pertinent that aseismic zones occur in the same depth range as that in which crustal low-velocity layers are reported, suggesting the presence of a region that undergoes ductile rather than brittle deformation.

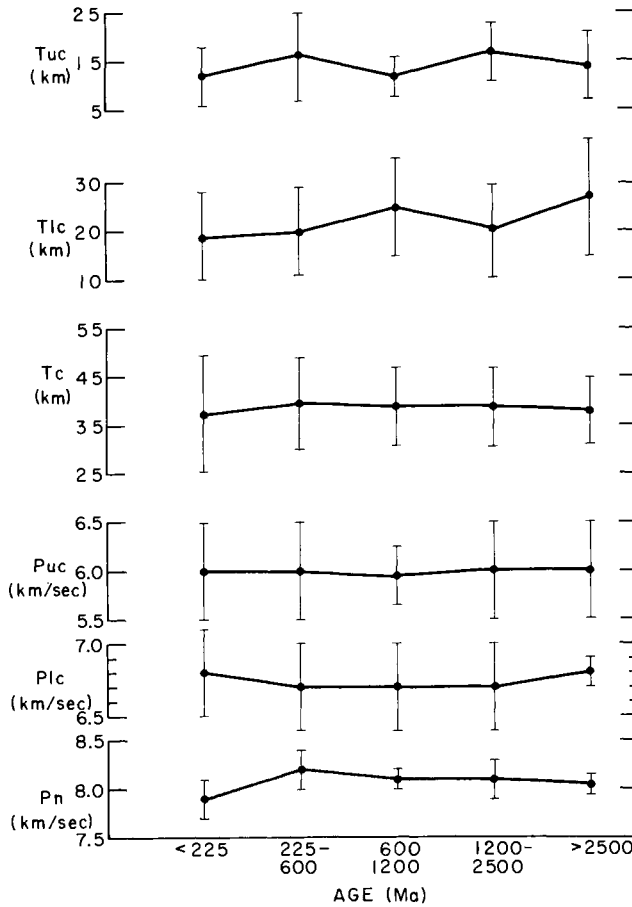


FIG 4.9 Variation of crustal thickness (T_c), layer thickness (T_{uc} = upper layer; T_{lc} = lower layer), crustal P-wave velocity (P_{uc} = upper layer velocity; P_{lc} = lower layer velocity), and P_n velocity with mean crustal age. After Condie (1973). Vertical bars are one standard deviation of the mean.

Crustal Structure and Mean Crustal Age

Seismic velocity and thickness data from the continents are shown as a function of mean crustal age in figure 4.9. Data are grouped into five age categories, and each point represents a mean value. Because crustal thickness is quite variable in Mesozoic–Cenozoic regions, seismic thicknesses are weighted according to areal distribution in these areas. Continental shelves and arcs (other than New Zealand and Japan) are excluded from the compilation, and mean crustal ages ignore the contributions of later Precambrian and Phanerozoic sediments on shields and platforms. Precambrian terranes that have been extensively reworked during younger orogenies are included with the younger crust.

The most striking features of both the thickness and velocity patterns are their approximate constancy with respect to mean crustal age. The results indicate that crustal thickness averages about 38 km regardless of age. Young crust (<225 Ma) ranges from very thick beneath mountain ranges (60–80 km) to thin along

continental margins and in arcs (20–30 km). Craton-forming processes eliminate thick and thin crust and stabilize crust with an equilibrium thickness of 35–40 km.

Results suggest that the thickness of the lower crust (T_{lc}) may increase with age, although the standard deviations are large. Seismic refraction data from Australia seem to support this trend, and suggest that the continental crust has thickened with age by underplating (Drummond and Collins, 1986). The Australia data also clearly indicate that the average P-wave velocity of the continental crust increases with the thickness of the lower crust. The low P_n velocity in the <225 Ma category probably reflects the common occurrence of warmer upper mantle in young orogenic areas. If only those areas <225 Ma in age with normal heat flow ($\leq 50 \text{ mW/m}^2$) are used, P_n velocity is about the same as is observed beneath older crust.

Crustal Reflection Profiles

Continental Areas

Seismic reflection profiles from the continental crust have been extremely helpful in enhancing our knowledge of crustal structure. The most detailed studies are those by the COCORP group in the United States (Brewer and Oliver, 1980; Smithson, 1986) and the BIRPS group in Britain (Matthews, 1986). Although there are still many problems with the interpretation of continental seismic profiles, data allow tentative conclusions regarding crustal structure.

1. Short, discontinuous and arcuate reflections in the middle and lower crust appear to record complexly folded metamorphic rocks.
2. Strong continuous events can be interpreted in terms of major low-angle faults or, at shallow levels, perhaps as layering in sediments and volcanics.
3. Blank or transparent regions in profiles may represent highly folded rocks with dips too steep to return reflected energy or massive granitic plutons.
4. Strong reflections from the lower crust that exhibit low dips ($< 5^\circ$) are common in many reflection profiles.
5. There is no evidence for a single mid-crustal reflection representing the Conrad discontinuity.
6. Reflections from the Moho are of variable intensity and character. Results indicate the boundary ranges from a simple reflector to a completely layered structure, or in some instances, to a nonreflecting or highly deformed boundary.
7. Large faults can be traced to great depth in the crust and some appear to extend into the upper mantle.

An example of a reflection profile in Archean crust that was reactivated during the Cenozoic is shown in fig 4.10. The most striking feature of the profile is a series of dipping reflectors that begin where the Wind River thrust crops out at the surface. These reflectors trace the thrust to a depth of about 24 km, with an average dip angle of 30–45 degrees. This structure suggests that compression was largely responsible for the formation of the Wind River Mountains during the Laramide orogeny. The arcuate reflections (A) shown on the profile are similar to patterns shown on synthetic seismograms for multiple deformed folds, and are interpreted as

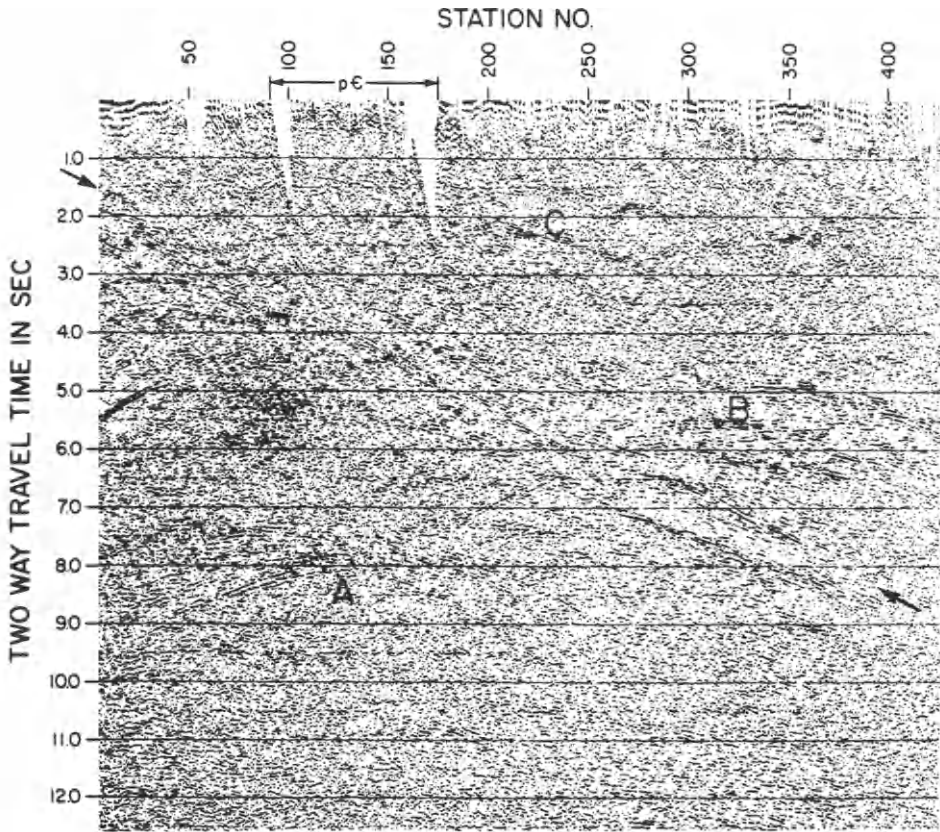


FIG 4.10 COCORP reflection profile across the southern Wind River Range, Wyoming. After Smithson et al. (1980). Arrows mark reflection from Wind River thrust. Deep complex crustal structure marked by arcuate reflections (A) and numerous composite reflections (B). Note: strong shallow reflection at C. Depth (km) is equal to about 3 times the 2-way travel time in seconds.

complexly deformed high-grade metamorphic rocks similar to those exposed in portions of the Wind River Range (Smithson et al., 1980). These results suggest the lower crust in this area may be similar to the deeply eroded Precambrian terranes now exposed at the surface in portions of the continents.

A large number of seismic reflection profiles are now available from a variety of crustal types, chiefly in the United States and Western Europe. Most major tectonic settings on the continents are represented by at least one reflection profile and some have several profiles. From this data base, four types of seismic reflection fabrics are recognized in the continents (fig. 4.11a) Allmendinger et al, 1987):

(1) *Craton*. From the surface to the base of the crust, cratons are characterized by: (a) a generally thin band of layered reflections corresponding to flat-lying Phanerozoic sediments (platforms only); (b) a relatively transparent region in the upper one-third to one-half of the crust; (c) a middle and lower crust dominated by diffractions and dipping reflections of various orientations; and (d) a marked downward decrease in the number of reflections and diffractions at the Moho.

(2) *Zone of foreland deformation*. Foreland deformational zones include thin-

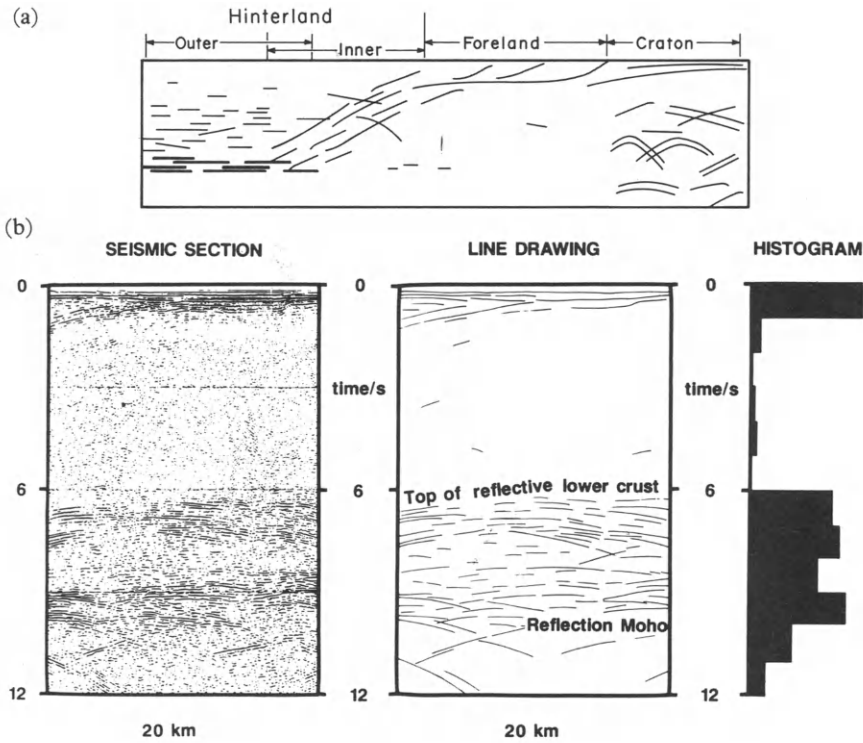


FIG 4.11 (a) Generalized cross-section showing spatial relations of reflection fabrics in a transect across an idealized orogenic belt. Modified after Allmendinger et al. (1987). (b) Segment of a seismic reflection profile from BIRPS southwest of Britain; line-drawing of the section showing the reflective lower crust; and histogram of reflection depths summed over 1 sec travel-time intervals. After Klemperer (1987).

skinned thrusts and decollements and associated folds. Except for complex reflections of sediments in the uppermost crust, few seismic reflectors are found in these zones. The paucity of reflectors at depth may result from poor signal penetration through the complex near-surface structures.

(3) *Inner hinterland*. This region is characterized by uniformly dipping reflectors at middle and lower crustal levels. The dip is towards the outer hinterland. The inner hinterland ranges in width from 50 to nearly 300 km and may represent the edge of the craton.

(4) *Outer hinterland*. This region has a highly layered lower crust and strong Moho reflections (fig. 4.11b). The depth to the top of the reflective lower crust varies between 10 and 20 km, and reflections from the Moho suggest a thickness of 1–6 km. Crust with these characteristics is not limited to orogenic belts and has been described from continental rifts and cratons as well. In terms of reflection characteristics, it is the most widespread type of continental crust.

Oceanic Areas

Reflection profiling in oceanic areas has been especially important in revealing detailed structure of continental margins. Interpretation of reflecting horizons in

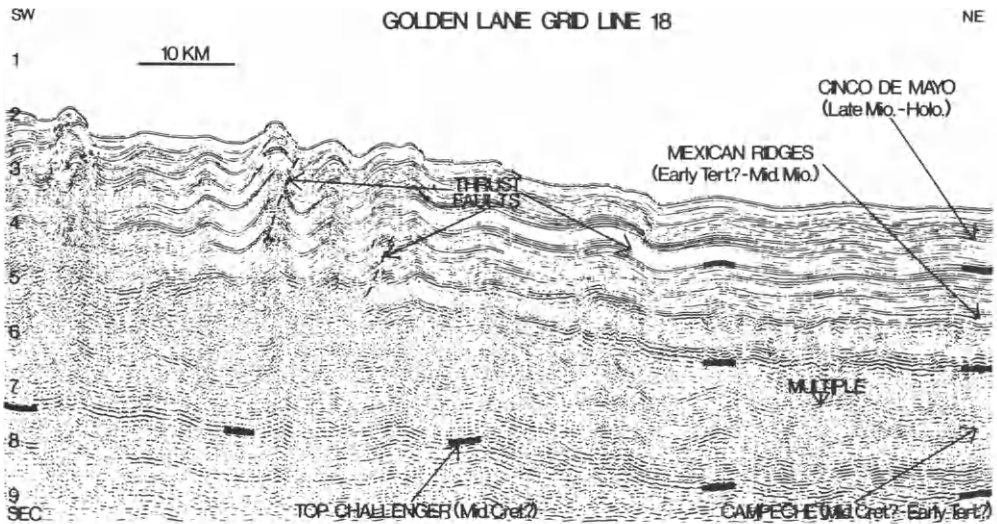


FIG 4.12 (a) Seismic reflection profile across the western continental margin of the Gulf of Mexico. From Buffler et al. (1979).

some instances has been verified from cores from the Deep Sea Drilling Project. Typical reflection profiles of two types of continental margins are given in fig. 4.12. The profile from the western margin of the Gulf of Mexico (a) exhibits detailed stratigraphy and structure in the upper 2–3 km. The strong discontinuous reflections of the Cinco de Mayo and Mexican ridges and the Campeche units are interpreted as alternating fine-grained pelagic sediments and turbidites. Reflections in the Challenger unit may represent, in part, deep-water carbonates. The most striking features of the profile are the nearly symmetrical folds with wave lengths of the order of 10 km and topographic relief of 500–700 m. These folds act as barriers to young sediments transported down the continental slope. Thrust faults also occur in the section and dip chiefly toward the continent at steep angles. The folds are probably not caused by salt doming, in that the salt-bearing unit (the Challenger unit) continues relatively undisturbed beneath the folds. The folds are interpreted to have developed in a seaward-sliding decollement triggered either by regional uplift to the west followed by gravity sliding into the Gulf of Mexico, or by compressive forces from the west.

A profile across the Aleutian trench off southern Alaska exhibits extremely complex structure (fig. 4.12b). The strong continuous reflection near the middle of the section appears to represent the contact of deformed ocean sediments with metamorphic basement. The upper strong reflector in the trench represents the contact between trench turbidities and oceanic sediments, whereas the strongest reflection in the trench occurs between the oceanic sediments and oceanic crust. These reflections can be followed for about 10 km toward the trench slope before they are cut by faults. Although they are deformed by steep faults which dip toward the continent, the reflections can be traced as far as the midslope terrace, where they appear to plunge into the continent. The curved features in the lower slope are diffractions caused by complex structure related to downslope mass movement and

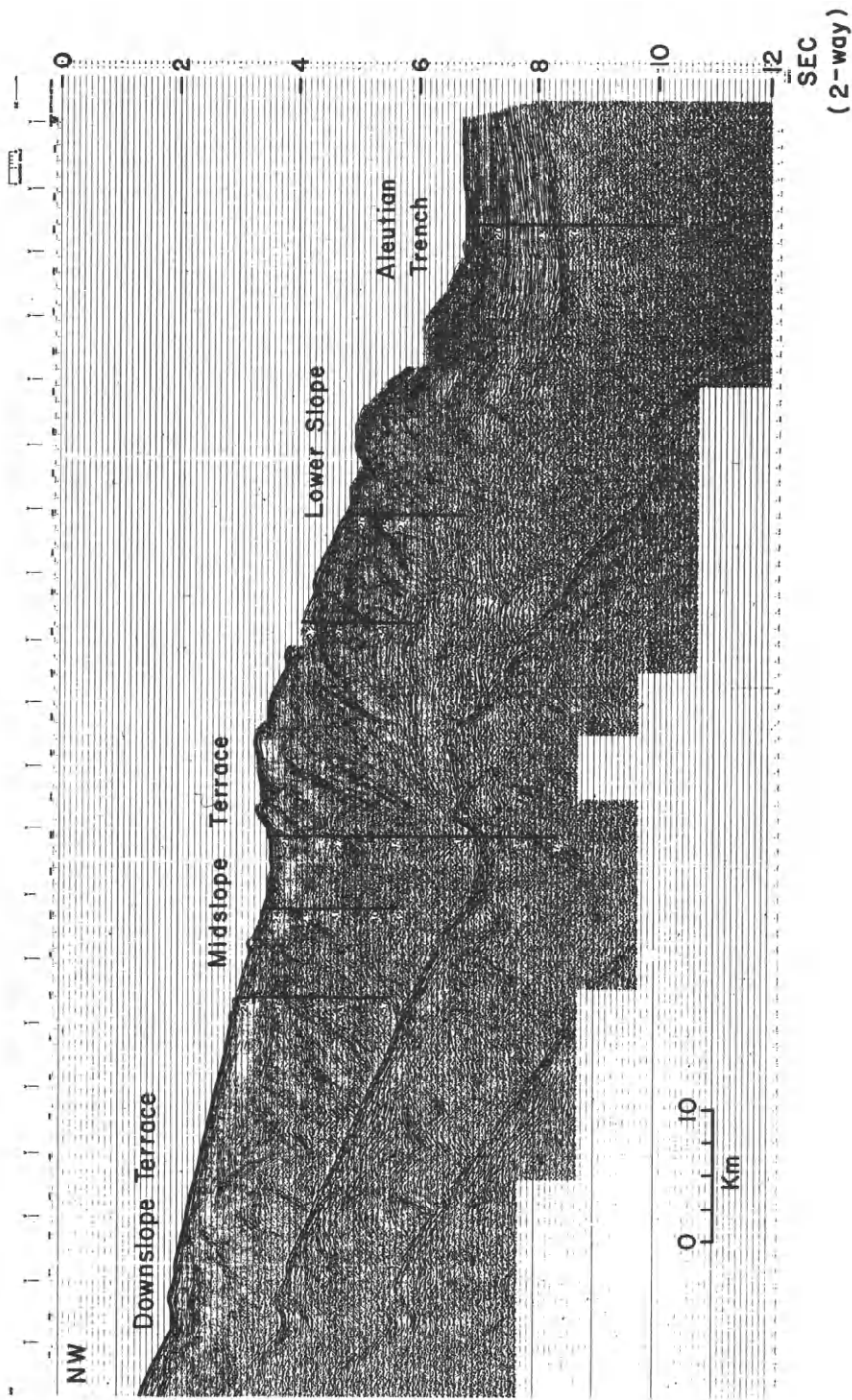


FIG 4.12 (b) Seismic reflection profile across the Aleutian trench near Kodiak Island, Alaska (from Von Huene, 1979).
Depth in (Km = 3 times the 2-way travel time (secs)).

faulting. Topographic ridges in this region may represent the surface expression of thrust faults dipping landward. The downslope terrace (forearc basin) appears to be composed of undeformed turbidites at shallow depth.

Origin of Reflections in the Lower Continental Crust

The strong reflectors in much of the middle and lower part of the continental crust have been relegated to many different causes (Smithson, 1986; Klemperer, 1987): (1) multiple intrusion of mafic sills; (2) layered igneous intrusions; (3) deeply buried sediments; (4) mylonite zones; (5) strain-banding developed from ductile deformation; and (6) open, fluid-filled cracks. Although more than one mechanism may be responsible for deep seismic reflections, it seems probable that such a widespread phenomenon has relatively few underlying causes. Data suggest that the depth to the zone of pronounced seismic reflectivity decreases as surface heat flow increases (Klemperer, 1987), and thus, that the reflections are related to temperature. Highly reflective zones are formed only at temperatures greater than 300–400°C. The correlation of reflectivity and heat flow, furthermore, suggests that crustal reflectivity must vary on the same time scale that heat flow varies (~ 10 Ma). Of the causes listed above, only ductile strain banding or fluid-filled cracks in the lower crust are capable of changing over geologic time. The other causes are relatively permanent features of the crust. The depth of the ductile–brittle transition in the crust increases as temperature declines, and thus the depth of ductile strain bands also increases. Fluids in cracks, as well as dehydration reactions, will also change depth in response to changing crustal temperatures and deformational styles. One advantage of calling upon fluids to cause deep crustal reflections is that fluids can also explain the relatively high electrical conductivity of the lower crust as discussed later in the chapter.

Heat Flow

Heat-flow Distribution

Surface heat flow in continental and oceanic crust is controlled by many factors, and for any given segment of crust it decreases with age. One of the most important sources of heat loss from the Earth appears to be hydrothermal circulation association with ocean ridges. Model calculations suggest that approximately 25% of total global heat flow can be accounted for by hydrothermal transport (Davies, 1980). Heat flow on continents and islands varies as a function of the age of the last magmatic event, the distribution of heat-producing elements, and erosion level. Considering all sources of heat loss, the total heat loss from the Earth is about 42×10^{12} W, which is composed of 12×10^{12} W from the continents and 30×10^{12} W from the oceans (Sclater et al., 1980). The equivalent heat flows are 55 and 95 mW/m², respectively, for a worldwide average heat flow of 81 mW/m². The difference between the average measured oceanic heat flow (67 mW/m²) and the calculated value (95 mW/m²) (table 4.1) is due to heat losses at ocean ridges by hydrothermal circulation. Models indicate that 70% of the average surface heat flow is lost from convection in the mantle by plate creation at ocean ridges, 20% by conduction and the remainder by radioactive decay of heat-producing elements in the continental crust.

Estimated average heat flows of major crustal types are tabulated in column 4 of the table 4.1. Shield areas exhibit the lowest and least variable continental heat-flow values, generally in the range of 30–40 mW/m², averaging about 38mW/m². Platforms and Paleozoic orogenic provinces are more variable—usually falling between 35 and 70mW/m² and averaging about 55 mW/m². The difference between shield and platform heat flows probably reflects the greater thickness of crust in most platform areas. Post-Paleozoic orogenic areas and island-arcs exhibit high and variable heat flow in the range of 60–80 mW/m². Measurements available from continental rift systems are high and variable. The high heat flow in some island-arcs and oceanic volcanic islands reflects recent volcanic activity in these areas. Heat flow in ocean basins generally falls between 35 and 60 mW/m², averaging about 50 mW/m². Ocean ridges, on the other hand, are characterized by extremely variable heat flow, ranging from less than 50 to greater than 200mW/m². Heat flow decreases as a function of increasing distance and age from ocean ridges (see fig. 4.16). Active and some inactive marginal-sea basins are also characterized by high heat flow (60–80 mW/m²), while most inactive basins exhibit typical ocean-basin heat flows (fig. 4.2). Inland-sea basins as a group exhibit variable heat flow (30–75 mW/m²), which reflects, in part, Cenozoic sedimentation rates in the basins.

Most evidence supports a correlation of high heat flow with thin crust, low P_n velocities, positive travel-time anomalies, shallow depth to the zone of seismic low-velocity, low Q , and high electrical conductivity in the upper mantle.

Heat Production and Heat Flow in the Continents

Heat-flow values are significantly affected by the age and intensity of the last magmatic event(s), the distribution of radioactive elements in the crust, and the amount of heat coming from the mantle. Information from the continents has made it possible to evaluate quantitatively the relative contributions of each of these effects. Continental surface heat flow (q_o) is linearly related to average radiogenic heat production (A_o) of near-surface granitic and gneissic rocks by the following equation:

$$q_o = q_r + A_o D \quad (4.1)$$

q_r and D are constants characteristic of crustal types; q_r , commonly referred to as the *reduced heat flow*, is the intercept value for rocks with zero heat production; and D , the slope of the line relating q_o to A_o , has units of depth and is commonly referred to as the *characteristic depth*. The characteristic depth is interpreted to reflect the thickness of the upper crustal layer enriched in heat-producing elements. The reduced heat flow is the heat coming from below this layer and includes both radiogenic heat in the lower crust and mantle heat.

Examples of the linear relationship between q_o and A_o for several crustal types are given in fig. 4.13. The line representing the Appalachian Mountains is typical of most continental platform areas and pre-Mesozoic orogenic areas. Characteristic depths range from 4.5 to 14 km. Recent studies of uplifted root zones of continental crust, such as the Pikwitonei province in Canada (Fountain et al., 1987), show that estimates of surface heat flow and surface heat production also yield the linear

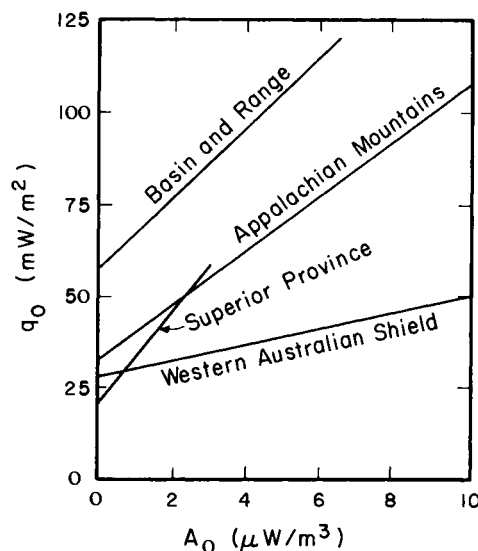


FIG 4.13 Surface heat flow (q_0) versus heat production of near-surface rocks (A_0) for several crustal types.

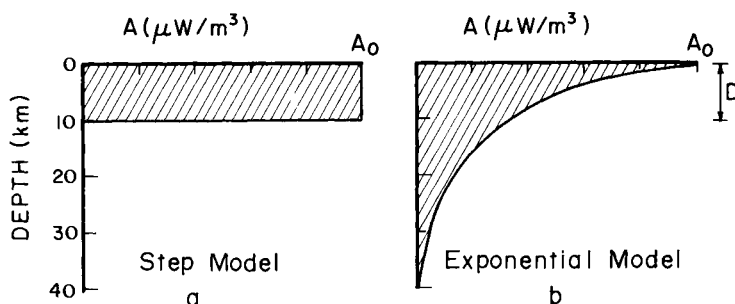


FIG 4.14 Two models of crustal heat production that satisfy the linear relationship between q and A . After Blackwell (1971), copyright by the American Geophysical Union. (a) Constant heat generation to depth D ; (b) heat generation decreasing exponentially with depth.

relationship, indicating that this relationship is independent of crustal level exposed at the surface. The chief factors distinguishing individual crustal types are the reduced heat-flow values (q_r) and, to a lesser extent, the range of surface heat flow and heat productivity values. Reduced heat flow distribution can be used to define *reduced heat flow provinces*.

Three vertical distributions of radioactivity in the crust have been proposed that are consistent with the linear q - A relationship. The step model (fig. 4.14a) involves a heat-producing surface layer of approximately constant thickness D and variable heat production, with q_r representing the combined heat coming from the lower crust and upper mantle (Blackwell, 1971). The exponential model involves an exponential decrease in radiogenic heat production according to the expression,

$$A = A_0 e^{-(x/D)}, \quad (4.2)$$

where A_0 is the near-surface heat production and A the heat production at depth x (Lachenbruch, 1968) (fig. 4.14b). A third model, the linear model (not illustrated), is characterized by a linear decrease in A with depth.

The following lines of evidence favor the exponential model as most nearly representing the actual distribution of radioactivity in the continental crust:

1. Radiogenic heat-production data from a few deep drill-holes (>3 km) suggest exponentially decreasing radioactivity with depth.
2. Radioactivity decreases in granitic rocks as a function of increasing inferred depths of emplacement in the crust (Swanberg and Blackwell, 1973).
3. Studies of the distribution of U, Th, and K in rocks of varying regional metamorphic grade indicate that granulite-facies terranes, which probably formed at lower crustal depths (20–40 km), have concentrations of radioactive elements much lower than upper-crustal metamorphic and plutonic terranes (Chapt. 9).
4. Thermodynamic models based on an equilibrium distribution of heat-producing elements in the Earth's gravitational field are consistent with an exponential decrease of these elements with depth (Turcotte and Oxburgh, 1972a).
5. The exponential model readily preserves the linear relationship of q and A under differential erosion, whereas the step and linear models do not (Lachenbruch, 1970). The low heat flow in shield areas, for instance, is readily explained in the exponential model by erosion of much of the surface layer which is rich in radioactive elements.

Temperature Distribution in the Continents

A *geotherm*, or the temperature distribution with depth in the Earth beneath a given surface location, is dependent upon surface and mantle heat flow and the distribution of thermal conductivity and radioactivity with depth. Estimates of geotherms in four crustal types are given in figure 4.15. Models are calculated assuming steady-state conditions. The shaded areas for each geotherm represent the combined uncertainties of the factors listed above. All three models described in the previous section for the distribution of radiogenic heat production with depth result in geotherms that lie within these shaded uncertainty areas. It is noteworthy that the temperature differences between crustal types are larger than the combined uncertainties of data used in calculating the geotherms. Properties of the geotherms include: (1) a nearly constant gradient in the upper crust, (2) divergence of geotherms resulting in temperature differences of up to 500°C between crustal types at Moho depths, and (3) convergence of the geotherms at depths greater than 250 km where convective heat transfer dominates. Results suggest the following Moho temperatures: Sierra Nevada, $325\text{--}375^{\circ}\text{C}$; Precambrian shield, $350\text{--}400^{\circ}\text{C}$; Mesozoic-Cenozoic orogenic areas, $500\text{--}600^{\circ}\text{C}$; and Basin and Range Province, $650\text{--}750^{\circ}\text{C}$.

Heat Flow and Crustal Thickness

For continental (and transitional) crust ≤ 40 km thick, average heat flow is inversely correlated with crustal thickness. The regression line is

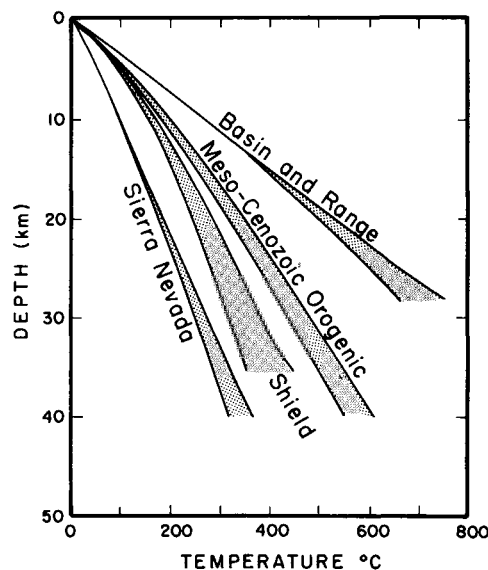


FIG 4.15 Geotherms in four continental crustal types. After Blackwell (1971) and Chapman (1986). Shaded areas represent uncertainty ranges for each geotherm.

$$q_0 = -2.48 (t - 3.48) + 61.2 \quad (4.3)$$

where q_0 is given in mW/m^2 and t in km (Bodri and Bodri, 1985). The correlation coefficient $r = -0.89$. Although various crustal types fall on different parts of the regression line and reflect different temperature regimes at depth, all must reflect similar rates of thinning of about 4 km for an increase in heat flow of 10 mW/m^2 . For continental crust >40 km thick, data suggest that heat flow and crustal thickness are positively correlated.

The origin of the inverse q - t relationship is not well understood. One possibility is that thickening of the continental crust occurs primarily by the underplating of mafic rocks with relatively low radiogenic heat production. Thus, the thicker the crust, the greater the amount of uplift and erosion of the upper heat-producing layer. In addition to the inverse correlation of heat flow and crustal thickness, high surface heat flow also generally corresponds to the following properties of the upper mantle: low P_n velocities, shallow depth to the LVZ, positive travel-time anomalies and high electrical conductivity. This suggests that crustal thickness is related to the depth of partial melting in the upper mantle and that convective upcurrents or plumes in the mantle may be responsible for the inverse q - t relationship.

The positive correlation of q and t in thick continental crust may reflect thickening of the continents during plate collisions (see Chapt. 6) since most of the sites with $t > 40$ km come from young collisional mountain belts. Thickening of the crust during a collision increases the total radiogenic heat contribution for a given column of rock above the Moho.

Age Dependence of Heat Flow

Average surface heat flow decreases with the average age of crustal rocks in both oceanic and continental areas (Sclater et al., 1980; Morgan, 1985). The range of

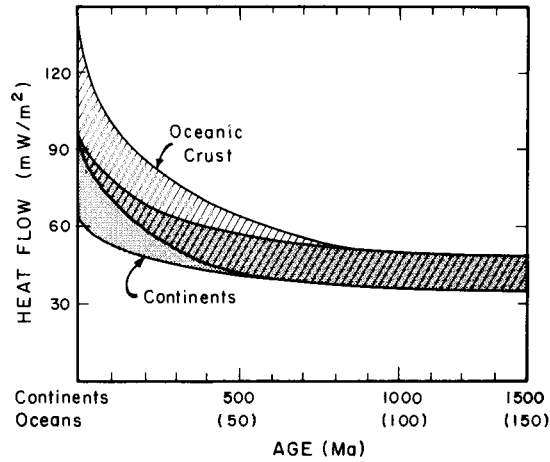


FIG 4.16 Age dependence of average heat flow variation for continental and oceanic crust. After Sclater (1971) and Morgan (1985). Numbers in parentheses on horizontal axis refer to ages in oceanic areas.

variation of average heat flow in continental and oceanic crust is plotted as a function of increasing mean crustal age in figure 4.16. Continental heat flow drops with age to an approximately constant value of 40–50 mW/m² in about 1 Ga and in oceanic areas, similar constant heat-flow values are reached in only 50–100 Ma. The continental heat flow can be considered in terms of three components, all of which decay with time (Vitarello and Pollack, 1980). Radiogenic heat in the upper crust contributes about 40% of the total heat flow in terranes of all ages. The absolute amount of heat from this source decreases with time in response to erosion of the upper crust. A second component of heat, contributing about 30% to the heat flow in Cenozoic terranes, comes from the residual energy of igneous activity associated with orogeny. These two components decay rapidly in a few hundred million years. The third and generally minor component comes from convective heat from within the mantle.

In oceanic areas, heat flow (q_0) falls off with the square root of the age of the crust in 10^6 years (a), according to

$$q_0 = 11.3 a^{-1/2} \quad (4.4)$$

to approximately 120 Ma (Sclater et al., 1980). As the lithosphere cools it contracts, and the depth to the ocean floor increases going away from ocean ridges. The depth of the ocean floor (d) back to about 70 Ma can be approximated by

$$d \text{ (meters)} = 2500 + 350 a^{1/2} \quad (4.5)$$

if the lithosphere is considered the upper thermal boundary of the mantle.

The age of the oceanic crust increases as a function of distance from ocean ridges and as a function of decreasing mean elevation as a consequence of seafloor spreading. The decrease in heat flow with distance to an active ocean ridge reflects cooling of new crust that was formed at the ridge by injection of magma. Models

TABLE 4.2 *Summary of Parameters for Reduced Heat flow Provinces*

Province	Age (Ma)	q_0 (mW/m ²)	q_r (mW/m ²)	D (km)
Basin and Range (USA)	0–65	77	63	9.4
Eastern Australia	0–65	72	57	11
Appalachian Mountains (USA)	100–400	49	28	7.5
United Kingdom	300–1000	69	24	16
Western Australian Shield	≥2500	39	30	4.5
India Shield	≥1800	71	38	15
Superior Province (Canada)	≥2500	34	22	14
Baltic Shield (Scandinavia)	≥1800	33	22	8.5

q_0 and q_r are average surface and reduced heat flow, respectively; D is the characteristic depth; from Vitorello and Pollack (1980) and Pujol et al. (1985).

assuming a lithospheric plate 50–100 km thick with basal temperatures in the range of 550–1500°C are consistent with observed heat-flow distributions adjacent to ocean ridges which decay to equilibrium levels of about 40 mW/m² in 50–100 Ma.

Reduced Heat Flow Provinces

It is clear from the data in table 4.2 that reduced heat flow values also decrease with the average age of the continental crust. No relationship, however, appears to exist between the characteristic depth D and mean crustal age. The reduced heat flow decays within 200–300 Ma to a value between 20 and 30 mW/m², which is about a factor of three faster than surface heat flow decays. Corresponding equilibrium q_r values for oceans are 30–40 mW/m². D is generally equated with the thickness of a surface layer in which most of the heat-producing elements are concentrated. Two interpretations of the change in D between reduced heat flow provinces have been proposed (Jessop and Lewis, 1978): D values reflect differences in crustal erosion level or differences in degree of fractionation of the crust. Because the crustal level exposed at the surface as deduced from metamorphic mineral assemblages does not correlate with D, nor does D correlate with mean crustal age, the first interpretation seems unlikely. The alternative interpretation necessitates variable intensities of fractionation of U, Th, and K into the upper crust which is not related to the age of the crust.

Continental and Oceanic Equilibrium Heat Flow

As described above, both continental and oceanic heat flow decay to similar equilibrium values of 40–50 mW/m². One model which accommodates this similarity is summarized in table 4.3. The model assumes a 10 km thick oceanic crust and a 40 km thick continental crust, with an exponential distribution of radiogenic heat sources in the continental crust. Equilibrium reduced heat flow values (q_r) are selected from the range of observed values in both crustal types. Both cases have similar temperatures by depths of 100 km, the average thickness of the lithosphere. The results clearly imply that more mantle heat is entering the base of the oceanic lithosphere (44 mW/m²) than of the continental lithosphere (12 mW/m²). Hence,

TABLE 4.3 *Heat Flow Model of Ocean Basin and Precambrian Shield*

Ocean basin $q_0 = 50 \text{ mW/m}^2$			Precambrian shield $q_0 = 38 \text{ mW/m}^2$		
(km)	A ($\mu\text{W/m}^3$)	q (mW/m^2)	(km)	A ($\mu\text{W/m}^3$)	q (mW/m^2)
10	0.42	4	35	0.7*	25
90	0.02	2	65	0.02	1
Mantle	0.02	44	Mantle	0.02	12

* $A = A_0 e^{-(x/D)}$ where $A_0 = 2.5 \mu\text{W/m}^3$ and $D = 10 \text{ km}$.

when a plate carrying a Precambrian shield moves over oceanic mantle, the surface heat flow should rise until mantle convective systems readjust so they are again liberating most heat beneath oceanic areas. This may account for the relatively high surface heat flow of the India Shield (table 4.2), which has moved over oceanic crust in the last 50 Ma (Chap. 8).

Gravity Anomalies

Continental Areas

Average Bouguer gravity anomalies expressed in milligals are given in table 4.1 for each major continental and oceanic crustal type. Shield and platform areas are characterized by broad regional anomalies of -10 to -50 mgal and occasional sharp anomalies of local importance. Anomalies with widths of hundreds of kilometers reflect inhomogeneities in the lower crust or upper mantle, and those of smaller size reflect near-surface rock types of fault zones. Gravity anomalies can be used to trace structural trends or rock units in Precambrian shields beneath the sedimentary cover in platform areas. One example is the Keweenaw basalts (1 Ga in age), which can be traced southwest from Lake Superior beneath the sedimentary cover as the Mid-Continent Gravity High (fig. 4.17). Contacts between Precambrian crustal provinces in shield and platform areas (such as between the Grenville and Superior Provinces in Canada) are also often expressed by gravity anomalies.

Phanerozoic orogenic areas are commonly characterized by large negative Bouguer anomalies of -200 to -300 mgal . Anomalies parallel near-surface structural and lithologic elements in these areas and may reflect thicker crust and warmer (low-density) upper mantle. Young orogenic belts near continental borderland and island-arc areas exhibit smaller anomalies (-50 to $+100 \text{ mgal}$). Large negative anomalies in the Basin and Range Province and over most continental rifts reflect thinning of the lower crust and the presence of shallow, low-density upper mantle in these areas. Small positive gravity anomalies superimposed on regional negative anomalies occur in the centers of some rifts and appear to reflect near-surface intrusions of mafic magma (Darracott et al., 1972; Daggett et al., 1986).

Oceanic Areas

Bouguer anomalies in oceanic volcanic islands range from large and positive ($+250 \text{ mgal}$) to small and variable (-30 to $+45 \text{ mgal}$). Negative Bouguer anomalies

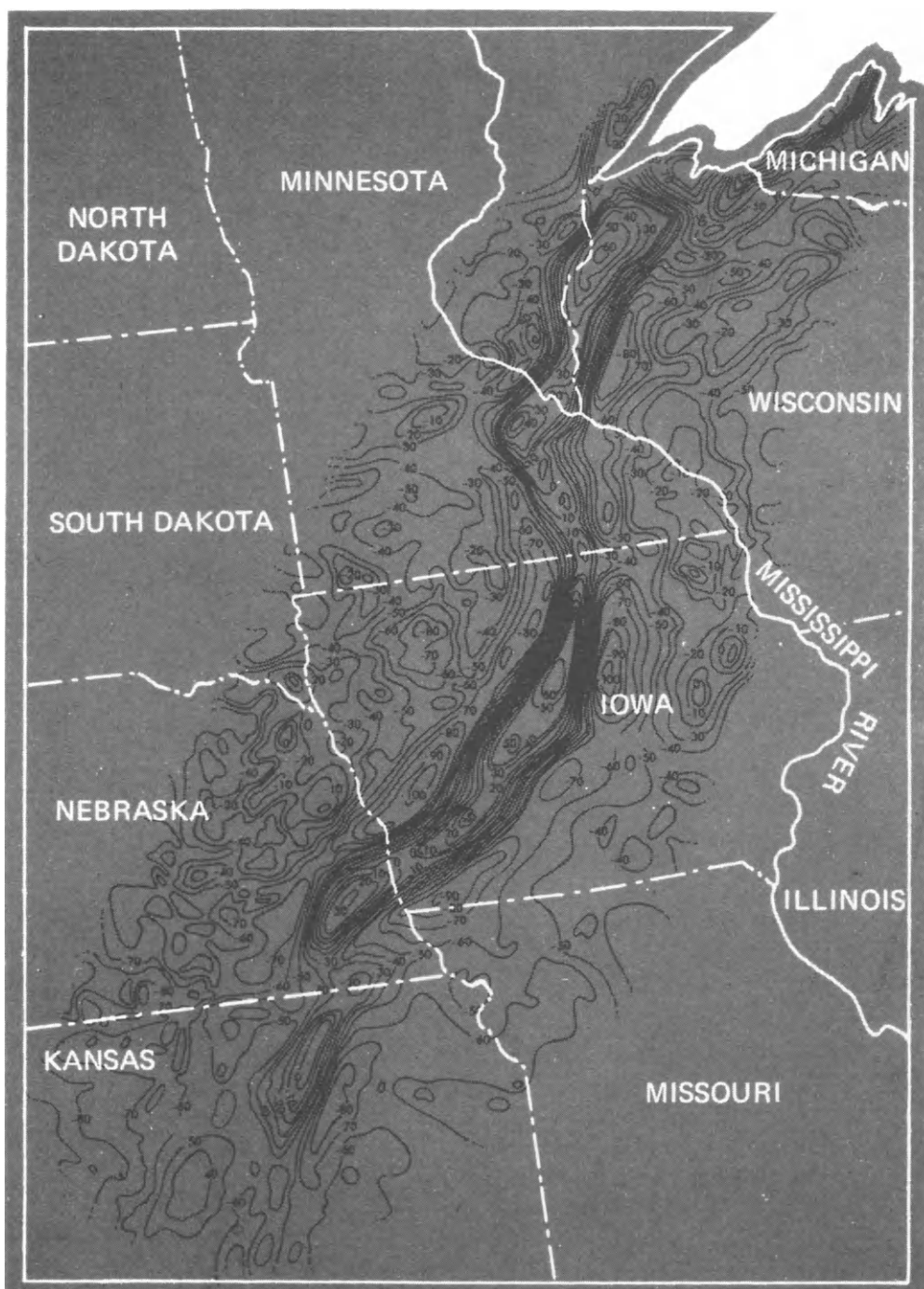


FIG 4.17 Bouguer gravity anomaly map of the north-central United States showing the Mid-Continent Gravity High. From Bouguer Anomaly Map of the United States, U.S. Geological Survey (1964). Contours in milligals.

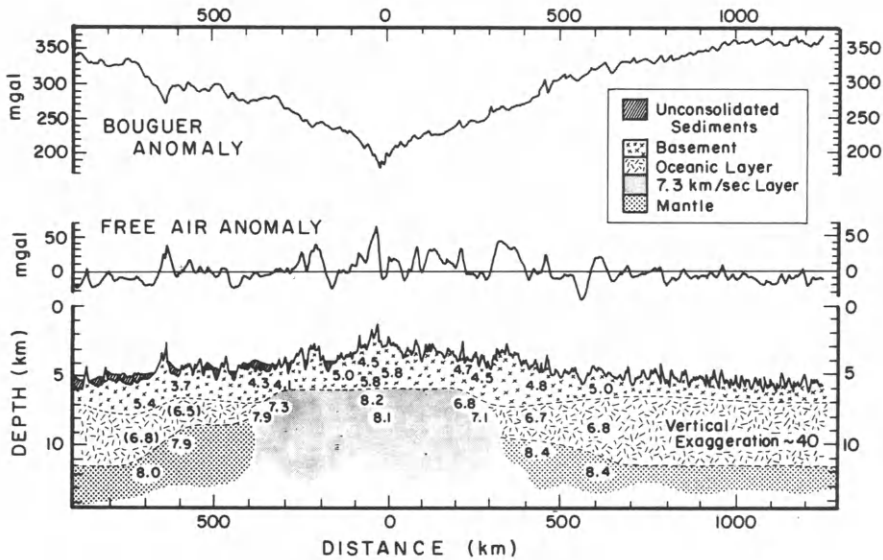


FIG 4.18 Gravity anomaly profiles and seismic structure across the northern Mid-Atlantic ridge. From Talwani et al. (1965), copyright by the American Geophysical Union. P-wave velocities are indicated.

observed over trenches reflect the descent of the less dense lithosphere into the upper mantle. Some gravity profiles suggest thinning of the crust on the oceanward side of trenches. Bouguer anomalies range from +250 to +350 in ocean basins, from +200 to +250 on ridges, and from 0 to +200 over marginal and inland-sea basins. Bouguer and free-air anomaly profiles across the northern Mid-Atlantic ridge are given in figure 4.18. Crustal structures are also shown from seismic data. In the North Atlantic, the Bouguer anomaly increases away from the Mid-Atlantic ridge (+200 mgal) toward the continents (to values of +350 mgal). The minimum in the Bouguer anomaly over ocean ridges suggests that they are approximately isostatically compensated by hot, low-density, upper mantle beneath them. Free-air anomalies in oceanic areas reflect, for the most part, topography on the ocean floor, as can be seen from inspection of the profile in figure 4.18. Smaller Bouguer anomalies in marginal and inland-seas result from thicker sediment layers in these basins.

Discussion

It has long been known that Bouguer anomalies roughly increase with the thickness of the crust. The relationship can be expressed by the following equation (after Grushinsky, 1967);

$$C = C_0 - x\Delta g, \quad (4.6)$$

where C is crustal thickness (in km), Δg is the Bouguer anomaly (in mgal), and C_0 and x are constants that depend on the crustal type. Generally, C_0 ranges from 31 to 38 km and x from 0.06 to 0.03. Oceanic crust has low values of C_0 and continental crust has high values. Because of the variation in C_0 and x , however, it is

TABLE 4.4 *Mean Velocities, Isostatic Anomalies and Depth to Moho*

V_{P_n} (km/sec)	Mean Crustal Velocity (km/sec)	Isostatic Anomaly (mgal)	Depth to Moho (km)
7.7	6.1	-30	25
8.0	6.4	-5	36
8.3	6.6	+20	46

After Woollard (1972).

not possible to calculate crustal thickness from a knowledge of only the Bouguer anomaly.

Isostatic anomalies are small over most crust, indicating a close approach to isostatic equilibrium. The major exceptions are arc-trench systems where lithosphere is being consumed by the asthenosphere. Thick roots exist under some mountains (eg., the Sierra Nevada) but not under others (eg., the Rocky Mountains), suggesting that the Airy hypothesis of isostatic compensation (Chapt. 1) is not always the dominant compensation mechanism. Lateral variations in the density of the crust and upper mantle as implied by seismic data tend to support the Pratt compensation mechanism. It is likely that both mechanisms operate in varying degrees in different areas. Isostatic compensation must occur on a time scale of less than 50 Ma, since both direct and indirect estimates of continental erosion rates indicate that otherwise, continents would be eroded to sea level in 25–50 Ma. A rough correlation exists between P_n velocity, mean crustal velocity, isostatic anomaly, and depth to Moho, as shown by the data in table 4.4. It is probable that isostatic transfer of mass occurs in the asthenosphere.

Isostatic gravity anomaly maps enhance short-wavelength anomalies caused by bodies at or near the Earth's surface and they emphasize regional fabrics (Simpson et al., 1986). Anomalies less than several hundred kilometers wide are particularly useful in identifying major lithologic or structural breaks in the continental crust. Most positive anomalies whose sources are known, reflect mafic rocks in rift settings or accreted or upthrust mafic lower crust. Most lows can be explained in terms of thick sediments, major felsic volcanic fields, batholiths or to downwarped crustal sections associated with continental collisions.

Magnetic Anomalies

Continental Areas

Magnetic anomalies in continental areas usually have amplitudes of a few hundred gammas, widths of ten to hundreds of kilometers, and lengths of hundreds to thousands of kilometers (MacLaren and Charbonneau, 1968; Zietz et al., 1971). They reflect upper crustal rock types or isothermal surfaces of Curie point distributions. In both shield and platform areas, magnetic anomalies characteristically exhibit broad "swirling" patterns. In Archean shield areas granitic-gneiss terranes usually correlate with magnetic highs and mafic volcanic belts with lows (fig. 4.19). Since granitic rocks do not exhibit large magnetizations, the magnetic

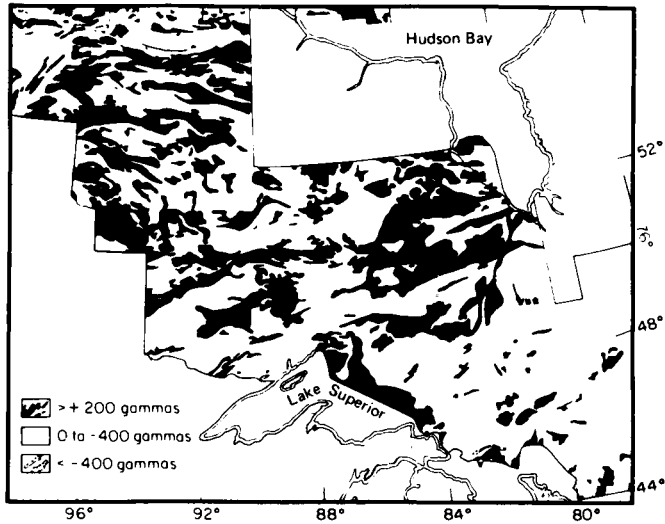


FIG 4.19 Magnetic anomaly patterns in the southern Canadian Shield. After Morley et al. (1967). Magnetic lows reflect mafic volcanic belts and highs (in black) are areas of granite and gneiss. Used with permission of Geological Survey of Canada.

highs of these terranes must reflect terranes beneath the granites and gneisses that are more magnetic than those beneath the mafic volcanic belts (Hall, 1968). Major near-surface intrusions, both mafic and felsic in composition, usually exhibit magnetic signatures. Structural discontinuities such as major faults and crustal-province boundaries also produce magnetic anomalies. Magnetic anomalies, like gravity anomalies, provide a means of tracing major rock units and structural discontinuities beneath sediment cover in platform and continental shelf areas.

Magnetic anomaly patterns in Phanerozoic orogenic areas and island-arcs are more complex and broken up, and show a wide range of amplitudes. Broad anomalies appear to be related to deeper crustal inhomogeneities and shallow, narrow anomalies to near-surface faults or mafic and ultramafic plutons. Overall, magnetic grain generally parallels major orogenic structures. However, in some young orogenic areas and, in particular, in those areas with high heat flow and thin crust, the magnetic grain does not parallel surficial geologic features, and anomaly amplitudes are generally less than 200 gammas. For instance, the Basin and Range Province, which exhibits overall N–S structural trends, is characterized by dominantly E–W-trending, small-amplitude magnetic anomalies. The small amplitude of magnetic anomalies in these regions is generally interpreted to indicate that the lower crust and/or mantle are above the Curie point of magnetite (578°C), the most common magnetic mineral. These peculiar anomaly trends may be related to older lithologic or structural elements in the crust at intermediate depths.

Oceanic Areas

Oceanic crust is characterized by linear to curvilinear magnetic stripes that roughly parallel ridge crests (figure 4.20). The anomalies are typically 5–50 km wide,

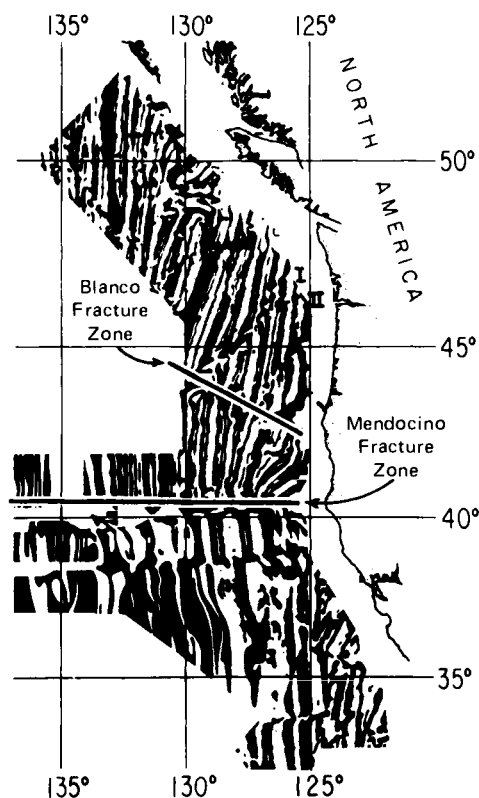


FIG 4.20 Linear magnetic anomaly patterns in the northeastern Pacific. After Raff and Mason (1961). Positive anomalies are black.

hundreds of kilometers long and range from about 400 to 700 gammas in amplitude (relative to sea level). The fact that they are roughly symmetrical about ocean ridges is an important feature that led to the seafloor spreading hypothesis. Magnetic anomalies are offset along transform faults such as the Blanco and Mendocino fracture zones shown in figure 4.20. Magnetic anomalies on the seafloor are produced in basalts emplaced in the central rift zones of ocean ridges by reversals in the Earth's magnetic field. Anomalies are less distinct in marginal-sea basins caused in part by the comparatively high temperatures beneath the basins (fig. 4.2), above or close to Curie temperatures. Linear anomalies have been mapped in the East Scotia and Tasman Sea basins, indicating the presence of now-extinct spreading centers in these basins (Barker, 1972; Hayes and Ringis, 1973). In some inland-sea basins the thick sediment cover appears to mask or partially mask magnetic anomalies. Weak anomalies varying from linear and subparallel to irregular-shaped have been detected in the Gulf of Mexico (Yungul, 1971).

Electrical Conductivity

Models for the electrical conductivity distribution in the continental crust are nonunique and complex (Keller, 1971; Hoak and Hutton, 1986). One or more high-conductivity layers are recognized between 20 and 40 km depth in most continental

crust. Highly conducting layers may correlate with low-velocity zones in the crust, as found in eastern New Mexico for instance (fig. 4.8). Linear crustal conductivity anomalies have been described from several continental regions. An example is the North American Central Plains (NACP) anomaly, which extends for a distance of nearly 200 km from southeastern Wyoming to the edge of the Canadian Shield in Saskatchewan (fig. 3.9) (Camfield and Gough, 1977). This anomaly lines up with major shear zones exposed in the Canadian Shield in northern Saskatchewan and with the Cheyenne shear zone in southeastern Wyoming (see fig. 4.32). It is possible that NACP is produced by dehydration reactions in a shear zone now chiefly covered by cratonic sediments extending from Wyoming to central Saskatchewan. Such a shear zone may represent a collisional plate boundary extending across north-central North America.

Two origins are generally considered for the high-conductivity of the middle and lower crust: the presence of water and partial melting. A third possibility, the presence of conductive graphite zones, may also be of local importance. Experimental data, combined with geothermal results, indicate that partial melting is unlikely in all but young continental crust in active orogenic areas. Aqueous fluids with high ionic contents are the most probable cause of highly conducting crustal layers (Hoak and Hutton, 1986). The normal lower crustal conductivities of 10^{-2} – 10^{-3} S/m can be caused by as little as 0.1–0.01% water. As previously discussed, free water may also be responsible for the strong seismic reflections in the lower continental crust, as well as for crustal low-velocity zones. One possible source of water in the lower crust is from dehydration reactions accompanying metamorphism. However, free water should diffuse upwards resulting in a dry lower crust, and thus some means of replenishing the water is necessary. The upper mantle is another possible source for water, and also provides a means of continuously recharging the lower crust as water is lost upwards. It has also been suggested that an impermeable zone in the lower crust traps water and slows its upward diffusion.

Metamorphism

Burial of rocks in the Earth's crust results in progressive metamorphism as the rocks are subjected to increasing pressure and temperature. Given that metamorphic reactions are more rapid with rising than with falling temperature, metamorphic mineral assemblages may record the highest P–T regime to which rocks have been subjected. Later uplift and erosion make it possible to directly study rocks that were once buried to various depths in the crust but are now exposed at the Earth's surface. Such rocks contain metamorphic mineral assemblages "arrested" at some burial depth, or *metamorphic grade*. Minor changes in mineralogy known as *retrograde metamorphism* may occur during uplift, but such changes commonly can be identified without great difficulty by studying textural features of the rocks. Progressive metamorphism is accompanied by losses of H₂O and other volatile constituents, with some high-grade terranes being almost anhydrous. Other geochemical changes may also accompany metamorphism, although changes in major-element concentrations are generally considered to be of minor importance.

Zones of increasing metamorphic grade can be classified into *metamorphic facies*, which represent limited ranges of load pressure, temperature and water content in

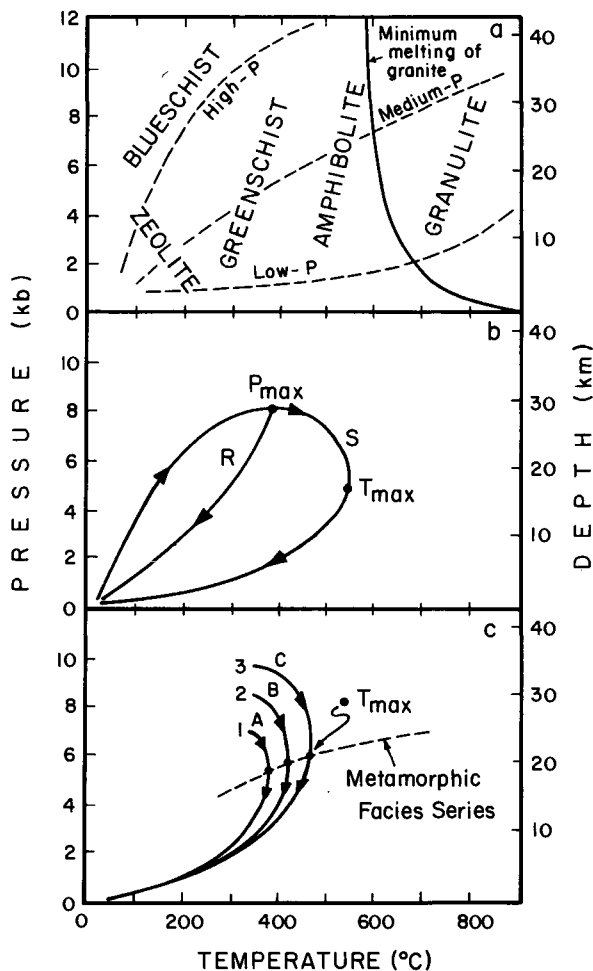


FIG 4.21 (a) Metamorphic facies and facies series (dashed lines) of regional metamorphism. After Miyashiro (1973) and Ernst (1973). P, Pressure. (b) Schematic pressure—temperature diagram showing burial trajectory followed by rapid (R) and slow (S) uplift. After Thompson and Ridley (1987). (c) Schematic pressure—temperature diagram showing a single burial event that transposes rocks to three depths, 1, 2, and 3. Also shown are the uplift trajectories, maximum temperatures (A, B, C) and the facies series.

the crust. Five major facies of regional metamorphism are recognized (fig. 4.21a). The *zeolite facies* is characterized by the development of zeolites in sediments and volcanics, and reflects temperatures generally less than 200°C and burial depths less than 5 km. The *greenschist facies* is characterized by the development of chlorite, actinolite, epidote and albite in mafic volcanics, and of muscovite and biotite in pelitic rocks. *Blueschist-facies* assemblages form at high pressures (>5 kbar) yet low temperatures (<400°C), and are characterized by such minerals as glaucophane, lawsonite, and jadeite. The *amphibolite facies* is characterized by kyanite, staurolite, and sillimanite in metapelites and by plagioclase and hornblende in mafic rocks. The *granulite facies* is characterized by a sparsity or absence of hydrous minerals and the appearance of pyroxenes.

Using experimental petrologic and oxygen isotope data it is possible to estimate the temperature and pressure at which metamorphic mineral assemblages crystallize. From these results, burial depths of metamorphic belts now exposed at the Earth's surface can be constrained. Because geothermal gradients vary from place to place in the Earth, the succession of metamorphic facies should also vary. A succession of metamorphic facies from a given geographic area is called a *facies series*. Three major facies series are recognized (fig. 4.21a). The low pressure series reflects a relatively steep geothermal gradient ($\geq 25^\circ\text{C}/\text{km}$) and is characterized by such minerals as andalusite, cordierite and sillimanite. The medium-pressure series reflects intermediate geotherms ($10\text{--}25^\circ\text{C}/\text{km}$) and is characterized by kyanite and garnet, and the high-pressure series ($\leq 10^\circ\text{C}/\text{km}$) and is characterized by glaucophane, jadeite and lawsonite. Low-pressure metamorphic belts of all ages are known, but they are particularly common in Archean terranes. Medium-pressure belts have also been described from crustal provinces of all ages, but they are uncommon in the Archean. Blueschist terranes make their first appearance in the late Proterozoic and do not become common until after mid-Paleozoic. These terranes appear to have formed on the trench side of arc systems in response to depressed geotherms in subduction zones. Their absence in pre-late Proterozoic terranes may reflect higher geothermal gradients beneath arc systems during the Precambrian.

Calculation of pressure–temperature–time (P–T–t) paths for metamorphic mineral assemblages is feasible if definitive mineral assemblages are found and if adequate experimental data exist. It is also necessary to employ various radiogenic isotopes with different blocking temperatures to date specific points on a P–T–t path. Although it is rarely possible to rigorously quantify P–T–t paths, data suggest that both facies series and P–T–t paths are characteristic of specific tectonic settings (Miyashiro, 1973).

Although many regional metamorphic terranes have undergone more than one period of metamorphism, thus making it very difficult to reconstruct P–T–t paths, much can be learned about crustal evolution by considering one-stage metamorphic models. From such models it is clear that for slow rates of uplift ($<1\text{ mm/a}$), rising segments of buried rocks will be heated due to the concentration of radiogenic heat sources at high levels in the continental crust. Because of this, the maximum temperature achieved on a P–T–t path is at a depth shallower than the maximum burial depth. Examples of P–T trajectories for single burial–uplift events are shown in figure 4.21b and c. In b, path R shows rapid burial followed by rapid uplift which preserves a high-P, low-T assemblage (P_{max}). Path S is for slower uplift, which results in recrystallization of the high-P assemblage and preservation of an assemblage recording the maximum temperature (T_{max}). In c, a single burial event transposes rocks to three different depths 1, 2, and 3 followed by slow uplift along trajectories A, B, and C, respectively. In this case the facies series preserved at the surface is approximately given by the line connecting T_{max} on each uplift curve.

Crustal Composition

Several approaches have been used to estimate the chemical and mineralogical composition of the crust. One of the earliest methods was based on chemical analysis

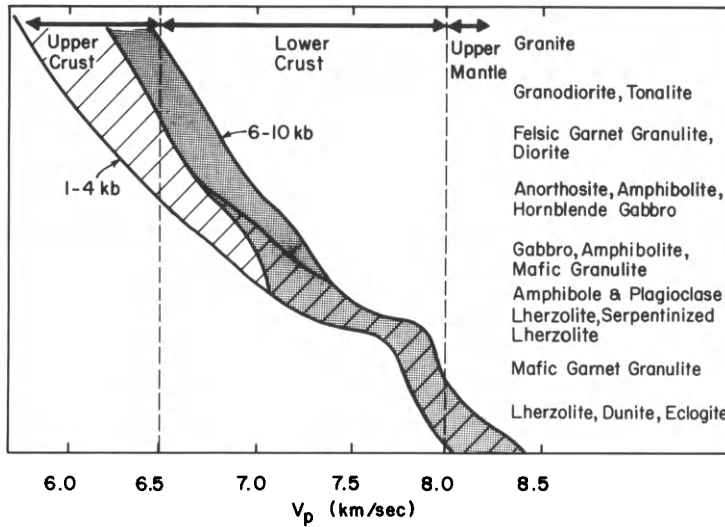


FIG 4.22 Measured P-wave velocity distribution in various igneous and metamorphic rocks at 6–10 kb and 1–4 kb pressure. Data from Birch (1960) and Christensen (1966). The general range of continental crust and upper mantle P-wave velocities are indicated by the arrows.

of glacial clays, which were assumed to be representative of large portions of the continents (Goldschmidt, 1954). Later estimates were based on mixing average basalt and granite compositions in ratios generally ranging from 1:1 to 1:3 (Taylor, 1964a) or by weighting the compositions of various igneous, metamorphic, and sedimentary rocks according to their inferred abundances in the crust (Poldervaart, 1955; Ronov and Yaroshevsky, 1969). More recently, it has been possible to estimate the composition of the continents based on the results of extensive sampling of rocks of various metamorphic grade from Precambrian shields (Eade and Fahrig, 1971; Holland and Lambert, 1972). Such rocks were formed at various depths in the crust and have been uplifted and exposed at the surface. From the results of experimental studies it is possible to outline roughly the P–T conditions under which metamorphic mineral assemblages form. More recent estimates of crustal composition employ seismic velocity measurements in various rocks, the chemical composition of fine-grained clastic sediments, exposed crustal cross-sections both of oceanic (ophiolites) and of continental crust, results from deep drilling on the ocean floor, and xenoliths of deep crustal rocks brought to the surface during volcanic eruptions.

Continental Crust

Seismic Constraints

One constraint on the composition of both the continental and oceanic crust comes from seismic velocities measured in the laboratory on rocks at elevated pressures. P-wave velocities measured at pressures corresponding to upper continental depths (1–4 kb) are consistent with an upper crust composed of large amounts of granitic rocks (or their metamorphic equivalents) (fig. 4.22). In terms of the lower crust,

measured seismic velocities allow a wide range of igneous and metamorphic rock types. Even felsic gneisses may be important at middle and lower crustal depths where $V_p \leq 6.8$ km/sec. Many lower crustal sections, however, have P-wave velocities in the range of 7.0–7.8 km/sec, and thus require a substantial proportion of mafic rocks in the lower crust. Clearly, seismic velocity measurements alone do not provide unique constraints on the composition of the crust.

Sampling of Precambrian Shields

Widespread sampling of metamorphic terranes exposed in Precambrian shields, and especially in the Canadian Shield, has provided an extensive sample base to estimate both the chemical and lithologic composition of the upper part of the Precambrian continental crust (Shaw et al., 1986). Both individual and composite samples have been analyzed. Results indicate that although the upper crust is lithologically heterogeneous, granitic rocks of largely granodiorite to tonalite composition dominate, and the weighted average composition is that of granodiorite.

Use of Fine-grained Terrigenous Sediments

Clastic sediments, and especially fine-grained terrigenous sediments, may represent well-mixed samples of continental crust. To use sediments to estimate crustal composition, however, it is necessary to evaluate losses and gains of elements during weathering, erosion, and diagenesis. Elements, such as rare earths (REE), Th and Sc that are relatively insoluble in natural waters and have short residence times in seawater ($<10^3$ years) may be transferred virtually quantitatively into terrigenous clastic sediments (Taylor and McLennan, 1985). The remarkable uniformity of REE in pelites and loess compared to the great variability observed in igneous source rocks attests to the efficiency of mixing during erosion and deposition. Studies of such elements as the REE indicate they are also very immobile during weathering and diagenesis, and ratios such as La/Sc and La/Yb are relatively resistant to change during these processes. REE distributions are especially constant in shales and resemble REE distributions in weighted averages from Precambrian shields. If the REE distributions and ratios of other relatively immobile elements in shales reflect the average composition of the upper continental crust, it is possible to use these distributions to evaluate abundance levels of other elements as proposed by Taylor and McLennan (1985). Estimates of the average composition of the upper continental crust using the composition of shales are in remarkable agreement with the weighted averages determined from Precambrian shields.

Crustal Sections

From combined field and laboratory studies it is now clear that partial sections of the continental crust are exposed at the Earth's surface. Most or all of these sections have been exposed in response to continental collisions (Chapt. 6). During such collisions, continental crust on the overriding plate is thrust over continental crust on the descending plate and erosion exposes deep crustal levels in the thrust plate (see fig. 6.21a). Metamorphic grade increases abruptly across the thrust and vastly dissimilar geologic provinces may occur on opposite sides of the thrust. Metamor-

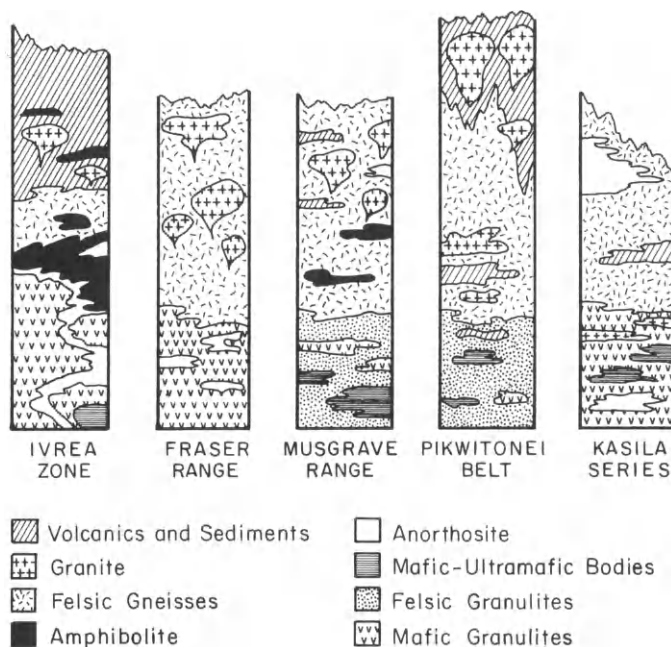


FIG 4.23 Generalized cross-sections of continental crust based on exposed sections of deep crustal rocks. Modified after Fountain and Salisbury (1981).

phic grade decreases gradually in going away from the thrust in the upper plate, reflecting progressively shallower crustal levels.

Examples of five partial sections of continental crust constructed from geologic mapping and geobarometry studies of metamorphic mineral assemblages are given in fig. 4.23. Each section is a schematic diagram illustrating the relative abundances of major rock types and the base of each section is a major thrust fault. The greatest depths exposed in each section are the order of 25–35 km. Each column has a lower granulite zone with mafic granulites (see fig. 4.24) dominating in three sections, and felsic granulites in the other two. The sections indicate considerable compositional variation at all metamorphic grades attesting to the heterogeneity of the continental crust at all depths. Mafic and ultramafic bodies and anorthosites occur at deep levels in some sections and probably represent layered igneous sheets intruded into the lower crust. Volcanic and sedimentary rocks are also buried to great depths in some sections. Intermediate and upper crustal levels are characterized by large volumes of intrusive tonalite to granodiorite.

More than anything else, the crustal sections indicate considerable variations in lithologic and chemical composition both laterally and vertically in the continental crust. The only large-scale progressive change in the sections is an increase in metamorphic grade with depth. Major changes in composition of the basal granulite terranes are consistent with the variation in P-wave velocities observed in the lower crust (fig. 4.22). Although there is no evidence for a widespread Conrad discontinuity in the sections, rapid changes in lithology may be responsible for discontinuities of a more local extent such as those observed in seismic reflection profiles.

Xenolith Studies

Xenoliths of crustal and upper-mantle rocks are common in kimberlite pipes and in some alkali basalts. If one can determine the depth from which xenoliths come, and estimate the relative abundances of various xenolith populations in the crust, it is possible to reconstruct a crustal cross-section (Kay and Kay, 1981). Although metamorphic xenoliths can be broadly ordered in terms of their crustal depths, metamorphic mineral assemblages in many lower crustal xenoliths are not definitive in determining depths more precisely. Even more difficult is the problem of estimating the relative abundances of xenolith types in the crust. Some lithologies may be oversampled and others undersampled by ascending kimberlite pipes. Hence, extreme caution should be used in reconstructing crustal sections from xenolith populations.

Xenolith-bearing volcanics and kimberlites occur in many different tectonic settings, thus giving a wide lateral sampling of the continents. Lower crustal xenoliths from arc volcanics are chiefly mafic in composition, and xenoliths of sediments are rare to absent. These results suggest that the root zones of modern arcs are composed chiefly of mafic rocks. Xenoliths from intraplate volcanics and kimberlites are compositionally diverse and have complex thermal and deformational histories. Metasedimentary rocks, however, are minor compared to metaigneous rocks. In general, xenoliths of mafic granulite are more abundant than felsic granulite, suggesting that a mafic lower crust is also important in cratons (Rudnick and Taylor, 1987).

Compositional Estimates

The average chemical composition of the upper continental crust is very well known from widespread sampling of Precambrian shields, geochemical studies of shales and exposed crustal sections. Data suggest an average composition similar to granodiorite (table 4.5, col. 1). The composition of the lower crust is less well constrained. Crustal sections, xenolith populations and the frequent occurrence of high P-wave velocities in the lower crust (7.0–7.8 km/sec) suggest that a large part of the lower crust is mafic in overall composition (table 4.5, col. 2) (fig. 4.24). In some collisional zones, however, the lower crust may be felsic as exemplified by exposed crustal sections (such as the Pikwitonei and Musgrave sections in fig. 4.23 and sections exposed in southern India) (table 4.5, col. 3). However, these crustal sections are underlain today by 35–40 km of crust, probably of granulite grade and perhaps composed in large part of mafic granulites. The crust in these areas may have thickened during continental collision (to 70–80 km), thus burying upper crustal rocks to granulite grade (35–40 km). Uplift and erosion of this crust brought the felsic granulites to the surface with a possible mafic granulite root still intact.

Estimates of total continent composition based on 1:1 mixtures of upper continental crust with felsic and mafic lower crust, respectively, are also given in table 4.5 (cols 4 and 5). For a mafic lower crust the bulk continent is similar in major-element composition to andesite, as suggested by Taylor (1977) (table 4.5, col. 6 and fig. 4.25). For a felsic lower crust, however, continents are tonalitic in composition. In terms of incompatible elements (Chapt. 7), a felsic lower crust results in a continent enriched in Ba, La, and Zr and depleted in Ti relative to a

TABLE 4.5 *Estimates of Average Compositions of the Continental and Oceanic Crust*

	Continental						Oceanic
	Upper	Lower		Total			7*
	1	2	3	4	5	6	
SiO ₂	65.5	49.2	61.0	57.4	63.3	58.0	49.6
TiO ₂	0.5	1.5	0.5	1.0	0.5	0.8	1.5
Al ₂ O ₃	15.0	15.0	15.6	15.0	15.3	18.0	16.8
FeOT	4.3	13.0	5.3	8.7	4.8	7.5	8.8
MgO	2.2	7.8	3.4	5.0	2.8	3.5	7.2
CaO	4.2	10.4	5.6	7.3	4.9	7.5	11.8
Na ₂ O	3.6	2.2	4.4	2.9	4.0	3.5	2.7
K ₂ O	3.3	0.5	1.0	1.9	2.2	1.5	0.2
MnO	0.1	0.2	0.1	0.2	0.1	0.1	0.2
P ₂ O ₅	0.2	0.2	0.2	0.2	0.2	0.2	0.2
Rb	110	2	10	56	60	42	4
Ba	800	50	780	425	790	350	60
Sr	325	500	570	410	450	400	180
La	30	10	20	20	25	19	3.5
Yb	2.0	1.0	1.2	1.5	1.6	2.2	2.7
Zr	220	30	200	125	210	100	100
Nb	25	3	5	14	15	11	5
U	2.5	0.1	0.1	1.3	1.3	1.3	0.2
Th	11	0.3	0.5	5.7	5.8	4.8	0.6
Cr	35	200	90	118	63	55	230
Ni	20	150	60	85	40	30	80

*Excluding sediments. Sediments contribute $\leq 5\%$ to crustal composition. Oxides in weight percent; trace elements in ppm. FeOT, total Fe as FeO.

1. After Taylor and McLennan (1985, table 2.6.2) and Shaw et al. (1986, table 1).
2. Mafic lower crust from Shaw et al (1986, table 3) and miscellaneous xenolith compositions.
3. Felsic lower crust from Weaver and Tarney (1984, table 2).
4. 1:1 mixture cols. 1 and 2.
5. 1:1 mixture cols. 1 and 3.
6. Average calc-alkaline andesite (from Taylor and McLennan, 1985, table 3.3).
7. Average ocean ridge basalt from miscellaneous sources.

continent with a mafic lower crust (fig. 4.25). Compared to average calc-alkaline andesite, the mafic lower crustal model is somewhat enriched in most incompatible trace elements. If the upper continental crust is felsic in composition and the lower continental crust largely mafic in composition, as most data suggest, how do these two layers form and how do they persist over geologic time? This question will be returned to in Chapter 10.

Oceanic Crust

In some respects the oceanic crust is more accessible than the continental crust for geochemical sampling. Fragments of oceanic crust are preserved on the continents as ophiolites (Chapt. 6), some of which contain "fossil Mohos". The chief problem with equating the composition of ophiolites to oceanic crust is that some or most ophiolites appear to have formed in back-arc basins and, in varying degrees, have geochemical signatures of arc systems (Chapt. 7). Other sources of data for

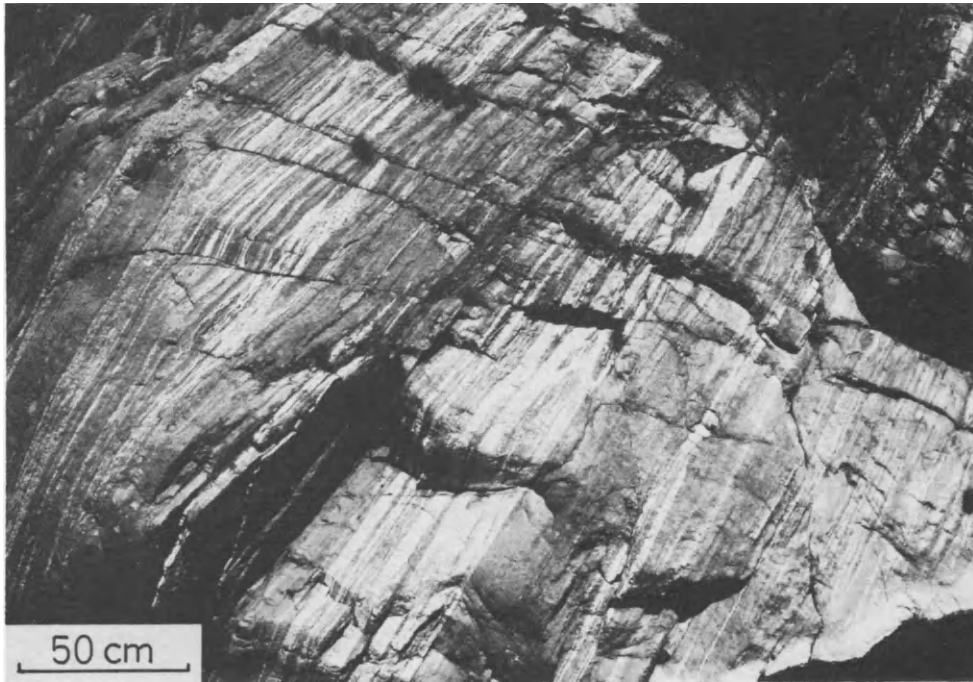


FIG 4.24 Layered mafic granulites from the Ivrea Zone in the Alps. Courtesy of K. R. Mehnert. Similar rocks may comprise large volumes of the lower continental crust.

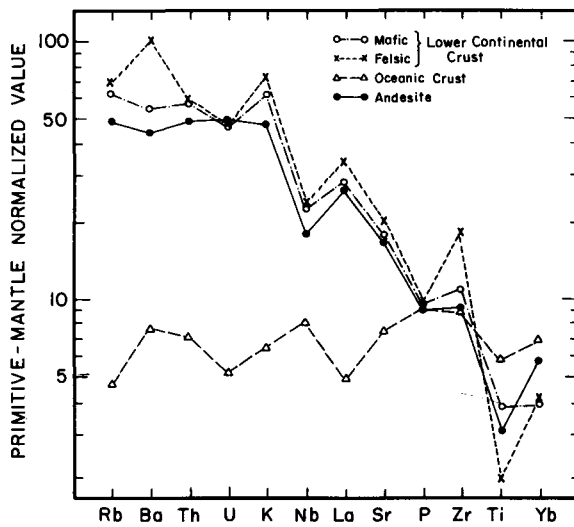


FIG 4.25 Incompatible element distributions in continental crust, average calc-alkaline andesite, and in oceanic crust. The two estimates for continental crust are from columns 4 and 5, table 4.5. Element concentrations are normalized to primitive mantle composition. From such a graph it is possible to observe relative and absolute deviations of crustal composition from mantle composition.

estimating the composition of oceanic crust are dredge samples from the ocean floor and drillcores retrieved from the Deep Sea Drilling Project. Studies of ophiolites and P-wave velocity measurements are consistent with basement and oceanic layers being composed largely of mafic rocks metamorphosed to the greenschist or amphibolite facies. The sediment layer is composed of pelagic sediments of variable composition and extent.

An estimate of the composition of oceanic crust is given in table 4.5 (col. 7). It is based on chemical analyses of ophiolites that do not have a subduction zone geochemical component (Chapt. 7) and from data for dredge and core samples that have penetrated the basement layer. Pelagic sediments, which contribute <5% to crustal composition, are ignored in the estimate. The average composition is that of tholeiitic basalt. Although ophiolites contain minor amounts of ultramafic rock and felsic rock, they are much less variable in lithologic and chemical composition than crustal sections of continental crust, suggesting that the oceanic crust is rather uniform in composition. Oceanic tholeiitic basalts are extruded or intruded at ocean ridges and undergo varying degrees of alteration by reaction with seawater and hydrothermal fluids. Fresh ocean ridge basalts are very low in highly incompatible elements (such as Rb, Ba, Th, U, K), but these elements are added to the basalts during alteration, giving rise to a flat and irregular pattern when normalized to primitive mantle (fig. 4.25). However, element contents are still much lower than corresponding contents in continental crust, as indicated on the same diagram.

Crustal Provinces

Introduction

Stockwell (1965) suggested that the Canadian Shield can be subdivided into structural provinces based on differences in structural trends and style of folding (fig. 4.26). Structural trends are defined by foliation, fold axes, bedding and sometimes by geophysical anomalies. Boundaries between the provinces are drawn where one trend cross-cuts another, along either unconformities or structural-metamorphic breaks. Large numbers of radiometric dates from the Canadian Shield indicate that structural provinces are broadly coincident with age provinces. Similar relationships have been described on other continents, and lead to the concept of a crustal province. A *crustal province* is a segment of the continental crust which records a similar range of radiometric dates and exhibits similar deformational histories. Crustal provinces range from 10^2 to 10^3 km wide and $>10^3$ km in length, and some larger ones have been divided into subprovinces. Linear to curvilinear crustal provinces that have undergone numerous episodes of deformation and metamorphism are sometimes referred to as *mobile belts*. Crustal provinces need not be coincident with crustal types. For instance, a crustal province may extend from a Precambrian shield into an orogenic belt or across more than one orogenic belt. On the other hand, an orogenic belt is both a crustal type and a crustal province. Radiometric dating using multiple isotopic systems is critical to defining and unraveling the complex, polydeformational histories of crustal provinces. In this respect it is useful to review the important radiometric dating methods and the interpretation of radiometric dates.

Radiometric dating is based on the decay of radiogenic nuclides to stable nuclides.

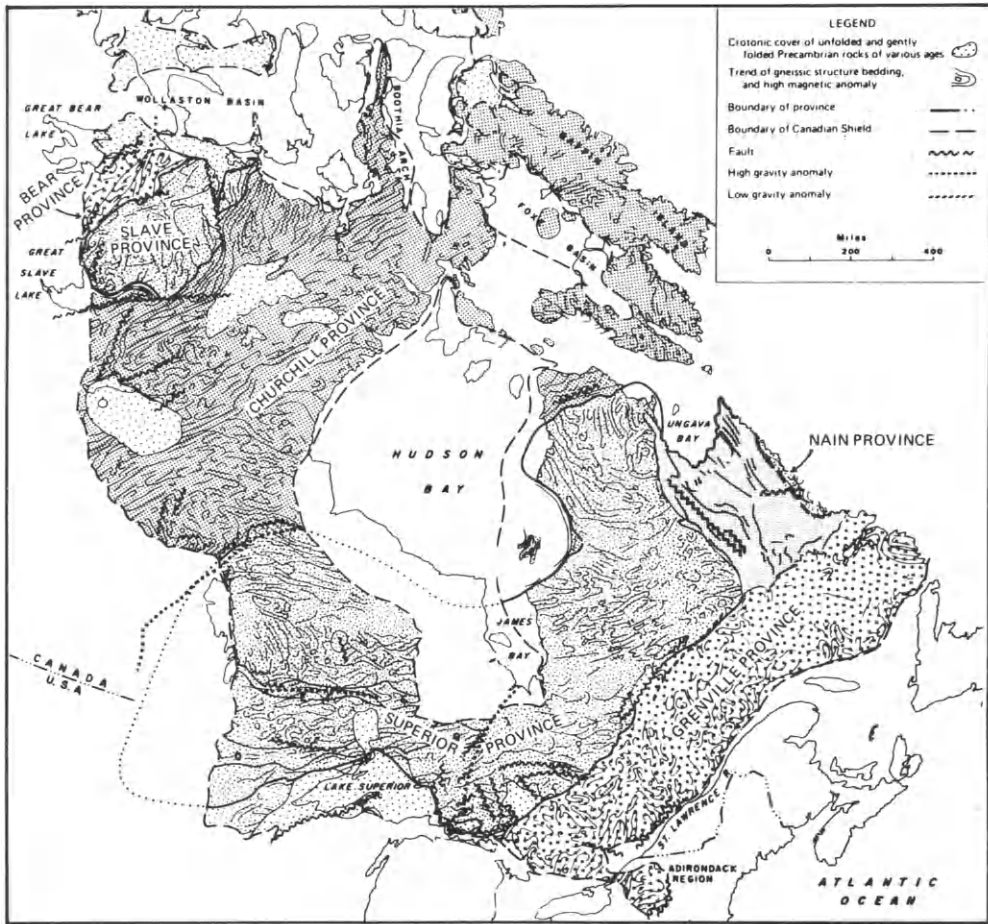


FIG 4.26 Precambrian crustal provinces of the Canadian Shield showing structural trends. After Stockwell (1965).

The following conditions must be satisfied in order to interpret radiometric dates in terms of specific geologic events:

1. The time interval over which a rock or mineral forms must be short compared to its age. This problem can be important in dating young rocks (≤ 10 Ma), especially metamorphic and plutonic rocks, and for authigenic minerals in young sediments.
2. Radioactive decay constants must be accurately known. Decay constants for ^{40}K , ^{235}U , ^{238}U , and ^{232}Th are known to better than 5%, and to better than 10% for ^{87}Rb and ^{147}Sm . Because older values of the decay constants were used to calculate dates prior to 1977, care should be taken when comparing dates to recalculate results using the new decay constants given in Steiger and Jager (1977).
3. The initial isotopic concentration of the daughter isotope (^{87}Sr , ^{206}Pb , etc.) must be determined before an accurate date can be calculated.

4. Losses or gains of parent or daughter isotopes must not have occurred since the time of the event being dated or, if they have, adequate corrections must be made for such losses or gains.

K–Ar Dates

The K–Ar dating method has served to delineate major crustal stabilization in orogenic belts, as well as to date young magmatic events. However, because of the mobility of argon and imprecise knowledge about just when argon becomes immobilized in a rock system, individual K–Ar dates are of limited value in bracketing the age of orogenies. Also, some minerals (such as pyroxenes) commonly accept excess argon into their structures and, if not corrected for, give ages that are too old. K–Ar dates are most useful in studying young, unaltered volcanic rocks and sediments, in detecting incipient metamorphism or alteration, and in defining regional epeirogenic uplift.

⁴⁰Ar/³⁹Ar Dates

In the ⁴⁰Ar/³⁹Ar method, K concentration is determined from ³⁹K by neutron activation, and thus, does not require a separate K determination. The principal advantage of the method is that argon can be released by stepwise heating of samples. In this way a spectrum of dates can be calculated from the ⁴⁰Ar/³⁹Ar ratio of each fraction, and it may be possible, by using different numerals, to relate these dates to stages in the cooling history of a rock. This method is important in aiding to unravel complex polydeformational histories of crustal provinces and in dating rock magnetization for paleomagnetic studies.

Rb–Sr Isochron Dates

The following expression is used to calculate the age of a geologic event using the Rb–Sr dating method (Faure, 1986):

$$(^{87}\text{Sr}/^{86}\text{Sr})_t - (^{87}\text{Sr}/^{86}\text{Sr})_i = (^{87}\text{Rb}/^{86}\text{Sr})_t(e^{\lambda t} - 1) \quad (4.7)$$

where $(^{87}\text{Sr}/^{86}\text{Sr})_t$ and $(^{87}\text{Sr}/^{86}\text{Sr})_i$ are the atomic ratios of the two Sr isotopes at the present time and at the time of the event being dated, respectively; $(^{87}\text{Rb}/^{86}\text{Sr})_t$ is the atomic ratio at the present time; λ is the decay constant of ⁸⁷Rb; and t is the age. This is the equation of a straight line with intercept $(^{87}\text{Sr}/^{86}\text{Sr})_i$ and a slope of $e^{\lambda t} - 1$, which is defined as an *isochron*. All portions of a rock system that formed at the same time, and had the same initial ⁸⁷Sr/⁸⁶Sr ratio [$(^{87}\text{Sr}/^{86}\text{Sr})_i$], and have not lost or gained either parent or daughter isotope since the rock formed, will lie on this line. Extrapolation of the line defined by samples of differing Rb/Sr ratios to ⁸⁷Rb/⁸⁶Sr = 0 gives the $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratio. In order to obtain accurate estimates of both the age and $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratio it is necessary to have a group of samples with a reasonably wide distribution of ⁸⁷Rb/⁸⁶Sr ratios. If an accurate isochron can be determined, the method provides a means of identifying single samples that have lost or gained parent or daughter, since such samples will probably plot off the isochron.

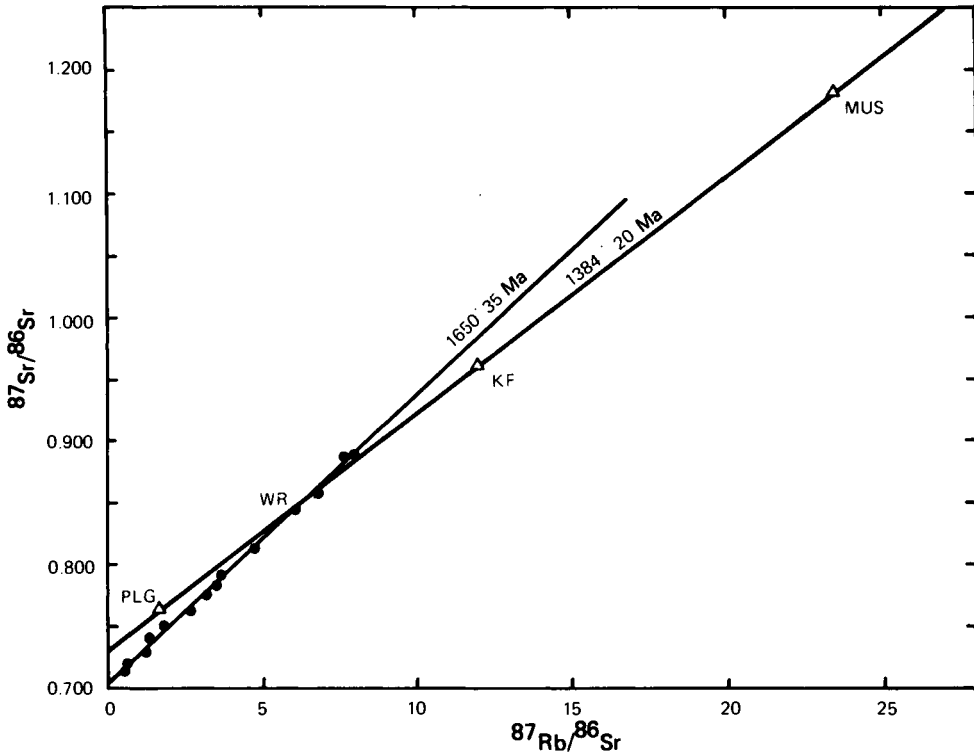


FIG 4.27 Rb–Sr isochron diagram for Precambrian granitic rocks from central Colorado. After Wetherill and Bickford (1965), copyright by the American Geophysical Union. Key: ● = whole-rock samples; △ = mineral separates from whole-rock WR. PLG = plagioclase; KF = K-feldspar, MUS = muscovite.

It is sometimes possible to distinguish two events using the isochron method. An example for Precambrian granitic rocks from central Colorado is illustrated in figure 4.27. Whole-rock samples of the granites define an isochron of 1650 Ma with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.703. When individual minerals in one sample are analyzed, an isochron of 1384 Ma [$(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.726$] is obtained. The latter isochron is interpreted as dating an isotopic exchange on a small scale between minerals, and probably records a time of regional metamorphism. The 1650 Ma whole-rock isochron is interpreted as the original formation date of the granites. However, when U–Pb zircon dates are available from the same rocks, they are often greater than whole-rock Rb–Sr isochron dates, suggesting that the isochron dates may be affected by later metamorphism.

Sm–Nd Isochron Dates

The Sm–Nd isochron dating method is similar to the Rb–Sr method. ^{147}Sm decays by alpha-particle emission to ^{143}Nd with a half-life of about 10^{10} a ($\lambda^{147}\text{Sm} = 6.54 \times 10^{-12} \text{ a}^{-1}$), and the standard isochron expression can be written by normalizing to ^{144}Nd , a stable isotope of Nd,

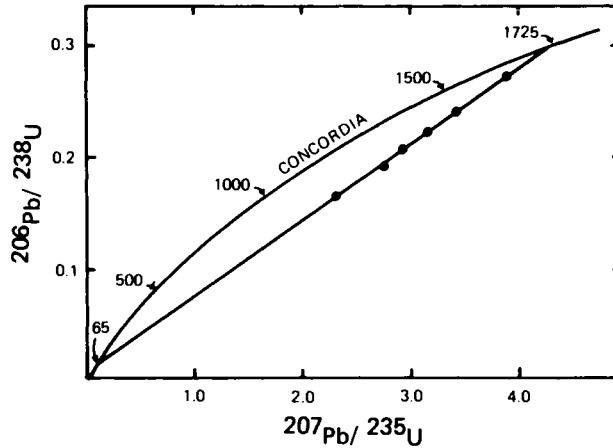


FIG 4.28 U-Pb concordia relationships of zircons from the Boulder Creek batholith, Colorado, interpreted in terms of an episodic lead-loss model. After Stern et al. (1971). Individual zircon analyses = ●; ages on concordia in Ma.

$$(^{143}\text{Nd}/^{144}\text{Nd})_t - (^{143}\text{Nd}/^{144}\text{Nd})_i = (^{147}\text{Sm}/^{144}\text{Nd})_i(e^{\lambda t} - 1) \quad (4.8)$$

where each term has an analogous definition to terms in the Rb-Sr isochron equation (4.7). Because both Sm and Nd are rare-earth elements, they are not strongly fractionated from each other, as are Rb and Sr, and hence the range in Sm/Nd ratios is generally much smaller than the Rb/Sr range on isochron plots. This lack of Sm-Nd fractionation also applies to secondary geological processes such as alteration and metamorphism, and hence the Sm-Nd dating method has the advantage over the Rb-Sr method that it can see through secondary events to primary igneous events. Also, because the Sm/Nd ratio is generally higher in mafic and ultramafic rocks than in felsic rocks, the Sm/Nd method provides a method of dating mafic and ultramafic rocks. Because the Sm-Nd ratio is generally not significantly changed by metamorphism, Sm-Nd dates are thought to record the ages that magma is separated from the mantle to form crust. Such ages are referred to as *crustal formation ages*, and will be discussed more fully in Chapter 10.

U-Th-Pb Dating Methods

Concordia Dates

When $^{238}\text{U}/^{206}\text{Pb}$ and $^{235}\text{U}/^{207}\text{Pb}$ dates for a given rock are the same, they are said to be *concordant*. If the $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ratios are plotted against each other, concordant samples of various ages define a single curve known as *concordia*. In figure 4.28, $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ratios in zircons from the Boulder Creek batholith in Colorado define a chord that intersects concordia at 1725 Ma and 65 Ma. The zircon dates are referred to as *discordant*, and the chord upon which they fall is a line of lead loss (or uranium gain). The upper intersection with concordia is generally interpreted as the initial formation age of the zircons (in this case assumed to be equal to the crystallization age of the Boulder Creek batholith). If an episodic model for lead loss is assumed, the lower intercept gives the age of a thermal event

that resulted in lead loss (65 Ma). The data can also be fitted to a continuous diffusion model for the loss of lead in which the lower end of the chord curves down towards the origin (zero age). Unless independent geologic evidence favors one of these models, it is not possible to decide which model is most suitable for a given rock.

As with Rb–Sr isochron dates, discordant U–Pb dates necessitate a reasonably wide range of $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ratios (or nearly concordant ages) to define accurately the intersection(s) with concordia. Zircon is the most common mineral used in U–Th–Pb dating, in that the initial amount of lead incorporated in zircon is generally negligible compared to uranium and thorium, and zircon is extremely refractory. It is generally agreed that the U–Pb zircon method is the most accurate method to date magmatic crystallization events, and especially events older than 100 Ma. Zircon dates have potentially the greatest resolving power for closely spaced Precambrian igneous events (approximately 5 Ma in rocks 3.0 Ga in age).

It is also possible to analyze individual zircons by evaporation techniques whereby a zircon is progressively evaporated on a mass spectrometer filament. From such an approach, more than one age can be obtained from zircons in the same rock. Use of an ion microprobe allows analysis of various domains within individual zircons that formed at different times. From these methods, it is possible to identify more than one geologic event in the same rock from different zircon populations or from various overgrowths on individual zircons.

Isochron Dates

Equations similar to (4.7) can be written for ^{238}U – ^{206}Pb , ^{235}U – ^{207}Pb , and ^{232}Th – ^{208}Pb , normalizing to the stable isotope ^{204}Pb . In addition, a third isochron expression relating $^{207}\text{Pb}/^{204}\text{Pb}$ to $^{206}\text{Pb}/^{204}\text{Pb}$ can be formulated by combining the ^{238}U – ^{206}Pb and ^{235}U – ^{207}Pb expressions, thus providing a total of four isochron dating relations in the U–Th–Pb system. Because of U loss during secondary geologic processes, the U–Pb isochron methods have not been very successful. The Th–Pb isochron method can be definitive in some geologic systems. The Pb–Pb isochron method has the advantage that it is not sensitive to recent losses of uranium or lead, but depends only on the requirement that all samples have remained closed to early secondary processes.

Common Lead Dates

Common lead is lead in a rock or mineral at the time of formation. Those minerals or rocks with extremely low U/Pb and Th/Pb ratios, such that no significant radiogenic lead has accumulated since they crystallized, are used in common lead dating methods. Galenas, K-feldspars and some whole-rock volcanic samples are examples of samples relatively free from radiogenic contamination. Lead model ages are calculated from a knowledge of present and initial lead isotope ratios ($^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}/^{204}\text{Pb}$), the age of the Earth, and various assumptions difficult to prove regarding the history of the Earth. Initial isotopic ratios are assumed to be equal to those observed in sulfide phases in iron meteorites, which contain negligible uranium and thorium, and the age of the Earth is inferred from meteorite ages. The model dates calculated are interpreted by some as the ages that lead was separated

from mantle source regions, probably by melting, followed by extrusion or intrusion of magma (or ore solutions) into the crust. Another interpretation is that the dates represent the last crust–mantle mixing event in a given area. Because of assumptions that are difficult to evaluate, lead model ages are generally considered the least accurate of the U–Th–Pb radiometric dates.

Disequilibrium Dating

The decay chains of ^{235}U and ^{238}U are broken by the loss of one or more daughter isotopes due to geological processes. Alteration, weathering, chemical or biochemical precipitation, and magma crystallization are examples of processes that disturb the uranium decay chains. It is possible to make use of these separations to date relatively young geologic events by (1) monitoring the activity of the radiogenic nuclide separated from a decay chain, or (2) monitoring the rate of growth of a daughter nuclide produced from a parent previously removed from the decay chain. These methods are particularly important in dating events during the last million years, and fill a gap between ^{14}C and K–Ar dating methods. Applications have been made to measure sedimentation rates in the oceans using the ^{230}Th and ^{231}Pa methods and to date carbonate-precipitating organisms using the ^{234}U and ^{230}Th methods. Growth of glaciers has also been studied employing ^{210}Pb .

Fission Track Dates

Natural fission of ^{238}U leaves damage tracks in glasses and minerals which can be enlarged by etching. The number of tracks, which is a function of the age of the sample and its uranium content, can be counted with a petrographic microscope and used to date the sample, provided certain conditions are met. The two most important conditions are that the samples cooled rapidly and that the track density is not affected by secondary processes. Because tracks anneal in minerals at different rates as a function of falling temperature, the fission-track method can also be used to study the thermal history of rocks and rates of uplift (Wagner and Reimer, 1972).

Radiogenic Isotope Tracer Studies

Radiogenic daughter isotopes continually increase in rocks at a rate proportional to the parent/daughter element ratios (Rb/Sr, Sm/Nd, U/Pb, etc.). Employing this observation it is possible to place constraints on the composition of source rocks from which magmas are derived. For instance, ultramafic source rocks have lower Rb/Sr and U/Pb ratios but higher Sm/Nd ratios than granitic rocks. Hence, magmas produced by partial melting of ultramafic rocks in the mantle will have lower $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios and higher $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than magmas produced by partial melting of granitic crustal rocks. This assumes that the isotopic ratios are not changed during melting or fractional crystallization, and experimental and theoretical considerations support this assumption. Small differences in parent/daughter ratios within the same type of source rock can also be detected, and these results can be interpreted in terms of source heterogeneity (Chapt. 7). It is possible to extend the application of isotopic tracer studies to older geologic terranes by subtracting the amount of radiogenic daughter isotope that has grown in igneous

rocks since their crystallization. The isochron equations can be used to make this correction by solving for the initial daughter isotope ratios.

Events that can be Dated

A radiometric date gives the age that a system was last open to migration of daughter isotopes in and out of a rock or between adjacent minerals. This date, in effect, records a blocking temperature for an isotope in a given mineral or rock. The *blocking temperature* is the temperature at which the diffusion rate of a given isotope in a mineral (or rock) becomes negligible compared to its rate of accumulation. The cooling rate for a rock is also an important factor bearing on blocking temperatures. The temperature at which Ar is immobilized is lower than that for Sr, and the blocking temperature of Sr is lower than that for Pb in zircons. Hence, various isotopic systems in minerals and rocks record different parts of the thermal history of a rock or mineral. It is not always obvious, however, which dates (if any) correlate with specific tectonism, magmatism or other geologic phenomena. The problem is further complicated by the fact that daughter isotopes may be only partially retained if cooling rates are slow, thus giving radiometric dates that may be meaningless in terms of specific geologic events. For these reasons, a rather complete knowledge of geologic field relationships is necessary if radiometric dates are to be accurately interpreted.

Magmatic Events

Magmatic crystallization is one of the most common types of events that can be dated. Felsic igneous rocks are generally more satisfactory for dating than mafic or intermediate igneous rocks, since they contain greater concentrations of radiogenic parent isotopes. An exception is the previously mentioned Sm–Nd system. Magmatic events may record tensional or compressional stresses in the crust, or they may be completely unrelated to any known crustal deformation. Their interpretation in terms of tectonics is critically dependent upon field relationships.

Diagenesis

Diagenetic dates sometimes may be estimated using K–Ar and Rb–Sr isochron methods. Glauconite, diagenetic micas or feldspars, or whole-rock samples are generally used. Several problems are encountered in attempting to obtain diagenetic ages: (1) Ar and, to a lesser extent, Sr leak from diagenetic minerals; (2) it is difficult to separate detrital and diagenetic components in sedimentary rocks; and (3) Sr may not equilibrate within diagenetic phases, resulting in poorly defined isochrons. Extreme care should be used in interpreting radiometric ages obtained from sediments or sedimentary rocks in terms of diagenetic ages, unless the above problems have been evaluated.

Orogenies

The term *orogeny* is used in different ways. In its broadest sense it means mountain building. However, in a more restricted way it refers to crustal deformation involving major folding, which occurs during a limited interval of geologic time. It is

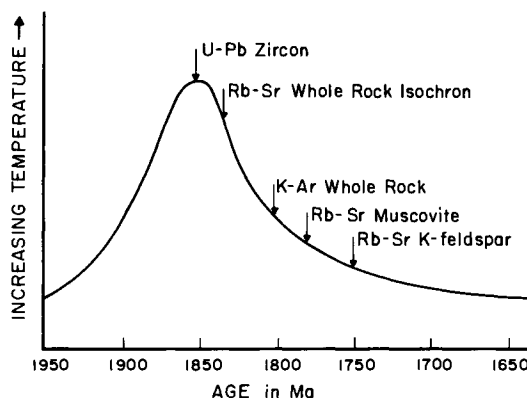


FIG 4.29 Diagrammatic representation of isotopic blocking temperatures associated with an idealized orogeny.

also commonly accompanied by volcanism, plutonism, and regional metamorphism. Regions that have undergone only faulting and broad uplift are excluded from this definition. Orogenies can be dated by dating *syntectonic igneous rocks*, which are igneous rocks emplaced during deformation. Alternatively, times of deformation can be bracketed by dating *pre-tectonic* and *post-tectonic igneous events*. As pointed out by Gilluly (1969), however, one should not equate all orogenies with magmatism, since some orogenies occur with little if any magmatic activity (such as the Antler orogeny in the western United States). Also, plutonism may be post-tectonic or, as with the 1.4–1.5 Ga plutons in the central and southwestern United States, *anorogenic* (ie., emplacement without regional deformation).

The exact times that various daughter isotopes become immobilized during an orogeny are not well known, but most occur after the thermal intensity peak (fig. 4.29). U–Pb zircon dates and Sm–Nd dates coincide most closely with maxima of magmatism, while Rb–Sr and K–Ar dates appear to record regional uplift during the final stages of orogeny. It is possible with Rb–Sr, U–Th–Pb and Sm–Nd dating methods to resolve 10–50 Ma at 3 Ga and in some instances, with U–Pb zircon dates, to resolve as little as 5 Ma at 3 Ga (Krogh and Davis, 1971). Dating of Phanerozoic orogenies suggests that orogenies last from 10 to 50 Ma. Dates from Precambrian mobile belts, on the other hand, commonly span 50–200 Ma, and several individual orogenies may be “hidden” in such a spread of dates.

Phanerozoic and late Precambrian orogenies are of two types: those related to convergent-plate boundaries and those related to continental collisions. Whether or not all Precambrian orogenies can be interpreted in a similar manner is a subject discussed in Chapter 9.

Regional Metamorphism and Uplift

Employing the concept of blocking temperature, isotopic dates represent the time elapsed since a mineral cools through its blocking temperature for a given isotope. A “whole-rock” blocking temperature refers to the average of the mineral blocking temperatures for a given rock. Magnetism is acquired by minerals in a similar manner as they cool through a “magnetic blocking temperature” known as the Curie temperature (Chapt. 5).

Because the blocking temperatures of various isotopes are different in various minerals, it is possible to monitor the history of an orogeny by using different isotopic systems. For instance, if we consider a simple orogeny characterized by a single thermal pulse (fig. 4.29), the Sm–Nd whole-rock and U–Pb zircon dates in syntectonic igneous rocks most closely define the maximum temperature. This assumes that the magmas are derived from the mantle and have not previously resided in the crust. If they had a previous crustal history, the Sm–Nd date may record the first crustal separation age, and the zircon date the time of remelting of the older crust. Rb–Sr whole-rock, K–Ar whole-rock, Rb–Sr mica and Rb–Sr feldspar dates have successively lower blocking temperatures and record progressive uplift and erosion of the orogenic belt. The Rb–Sr whole-rock isochron date may, in some cases, correspond to the maximum temperature of regional metamorphism. When K–Ar and Rb–Sr mineral dates are plotted on a map, each isotope in a given mineral may define a surface known as the *metamorphic veil*. Such a surface corresponds to times when large segments of crust were elevated through specific mineral blocking temperatures and may allow estimates to be made of the rates of epeirogenic uplift. From such studies it appears that the uplift history following major orogeny may last from 50 Ma to, in some instances, 100 Ma, as illustrated in figure 4.29.

Defining Crustal Provinces

The definition of crustal provinces is not always unambiguous. Most crustal provinces contain rocks of a wide range in age. For instance, the Limpopo Belt in Southern Africa (Chapt. 9) includes rocks ranging from 3800 to 2500 Ma in age, and records several periods of complex deformation and regional metamorphism. The Grenville Province in North America records a similar poly-deformational history with rocks ranging in age from 2700 to 1000 Ma. Some parts of crustal provinces represent new mantle-derived crust and others represent reworked older crust. *Reworking* or *overprinting* refer to later periods of deformation, metamorphism, and partial melting that affect older crust. It is possible, in some instances, to map regions of reworked crust and these are sometimes referred to as relict-age subprovinces. In some areas, and especially in central and southern Africa, provinces may cross-cut one another, and the older provinces have superimposed deformational fabrics and metamorphism from the younger provinces. Structural trends within provinces range from linear to exceedingly complex swirling patterns reflecting polyphase deformation (fig. 4.26).

Metamorphic grade is highly variable in crustal provinces ranging from prehnite-pumpellyite to granulite facies. The highest-grade zones are accompanied by migmatite development and occur in the central parts of mobile belts. On the whole there seems to be a time-dependent trend from low-pressure to high-pressure metamorphic series in going from Archean to younger provinces. Mesozoic-Cenozoic provinces commonly exhibit paired metamorphic belts. Some provinces are characterized by extensive thrusting and nappe formation and record substantial shortening of the crust. Examples are the Nagssugtoqidian belt in Greenland and the Alpine–Tethyan belt in Eurasia (Bridgwater et al., 1973). These appear to record one or more continental collisions. Provinces (or portions thereof) that evolve at

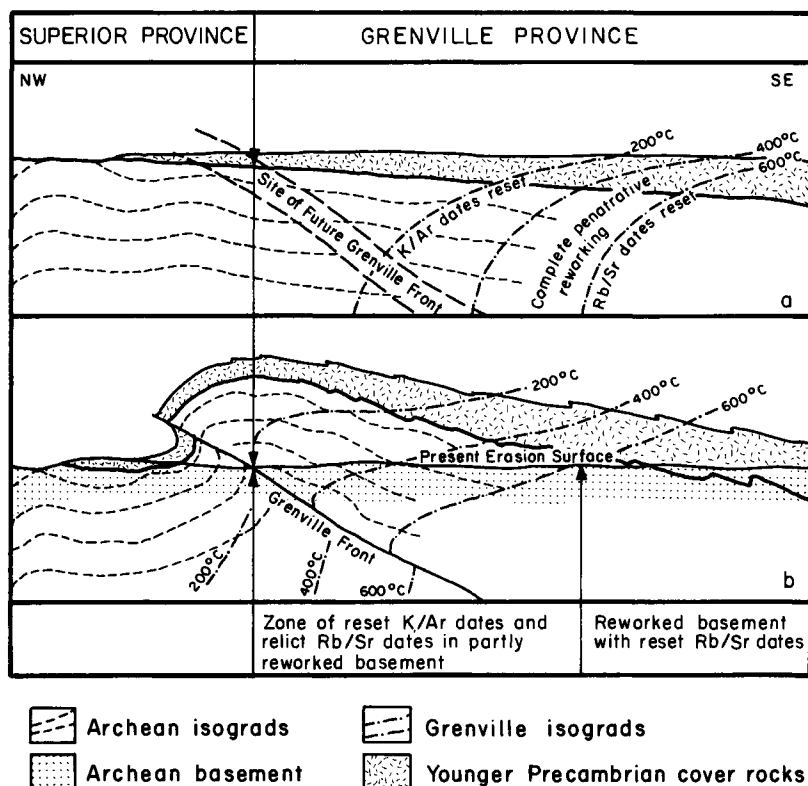


FIG 4.30 Digrammatic cross-section of the Grenville Front before (a) and after (b) uplift. After Wynne-Edwards (1972).

relatively high temperatures and low pressures, like the Ketilidian in southern Greenland and the Hercynian in Europe, are characterized by extensive granitic plutonism and dominantly vertical deformation patterns. Boundaries between crustal provinces may be parallel to the structural trends within adjacent provinces, or one province may cross another at steep angles, as is represented by the contact of the Grenville Province where it crosses Proterozoic provinces in eastern Canada (fig. 4.26). Contacts between crustal provinces fall into one or a combination of four categories: (1) faults or shear zones; (2) unconformities; (3) rapid changes in metamorphic grade; or (4) intrusive contacts. Magnetic and gravity anomalies also characterize provincial boundaries (fig. 4.26), reflecting juxtaposition of rocks of differing densities and magnetic susceptibilities. Changes in crustal thickness may also occur at province boundaries represented, for instance, by the increase from 35–40 km to 45 km at the Grenville Front between the Grenville and Superior Provinces (fig. 4.3).

The Grenville Front has been the subject of numerous geological and geochronological studies. This boundary varies from a shear zone to a metamorphic transition zone along strike (Wynne-Edwards, 1972). It ranges in width from a few to many kilometers and involves the overprinting of one metamorphic episode (the Kenoran at 2.7 Ga) by another (the Grenville at 1.0 Ga). The Grenville isograds cut across Kenoran isograds at steep angles near the Grenville Front (fig. 4.30). K–Ar dates are

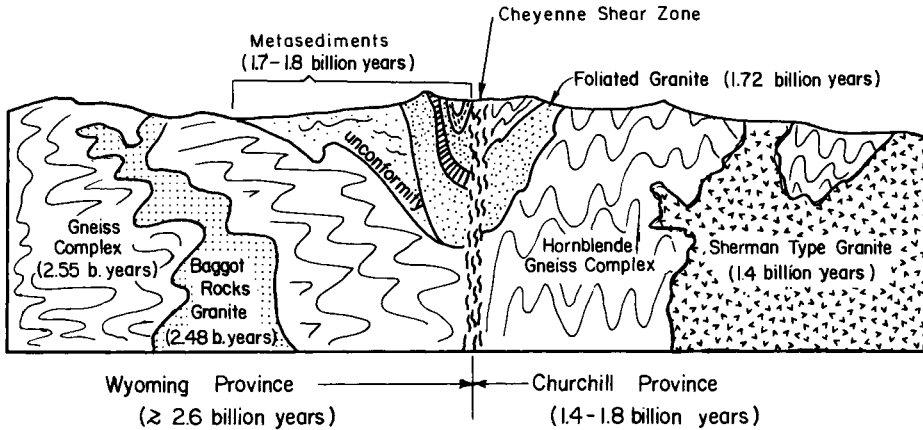


FIG 4.31 Schematic cross-section of the boundary between the Wyoming Province and Proterozoic provinces in southeast Wyoming. Modified after Hills et al. (1968). Location of section shown in figure 4.32.

reset at rather low temperatures ($\sim 200^{\circ}\text{C}$) and gradually decrease from 2.7 Ga to about 1.0 Ga over a zone of variable width across the front. At some locations the Grenville Province is thrust over the Superior Province, transporting older gneisses with completely reset K–Ar dates into contact with Archean rocks of the Superior Province.

Shear zones between provinces may be accompanied by extensive mylonite zones ranging up to tens of kilometers in width. A cross-section of such a boundary between the Wyoming Archean Province and Proterozoic provinces to the south is shown in figure 4.31. The Cheyenne shear zone, which is exposed in the Medicine Bow Mountains in southeastern Wyoming, is a near-vertical shear zone separating Archean gneisses from a gneissic complex (1.8 Ga) and granitic rocks (1.4 Ga). It is complicated by the fact that deformed metasedimentary rocks (1.7–1.8 Ga in age) rest unconformably on the Archean gneisses and are also cut by the shear zone. The shear zone, which is up to several kilometers wide, is composed chiefly of mylonitized quartzo-feldspathic gneisses.

Some province boundary fault zones exhibit transcurrent motions with up to several hundred kilometers of displacement. Others pass from flat-lying to steep structures and may have thrust or transcurrent offsets (Daly, 1986). Boundaries of structural provinces may or may not coincide with age-province boundaries. Age-province boundaries may be gradual, extending for tens to hundreds of kilometers beyond structural boundaries. Such gradational boundaries are produced by a gradual change in either metamorphic grade or in degree of granitization. An example is illustrated in figure 4.32, where a wide metamorphic “overprint” extends north from the Cheyenne shear zone. Although Precambrian rocks crop out at only a few places in this part of the Rocky Mountains, K–Ar mineral dates indicate the presence of a zone several hundred kilometers in width, in which argon in Archean rocks was partially mobilized at 1.8 Ga. K–Ar mineral dates in this zone are generally < 1.8 Ga. The 1.8 Ga contour in figure 4.32 is the maximum distance in from the contact where biotite began to retain argon at 1.8 Ga.

Major continental crustal provinces are shown on Plate I. Because of extensive

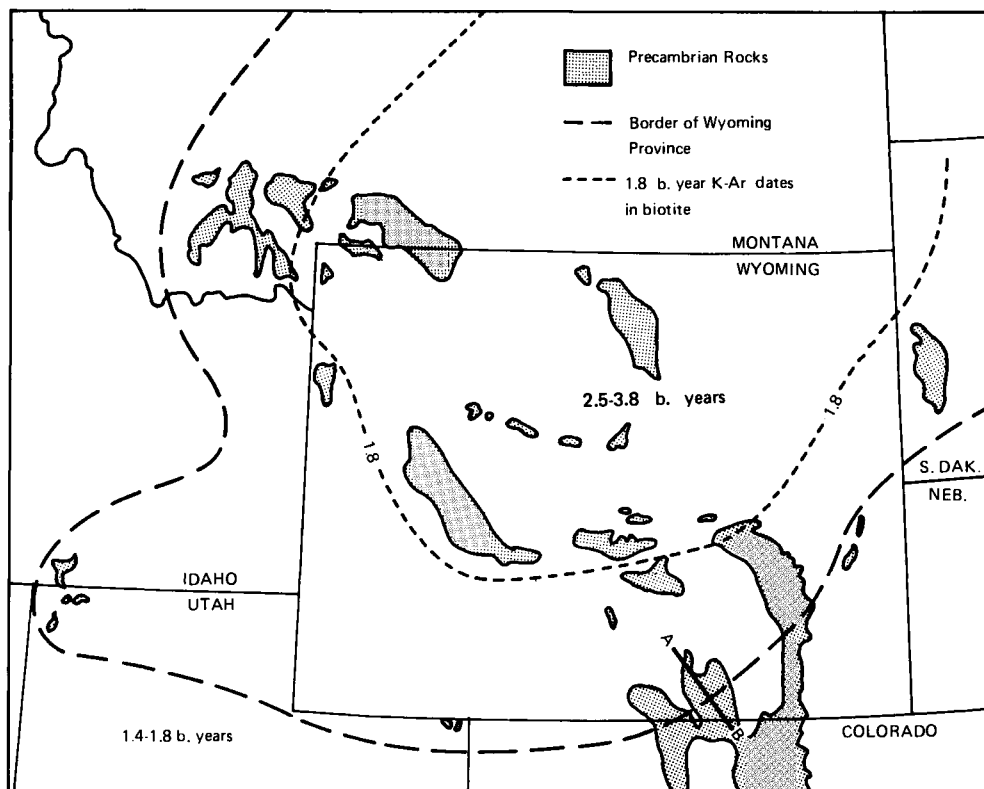


FIG 4.32 Map of the Wyoming Province showing the effect of later thermal heating on the outer parts of the province.

reworking of older crust in some areas, only the most intense and widespread orogenic events are used in delineating crustal provinces. Most of the provinces <2.5 Ga in age contain variable amounts of reworked older crust. Early Archean (≥ 3.5 Ga) crustal provinces, although widely distributed, are very small (chiefly <500 km across) and may represent remnants of the Earth's first continents (Chapt. 10). Late Archean provinces (2.5–3.5 Ga) are widespread on all continents and probably underlie much of the platform sediment, especially in Canada and Siberia. Early Proterozoic provinces (1.6–2.5 Ga) are widespread in North America and in the Baltic Shield in Europe. They are of lesser importance in South America, Africa, and Australia, and are of largely unknown extent beneath Asia and Antarctica. Mid-Proterozoic provinces (0.8–1.6 Ga) appear to be of minor extent, the largest province being the Grenville Province in North America. Late Proterozoic provinces (0.6–0.8 Ga) are of importance only in South America, Africa, and perhaps Antarctica.

Summary Statements

1. The Earth's crust is broadly divisible into continental (30–80 km), transitional (15–30 km) and oceanic (5–15 km) types. Continental plus transitional crust comprise 41% by area and 79% by volume of the total crust. The crust can be further divided into the 12 crustal types as summarized in table 4.1.
2. Seismic velocity results indicate that both oceanic and continental crust are

layered and velocities increase with depth. Oceanic crust is composed of sediment, basement and oceanic layers with increasing depth. Layer models for continental crust suggest variable numbers of layers. Crust in arc systems is particularly heterogeneous. The thickness of the lower continental crust appears to increase with age.

3. Seismic and electrical conductivity data indicate the common presence of one or more low-velocity, high conductivity layers in the middle or lower continental crust. These layers may be caused by the presence of free water, derived either from dehydration reactions or from the mantle.
4. Four types of seismic reflection fabrics are recognized in the continents: the craton, foreland deformation zone, and the inner and outer hinterlands. All of these fabrics may exhibit strong reflectors in the middle and lower crust. A correlation of depth to these reflections with heat flow suggests their depth varies over short times (10 Ma). Of the origins presented to explain these reflectors, only ductile strain banding or fluid-filled cracks seem capable of explaining their changing depths.
5. Heat is lost from the oceanic crust primarily at oceanic ridges, where losses by hydrothermal transport are important. Continental heat loss depends upon age of the last magmatic event, distribution of radiogenic heat sources with depth and erosion level. Approximately 70% of the Earth's heat loss is from oceanic ridges.
6. Surface heat flow from the continents is linearly related to crustal near-surface heat productivity. The zero intercept, the reduced heat flow, is the heat coming from the lower crust and mantle, and D , the slope (the characteristic depth) is the thickness of the upper layer enriched in radiogenic heat-producing elements. A model in which radiogenic heat sources decrease in abundance exponentially with depth best explains this relationship.
7. The average heat flow of the crust decreases with mean crustal age. In the continents, heat flow decreases to 40–50 mW/m² in about 1 Ga and to similar values in the oceans in 50–100 Ma. Reduced heat flow in the continents also decreases with mean crustal age.
8. Bouguer anomalies in continental areas reflect near-surface rock types and structural discontinuities, whereas in oceanic areas they reflect the distribution of hot, low-density upper mantle. Large negative anomalies over young orogenic areas reflect thicker crust and/or warmer, less dense upper mantle. Small isostatic anomalies in the crust indicate a close approach to isostatic equilibrium.
9. Magnetic anomalies in continental areas reflect near-surface rock types and structural discontinuities. Except in regions of thin crust and high heat flow, magnetic grain generally parallels orogenic structures. Oceanic crust is characterized by linear magnetic anomalies paralleling ocean ridge systems. Such anomalies are less distinct and more variable in marginal-sea and inland-sea basins.
10. Metamorphic mineral assemblages indicate that various levels of the continental crust are exposed at the Earth's surface. Successions of metamorphic facies (facies series) as well as pressure–temperature–time paths of crustal blocks are characteristic of specific tectonic settings.

11. The composition of the continental crust is constrained by seismic velocity distributions, the composition of Precambrian shields, the composition of shales, exposed crustal sections (both oceanic and continental), and the composition of crustal xenoliths in volcanic rocks.
12. Results indicate that the upper continental crust is lithologically and chemically heterogeneous and has an average composition of granodiorite. The lower crust is probably composed chiefly of mafic rocks, although some collisional zones may contain large amounts of felsic granulites. The average compositions of both the lower continental crust and the oceanic crust are basaltic and the average composition of continents is probably similar to andesite.
13. Crustal provinces are segments of continental crust that record a similar range of radiometric ages and exhibit similar deformational histories. They typically exhibit polyphase deformation and metamorphism.
14. Crustal province boundaries are faults or shear zones, unconformities, rapid changes in metamorphic grade, or intrusive contacts.
15. Orogenies are best dated with U–Pb zircon dates from syntectonic igneous rocks. Employing the different blocking temperatures of isotopes in various minerals, it is possible to monitor metamorphism and uplift following an orogeny, which may last from 50 to 100 Ma.

Suggestions for Further Reading

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CHAPTER 5

Plate Tectonics

Introduction

To understand the role of plate tectonics in the geologic history of the Earth, it is necessary to understand processes operating in modern plate regimes. A large amount of our information on modern plate tectonics comes from the fields of seismology and geomagnetism. Also, geological data retrieved by the Deep Sea Drilling Project provide major constraints on the interpretation of geophysical data. This chapter will review some of the important data, as well as interpretations that led to formation of the plate tectonics model for the Earth.

Geomagnetism

Rock Magnetization

To understand the magnetic evidence for seafloor spreading, it is necessary to understand how rocks become magnetized in the Earth's magnetic field. When a rock forms it may acquire a magnetization parallel to the ambient magnetic field, referred to as *primary magnetization*. The most important minerals controlling rock magnetization are magnetite and hematite. Information about both the direction and intensity of the magnetic field in which a rock formed can be obtained by studying its primary magnetization. However, it is not always easy to identify primary magnetization, in that rocks often acquire later magnetization known as *secondary magnetization*. Secondary magnetizations must be removed by demagnetization techniques prior to measuring primary magnetization. Magnetization measured in the laboratory is called *natural remanent magnetization* (NRM). Rocks may acquire NRM in several ways, of which only three are important in paleomagnetic studies.

(1) *Thermal remanent magnetization (TRM)*. TRM is acquired by igneous rocks as they cool through a blocking temperature(s) for magnetization of the constituent magnetic mineral(s). This temperature, known as the *Curie temperature*, ranges between 500°C and 600°C for iron oxides. The direction of TRM is almost always parallel and proportional in intensity to the applied field.

(2) *Detrital remanent magnetization (DRM)*. Clastic sediments generally contain small magnetic grains which became aligned in the ambient magnetic field as they fell through water during deposition or during compaction of the sediment. Such magnetization is known as DRM.

(3) *Chemical remanent magnetization (CRM)*. CRM is acquired by rocks during secondary processes if new magnetic minerals are formed. It may be produced during weathering, alteration, or metamorphism.

NRM is described by directional and intensity parameters. Directional parameters include *declination*, or the angle with respect to true north and *inclination*, or dip from the horizontal. A *paleomagnetic pole* can be calculated from the declination and inclination determined from a given rock.

Reversals in the Earth's Magnetic Field

Some rocks have acquired NRM in a direction opposite to that of the Earth's present magnetic field. Such magnetization is known as *reverse magnetization* in contrast to *normal magnetization* which parallels the present Earth's field. Experimental studies show that simultaneous crystallization of some Fe–Ti oxides with different Curie temperatures can cause these minerals to become magnetized with an opposite polarity to the ambient field. This *self-reversal magnetization* is related to ordering and disordering of Fe and Ti atoms in the crystal lattice. Although self-reversal has occurred in some young lava flows, it does not appear to be a major cause of reverse magnetization in rocks. The strongest evidence for this comes from correlation of reverse magnetization between different rock types from widely separated localities. For instance, reversed terrestrial lava flows correlate with reversed deep-sea sediments of the same age. It is clear that most reverse magnetization is acquired during periods of reverse polarity in the Earth's magnetic field.

One of the major discoveries in paleomagnetism is that stratigraphic successions of volcanic rocks and deep-sea sediment cores can be divided into sections that show dominantly reverse and normal magnetizations. *Polarity intervals* are defined as segments of time in which the magnetic field is dominantly reversed or normal. Using magnetic data from volcanic rocks and deep-sea sediments, a geomagnetic time scale was formulated (Cox, 1969), extending back to about 5 Ma (fig. 5.1). Although polarity intervals of short duration ($\leq 50,000$ years), cannot be resolved with K–Ar dating of volcanic rocks, they can be dated by other methods in deep-sea sediments, which contain a continuous (or nearly continuous) record of the Earth's magnetic history for the last 100–200 Ma. The last reversal in the magnetic field occurred about 20,000 years ago (the Laschamp).

Two types of polarity intervals are defined on the basis of their average duration: a *polarity event* or *subchron* (10^5 – 10^6 a) and a *polarity epoch* or *chron* (10^6 – 10^7 a). A polarity chron may contain several-to-many polarity subchrons and can be dominantly normal (e.g. the Brunhes), dominantly reversed (e.g. the Matuyama), or mixed. Larger intervals (10^7 – 10^8 a) with few if any reversals are known as *superchrons*. Based on the distribution of oceanic magnetic anomalies it is possible to extrapolate the geomagnetic time scale back more than 100 Ma. Independent testing of this extrapolation from dated basalts indicates the predicted time scale is correct to within a few percent back to at least 10 Ma. Results suggest that over the last 80 Ma the average length of polarity subchrons has decreased with time. Reversals in the Earth's field are documented throughout the Phanerozoic, although the geomagnetic time scale cannot be continuously extrapolated beyond about 200 Ma,

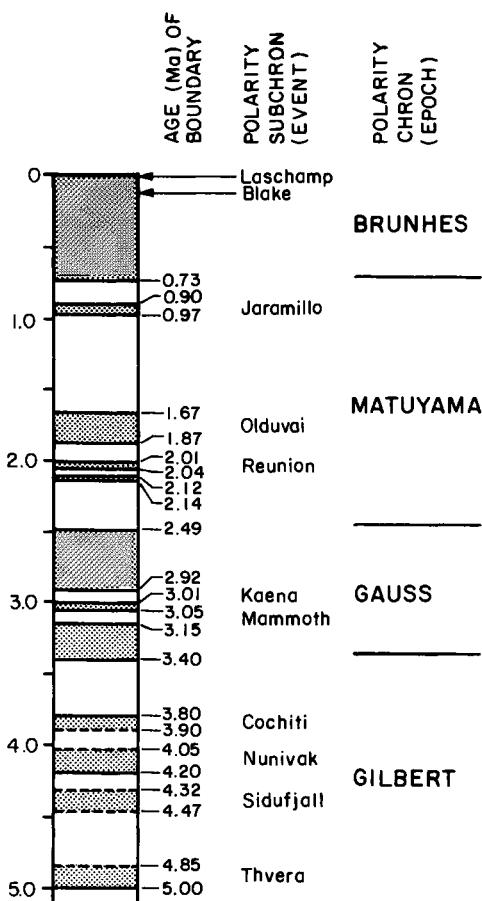


FIG 5.1 Late Cenozoic geomagnetic timescale based on data from terrestrial basalts and deep-sea sediments.

the age of the oldest oceanic crust. Reversals have been identified in rocks as old as 3.5 Ga.

The percentage of normal and reverse magnetization for any increment of time has also varied with time. The Mesozoic is characterized by dominantly normal polarities, while the Paleozoic is chiefly reversed (fig. 5.2). Periodic variations are suggested by the data at about 300, 110, and 60 Ma (Irving and Pullaiah, 1976). Statistical analysis of reversals in the magnetic field indicates a strong periodicity at about 30 Ma. Six superchrons are identified in the last 350 Ma, only two of which show relatively uniform polarities. These are the Cretaceous normal (CN) and Permian–Carboniferous reversed (PCR) superchrons (fig 5.2). Statistical analysis of the youngest and best-defined part of the geomagnetic time scale (<185 Ma) shows an almost linear decrease in the frequency of reversals back to the Cretaceous, reaching zero in the CN superchron. The inversion frequency appears to have reached a maximum about 10 Ma ago and has been declining to the present. Causes of changes in reversal frequency are generally attributed to changes in the relief or/and electrical conductivity along the core–mantle boundary. Both of these

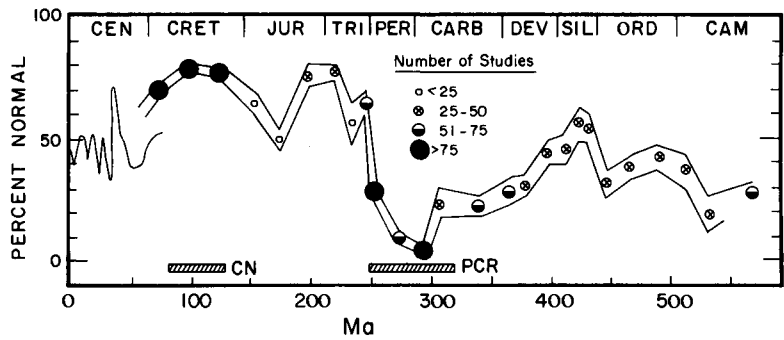


FIG 5.2 Distribution of Phanerozoic magnetic reversals averaged over 50 Ma periods. Also shown are the Cretaceous normal (CN) and Permian–Carboniferous reversed (PCR) superchrons. Modified after Piper (1987).

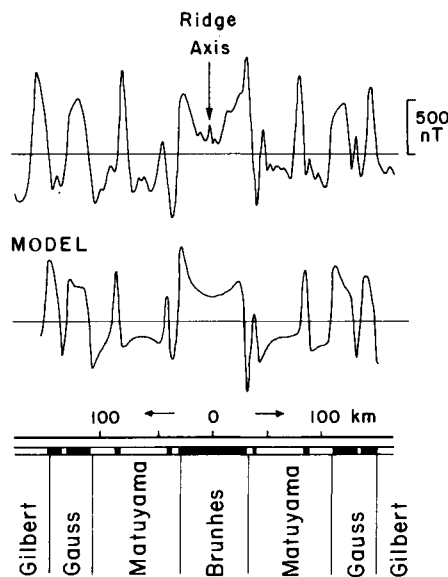


FIG 5.3 Observed and model magnetic profiles across the East Pacific rise at 51°S latitude and corresponding correlations with magnetic polarity intervals assuming a spreading rate of 4.4 cm/a. After Vine (1966), copyright by American Association for the Advancement of Science.

parameters are temperature-dependent and require long-term cyclical changes in the temperature at the base of the mantle. This in turn, implies that heat transfer from the lower mantle is episodic. A possible source of episodic heat loss from the core is latent heat released as the inner core grows by episodic overplating of Fe–Ni.

One of the most profound discoveries in terms of its impact on the geological sciences, as reported by Vine and Matthews (1963), is that linear magnetic anomaly patterns on the ocean floors correlate with reversed and normal polarity intervals in the geomagnetic time scale. This is caused by basalts injected into axial zones of ocean ridges during normal and reversed magnetization intervals. An example of such a correlation for the East Pacific rise is given in fig. 5.3. The suggested

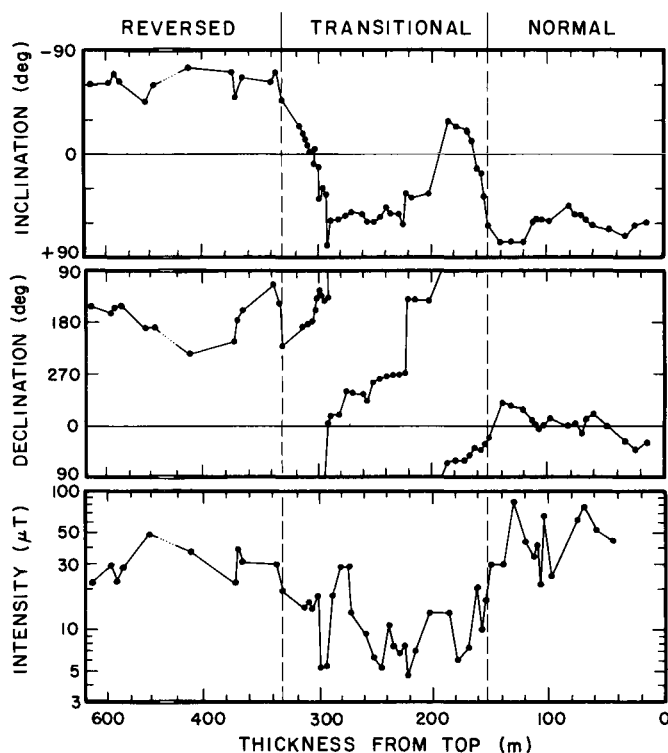


FIG 5.4 Record of a Mid-Tertiary magnetic reversal in successive basalt flows from the Steens Mountain area, Oregon. After Prevot et al. (1985).

correlations with polarity intervals are indicated at the bottom of the figure. A model profile for a half spreading rate of 4.4 cm/a is also shown, and is very similar to the observed profile. Both the Jaramillo and Olduvai events produce sizeable anomalies in the Matuyama chron. The Kaena and Mammoth events in the Gauss chron are not resolved, however. The lower limit of resolution of magnetic events in anomaly profiles with current methods is about 30,000 years. Although the distribution of magnetic anomalies seems to correlate well with polarity intervals, the amplitudes of anomalies can vary significantly between individual profiles. Such variation reflects, in part, inhomogeneous distribution of magnetite in basalts. Results suggest that most of the magnetization resides in the upper 0.5 km of basalts in the oceanic crust (Harrison, 1987). Less than 20% of the magnetization probably occurs below 0.5 km depth.

Several detailed studies across magnetic reversals in stratigraphic successions provide information on the timing and details of the magnetic field behaviour during a reversal. In those few sections where magnetic reversals are well dated, they occur over time intervals of 5000 to 10,000 years. One record for a section across Tertiary basalt flows in Oregon (~15 Ma in age) is shown in figure 5.4. In general, the field intensity is low during a reversal, and suggests that a decrease in the dipole field precedes a reversal (Prevot et al., 1985). Results from other geographic areas suggest that the average magnitude of the intensity during a transition is latitude-dependent. Notable also during the Steens transition are rapid and irregular changes in

inclination and declination. In sediment successions, where the record of magnetic transitions is often more complete, two scales of change in declination are resolved (Valet et al., 1986). Short-term changes ($\leq 10^2$ a), which typically show ≤ 10 degrees change in declination, increase in amplitude to 40–60 degrees during a reversal. This is generally interpreted to reflect an increase in turbulence in the outer core during a magnetic transition. Longer-term changes with a periodicity of 2000–4500 years are observed in both declination and inclination, and appear to be caused by variations in the non-dipole field. These are not affected by magnetic reversals.

The origin of magnetic reversals is poorly understood, and two general models have been proposed. In one, the main dipole field remains axial but decreases in strength during a reversal, so that the non-dipole field dominates during the transition. In the other, both dipole and non-dipole components decrease during the transition. Many theoretical models involving various changes in fluid motions in the outer core have been proposed to account for magnetic reversals. None of them, however, are well constrained or widely accepted.

Paleomagnetism

Paleomagnetism is the study of NRM in rocks, and has an important objective, the reconstruction of the direction and strength of the geomagnetic field over geologic time. To accomplish this goal, oriented samples are collected in the field, the NRM is measured in the laboratory with a magnetometer, and calculated paleopole positions are plotted on stereographic or equal-area projections (Piper, 1987). For the paleopole to have any significance in terms of continental drift, however, numerous problems must be addressed as follows.

(1) *Secular variation*. Variations in the Earth's magnetic field on timescales > 5 a are known as *secular variation*. Reliable observations on secular variation are available for over 300 a. Such variations include a westward drift of the magnetic field of about $0.2^\circ/\text{a}$, a decay in the dipole component of about 7% since 1845, and drift of the magnetic poles. Using dated archeological remains it is possible to estimate the trajectory of the North Magnetic Pole for several thousand years. Results show a path around and within a few degrees of the rotational pole such that the average position is equal to the rotational pole. This line of evidence suggests that the magnetic and rotational poles have been coincident in the geologic past.

(2) *Dipole field*. The present Earth's field can be approximated as geocentric dipole inclined at 11.5° to the rotational axis. The *non-dipole field*, which remains after subtraction of the dipole field from the measured field, is about 5% of the total field. In reconstructing ancient pole positions a dipole field is assumed to have existed in the geologic past. The consistency of pole positions from different parts of a continent at the same geologic time, and the world-wide agreement of paleoclimatic regimes with latitudes calculated from paleomagnetic studies, supports this assumption. Because of the axial symmetry of the dipole field, however, paleolongitudes cannot be determined from paleomagnetic data.

(3) *True polar wander*. Ancient pole positions are generally illustrated by plotting an *apparent polar wandering (APW) path* for a continent (or segment thereof) relative to its present position (fig. 5.5). It is referred to as an apparent path because it may

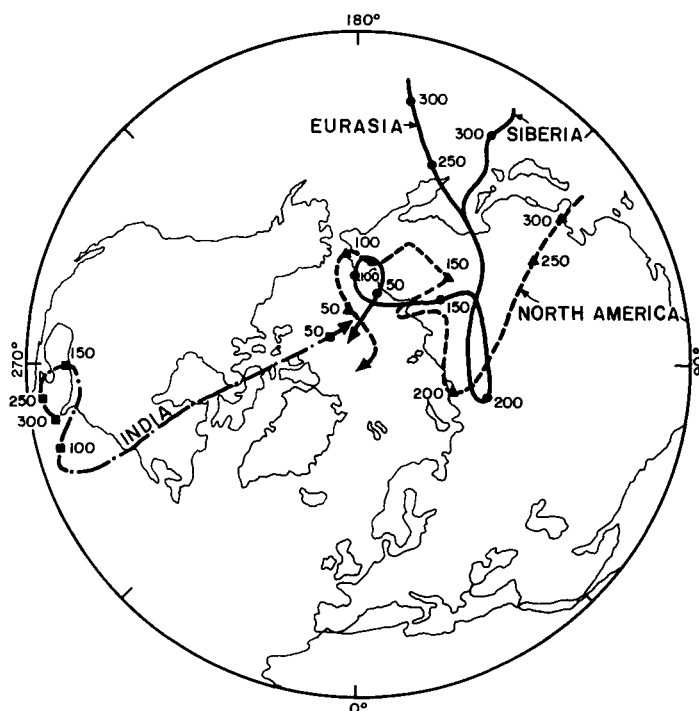


FIG 5.5 Apparent polar wandering paths of North America, Eurasia, Siberia, and India for the last 300 Ma. Ages given in Ma ago.

result from either plate motions or *true polar wander* (TPW), or both. True polar wander results from movement of the entire lithosphere with respect to the mantle or of the mantle relative to the core. To rigorously separate the two effects requires one of three approaches: (1) evaluation of the net motion of the entire lithosphere relative to the Earth's spin axis (the vector-sum method), (2) the mean-lithosphere method based on separating random plate motions from the mean motion of the lithosphere, and (3) the hotspot method, based on comparing plate motions to hotspots that are assumed to be fixed relative to the paleomagnetic axis (Gordon, 1987). Results of these studies suggest, but do not yet necessitate, that TPW has an affect on APW paths. Part of the problem is the uncertainty in the amount of motion between hotspots. Although the contribution of TPW is unknown, the fact that it affects equally all calculated poles of a given age indicates that it does not influence the *relative* plate motions deduced from APW paths. However, for a single supercontinent or for two continents with similar APW paths, the effects of TPW and plate motion cannot be distinguished.

(4) *Normal or reversed pole.* For an isolated paleopole it is not possible to tell if the pole is reversed or normal. This decision must be based on other information such as paleoclimatic data or an APW path constructed from other geographic localities but including the age of the measured sample.

(5) *Deformation.* The most reliable pole positions come from nearly flat-lying rock units. In rocks that acquired NRM prior to deformation it may be possible to estimate pole positions by "unfolding" the rocks. Strain associated with folding also must be removed in that it affects NRM. It is only in strata that have undergone

rather simple folding that these structural corrections can be made. However, in complexly deformed metamorphic terranes, $^{40}\text{Ar}/^{39}\text{Ar}$ dating indicates that major NRM is usually acquired after deformation, possibly during regional uplift. Thus, in many instances it is possible to obtain meaningful post-deformational paleopoles from strongly deformed and metamorphosed rocks.

(6) *Dating of NRM.* The Curie temperatures of major magnetic minerals (Fe–Ti oxides) lie between 500°C and 600°C. Rocks become magnetized when they are cooled through their Curie temperatures, and it is possible to date the magnetization using an isotopic system with a blocking temperature similar to the Curie temperatures. The $^{40}\text{Ar}/^{39}\text{Ar}$ method has proved useful in this respect, generally using biotite or hornblende. Extreme caution should be used, however, in assigning $^{40}\text{Ar}/^{39}\text{Ar}$ dates to rocks that have multiple magnetizations.

(7) *Separation of remanence components.* One of the most difficult problems in paleomagnetic studies is that of separating, identifying, and establishing the sequence of acquisition of multiple magnetizations. This is particularly a problem in metamorphic rocks where TRM or DRM may be overprinted with one or more CRMs of metamorphic origin. Standard techniques for separating magnetic components include stepwise treatment of samples either by alternating magnetic fields or by heating. Depending on the history of a rock, it may be simple, difficult or even impossible to separate and date the various magnetic components.

In the paleomagnetic method, APW paths of continents (or segments thereof) are plotted on equal-area projections of the Earth generally relative to the present positions of continents. If continents show the same APW paths it is clear they did not drift relative to each other. The fact that paths diverge from each other going back in time, however, indicates that continents have drifted separately (fig. 5.5). The pole may or may not have wandered over the same period of time. It is noteworthy that Europe and Siberia drifted independently until the Triassic, when their APW paths converged. This reflects collision and suturing of these plates in the Triassic (see Chapt. 8). To illustrate the use of APW paths in reconstructing plate motions, four tectonic scenarios with corresponding APW paths are shown in figure 5.6. The simplest case is for a single continent in which paleopole positions from various cratons lie on the same APW path (case A). In this case the cratons have been part of the same continent over the period of time represented by the APW path. Such paths characterize most Precambrian cratons within a given continent for the last 600 Ma, central and eastern Asia being the major exception. Case B illustrates the aggregation of a supercontinent by successive collisions of continents. Two continents collide at time 4 and are joined by two more continents at times 8 and 10 to form a supercontinent. A *Wilson cycle*, or the opening and closing of an ocean basin, is illustrated in case C. At time 4 a continent is rifted apart and a major ocean basin opens, reaching a maximum width at time 8, when the two continental fragments begin to close again. The original continent is reformed at time 12 by collision of these fragments. The offset in APW paths from times 1–4 reflects the fact that the fragments did not return to their original position. The opening and closing of the Iapetus Ocean during the Paleozoic (Chapt. 8) is reflected by the APW paths of continents surrounding the Atlantic. Case D illustrates offset along a major continental transform fault where motion begins at time 8.

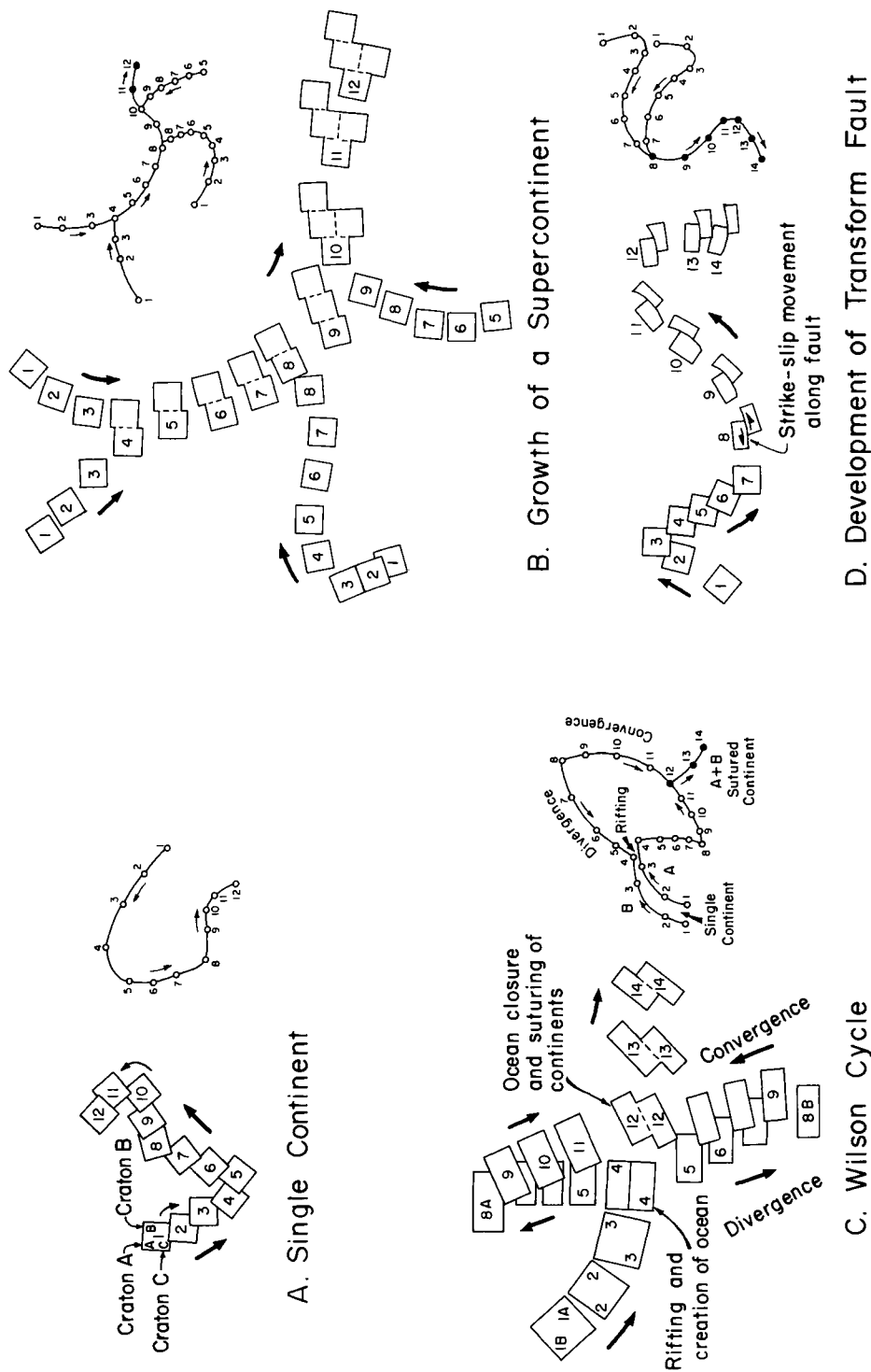


FIG 5.6 Apparent polar wandering paths for various tectonic scenarios. Modified after Piper (1987). Numbers represent relative time (1 being oldest).

The paleomagnetic method has the unique property of being able to see through the break-up Pangea 200 Ma ago. In principle it is possible to reconstruct continental positions prior to 200 Ma provided the problems listed above can be overcome. Results from paleomagnetic studies, as well as from geological studies, indicate that continents have grown by successive microcontinent or arc collisions, and that they may be fragmented and pulled apart by rifting, leading to the production of new ocean basins.

Plate Boundaries

Introduction

Earthquakes occur along rather narrow belts on the Earth today (fig. 5.7), and these belts mark boundaries between lithospheric plates. There are four types of seismic boundaries, distinguished by their epicenter distributions and geologic characteristics: (a) ocean ridges, (b) subduction zones, (c) transform faults, and (d) collisional zones. Provided a sufficient number of seismic recording stations with proper azimuthal locations are available, it is possible to determine the directions of first motion at sites of earthquake generation. Such investigations are known as *first-motion studies* or *fault-plane solutions*. First-motion studies provide major constraints on plate motions.

Modern plates range in size from $<10^4$ km² to over 10^8 km², and plate margins do not usually coincide with continental margins (Plate I). Seven major plates are recognized: the Eurasian, Antarctic, North American, South American, Pacific, African, and Australian plates. Intermediate-size plates (10^6 – 10^7 km²) include the Philippine, Arabian, Nazca, Cocos, Caribbean, and Scotia plates. In addition there appear to be more than 20 plates with areas of 10^5 – 10^6 km². Both plate theory and first-motion studies at plate boundaries indicate that plates are created at ocean ridges, consumed at subduction zones, and slide by each other along transform faults. At collisional zones, plates carrying continents may become sutured together. Plates diminish or grow in area depending on the distribution of convergent and divergent boundaries. The African plate, for instance, with exception of the northern boundary, is entirely surrounded by active spreading centers and is growing. If the surface area of the Earth is to be conserved, other plates must be diminishing in area as the African plate grows, and this is the case for plates in the Pacific area. Plate boundaries are dynamic features, not only migrating about the Earth's surface but changing from one type of boundary to another. In addition, new plate boundaries can be created in response to changes in stress regimes in the lithosphere. Also, plate boundaries can become extinct as two plates become part of the same plate. Small plates ($\leq 10^6$ km²) occur most frequently near continent–continent or arc–continent collisional boundaries and are characterized by rapid, complex motions. Examples are the Turkish–Aegean, Adriatic, Arabian, and Iran plates located along the Eurasian–African continent–continent collision boundary, and several small plates along the continent–arc collision border of the Australian–Pacific plates. The motions of small plates are controlled largely by the compressive forces of large plates.

Continental margins are of two types, active and passive. An *active continental margin* is found where either a subduction zone or a transform fault coincides with

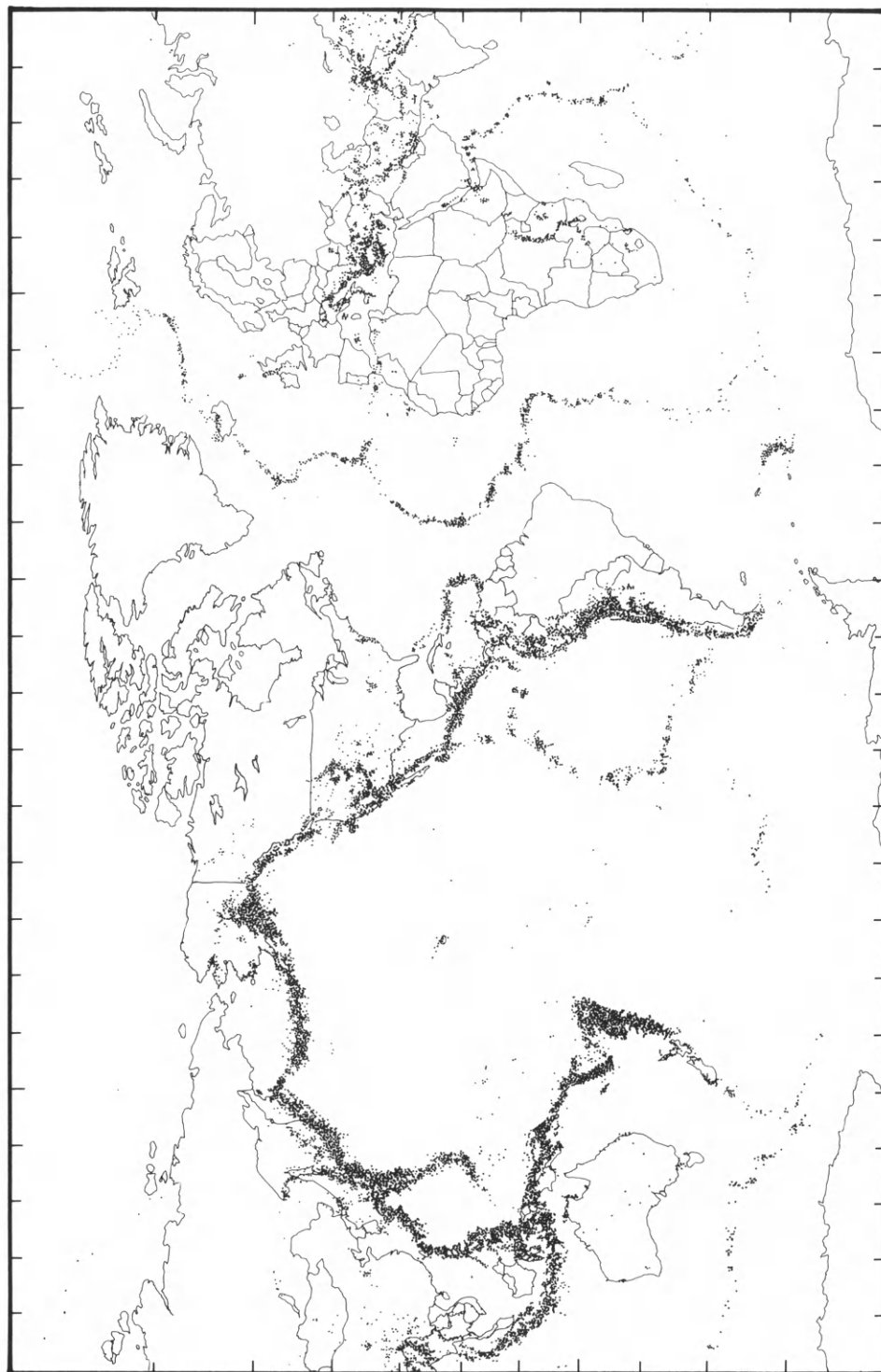


FIG 5.7 Distribution of world earthquakes 1961–1969 (from National Earthquake Information Center NEIC-3005).

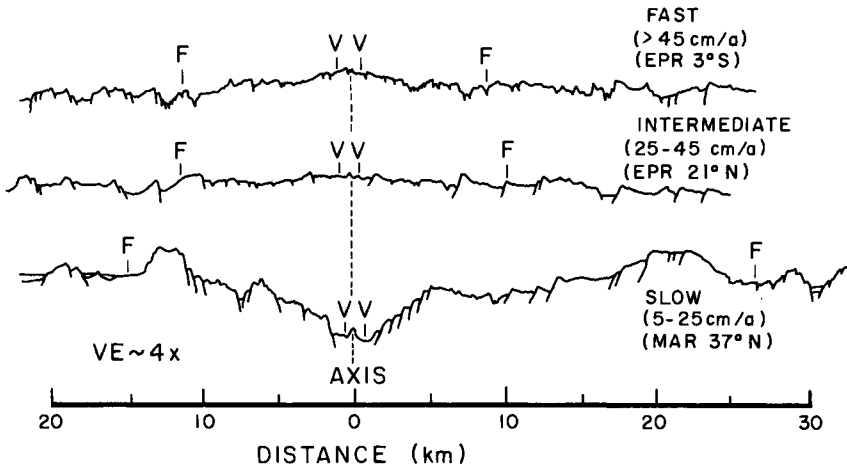


FIG 5.8 Axial topographic profiles across three ocean ridges. EPR, East Pacific rise; MAR, Mid-Atlantic ridge. V and F indicate widths of active volcanism and active faulting, respectively. After MacDonald (1982).

continent–ocean interface. Examples are the Andean and Japan continental-margin arc systems and the San Andreas fault in California. *Passive continental margins* occur along the edges of opening ocean basins like the Atlantic basin. Such margins are characterized by minimal tectonic and igneous activity.

Divergent Boundaries (Ocean Ridges)

Ocean ridges are characterized by shallow earthquakes limited to axial rift zones. These earthquakes are generally small in magnitude, commonly occur in swarms, and appear to be associated with intrusion and extrusion of basaltic magmas. First-motion studies indicate that rift earthquakes are produced dominantly by vertical faulting, as is expected if new lithosphere is being injected upwards. Most faulting occurs in the depth range of 2–8 km and some ruptures extend to the seafloor. The median valley of ocean ridges varies in geological character due to the changing importance of tectonic extension and volcanism. In the northern part of the Mid-Atlantic ridge, stretching and thinning of the crust dominate in one section while volcanism dominates in another. Where tectonic thinning is important, faulting has exposed gabbros and serpentinites from deeper crustal levels. Volcanic features range from large ridges (≥ 50 km long) in sections of the median valley in which volcanism has dominated, to small volcanic cones in sections dominated by extension. The axial topography of fast- and slow-spreading ridges varies considerably. A deep axial valley with flanking mountains characterizes slow-spreading ridges, while relatively low relief, and in some instances a topographic high, characterizes fast-spreading ridges (fig. 5.8). Model studies suggest that differences in horizontal stresses in the oceanic lithosphere may account for the relationship between ridge topography and spreading rate (Morgan et al., 1987). In oceanic lithosphere, which thickens with distance from a ridge axis, horizontal extensional stresses can produce the axial topography found on slow-spreading ridges. In fast-spreading ridges, however, the calculated stresses are too small to result in appreciable topography.

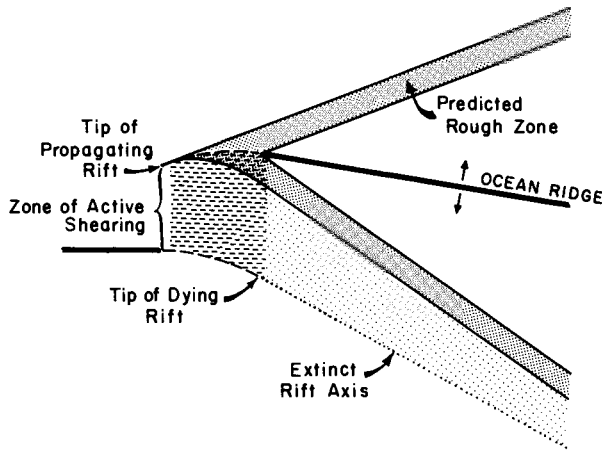


FIG 5.9 Crustal patterns produced by propagating and dying oceanic rifts. After Hey et al. (1980).

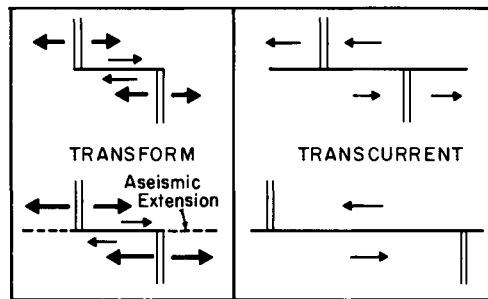


FIG 5.10 Motion on transform and transcurrent faults relative to an ocean ridge axis (double vertical lines). Note that offset increases with transcurrent motion while it remains constant with transform motion. Bold arrows refer to spreading directions, small arrows to plate motions.

Evidence suggests that ocean ridges grow and die out by lateral propagation. One model for propagation of a medial rift zone is illustrated in fig. 5.9. The model assumes that the spreading rate of the dying rift decreases to zero while the rate on the growing rift increases from zero to full rate over approximately the same distance. The change from fast to slow spreading produces a zone of rough topography which forms a "V", with the apex at the tip of the growing rift. A zone of intense shearing is produced in the crust between the tip of the propagating and the tip of the dying rift. Offset magnetic anomalies and bathymetry consistent with propagating rifts, with and without transform faults, have been described along the Galapagos ridge and in the Juan de Fuca plate (Hey et al., 1980).

Transform Faults and Fracture Zones

Transform faults differ from transcurrent faults in that the sense of motion relative to offset along ocean ridge axes is opposite to that predicted by transcurrent motion (Wilson, 1965) (fig. 5.10). It is clear that the motion observed along transform faults

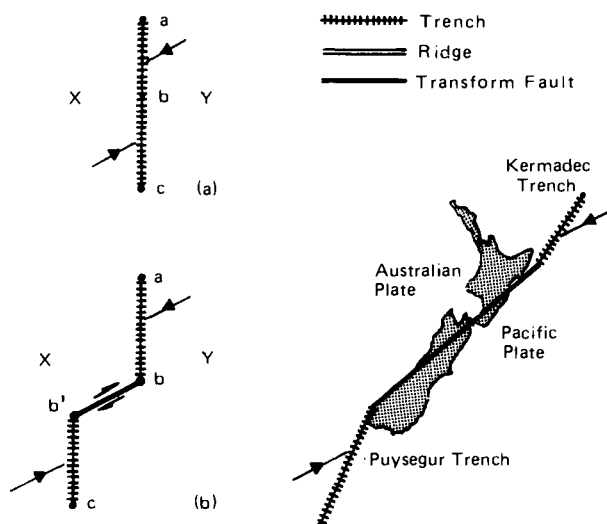


FIG 5.11 Origin of the trench-trench Alpine transform fault in New Zealand. After McKenzie and Morgan, 1969.

cannot have produced the offsets of ridge axes. These offsets may have developed at the time spreading began, and reflect inhomogeneous fracturing of the lithosphere. Transform faults, like ocean ridges, are characterized by shallow earthquakes (<50 km deep). Both geophysical and petrological data suggest that most oceanic transforms are “leaky” in that magma is injected along fault surfaces producing strips of new lithosphere (Garfunkel, 1986). Geological observations in transform shear zones exposed in ophiolites support this conclusion. Transforms cross continental or oceanic crust and may show apparent lateral displacements of many hundreds of kilometers. First-motion studies of oceanic transform faults indicate strike-slip motion in a direction away from ocean ridges (fig. 5.10). Also, as predicted by seafloor spreading, earthquakes are restricted to areas between offset ridge axes. Transform faults may produce large structural discontinuities, and in some cases structural and topographic breaks known as fracture zones are left on the ocean floor. *Fracture zones* include ridge-ridge transform faults and their aseismic extensions (Plate I) and the transform faults may or may not be active.

There are three types of transform faults: ridge-ridge, ridge-trench, and trench-trench faults. Ridge-ridge transform faults are the most common type, and many are shown in Plate I. Most ridge-ridge transforms retain a constant length as a function of time (fig. 5.10), whereas ridge-trench and trench-trench transforms decrease or increase in length as they evolve. The Alpine transform fault in New Zealand is an example of how a trench-trench transform may increase in strike length (fig. 5.11). In this figure the upper trench segment a-b (destined to become the Kermadec trench) consumes only plate Y (the Pacific plate) and the lower trench b-c (destined to become the Puysegur trench) consumes only plate X (the Australian plate). As Y is consumed between a and b, but not between b and c, b-c must be steadily offset from a-b to form two trenches joined by a transform fault.

Because of the time-dependent thermal behavior of oceanic lithosphere, transform

faults must have dip-slip as well as strike-slip components. Studies of oceanic transform-fault topography and structure indicate that zones of maximum displacement are very localized (<1 km wide) and are characterized by an anastomosing network of faults. Steep transform valley walls are composed of inward-facing scarps associated with normal faulting. Large continental transform faults form where pieces of continental lithosphere are squeezed within intracontinental convergence zones such as the Anatolian fault in Turkey. Large earthquakes ($M \geq 8$) separated by long periods of quiescence occur along “locked” segments of continental transforms, whereas intermediate-magnitude earthquakes characterize fault segments in which episodic slippage releases stresses. Aseismic slip occurs along well-lubricated segments of transforms. Large earthquakes along continental transforms appear to have a period of about 150 years.

Ridge segments between oceanic transforms behave independently of each other. This may be caused by instability in the convective upcurrents that feed ocean ridges causing these upcurrents to segment into regularly spaced rising diapirs, with each diapir feeding a different ridge segment. Transforms may arise at the junctions of ridge segments because magma supply between diapirs is inadequate for normal oceanic crustal accretion. The persistence of transforms over millions of years indicates that asthenospheric diapirs retain their integrity for long periods of time. Employing fixed hotspot models of absolute plate motion, it appears that both ridge axes and transforms migrate together at a rate of a few centimeters per year. This, in turn, requires that mantle diapirs migrate, and suggests that ocean ridge segments and diapirs are decoupled from underlying mantle flow.

Triple Junctions

Triple junctions are points where three plates meet. Such junctions are a necessary consequence of rigid plates on a sphere, since this is the common way a plate boundary can end. There are 16 possible combinations of ridge, trench, and transform-fault triple junctions (McKenzie and Morgan, 1969), of which only six are common. These are illustrated in figure 5.12, with examples. Triple junctions are classified as stable or unstable, depending on whether or not they preserve their geometry as they evolve. The geometric conditions for stability are described with vector velocity triangles in the figure, and only RRR triple junctions are stable for all orientations of plate boundaries. It is important to understand evolutionary changes in triple junctions, because changes in their configuration can produce changes that superficially resemble changes in plate motions. Triple junction evolution is controlled by the lengths of transform faults, spreading velocities, and the availability of magma.

Convergent Boundaries (Subduction Zones)

Convergent plate boundaries are defined by earthquake hypocenters that lie in an approximate plane and dip beneath arc systems. This plane, known as the *seismic zone* or *Benioff zone*, dips at moderate to steep angles and extends in some instances to the 670 km seismic discontinuity. The Benioff zone is interpreted as a brittle region in the upper 10–20 km of descending lithospheric slabs. Modern Benioff

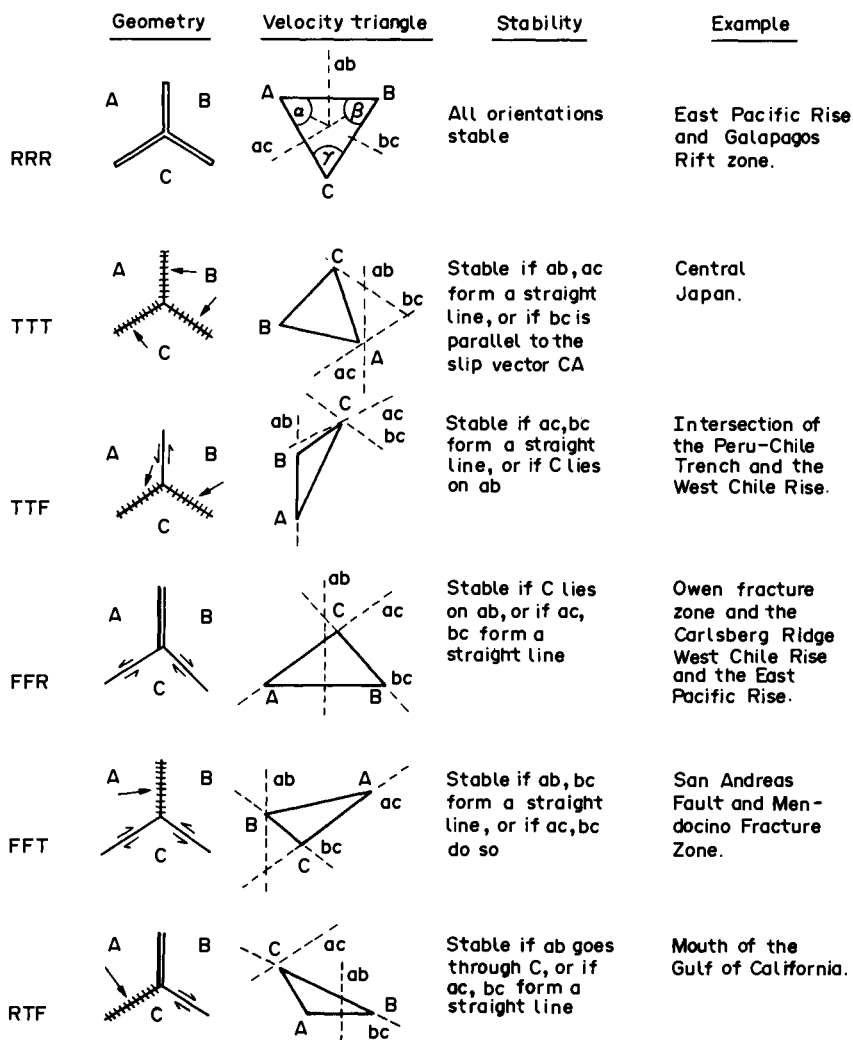


FIG 5.12 Geometry and stability requirements of six common triple junctions. Modified after McKenzie and Morgan (1969). Dashed lines ab, bc, and ac in the velocity triangles join points, the vector sum of which leave the geometry of AB, BC, and AC, respectively, unchanged. The junctions are stable only if ab, bc, and ac meet at a point.

Key: ———, trench; =====, ridge; ———>———, transform fault.

zones vary significantly in hypocenter distribution and in dip (fig. 5.13). Some, such as the seismic zone beneath the Aleutian Islands, extend only to depths <300 km, while others extend to the 670 km discontinuity (fig. 5.13 a and b, respectively). In general, seismic zones are curved surfaces with radii of curvature of several hundred kilometers and with irregularities on scales of <100 km. Approximately planar seismic zones are exceptional. Seismic gaps in some zones (for example c and e) suggest, although do not prove, fragmentation of the descending slab. Dips range from 30° to 90°, averaging about 45°. Considerable variation may occur along strike in a given subduction zone, as exemplified by the Izu-Bonin arc system in the

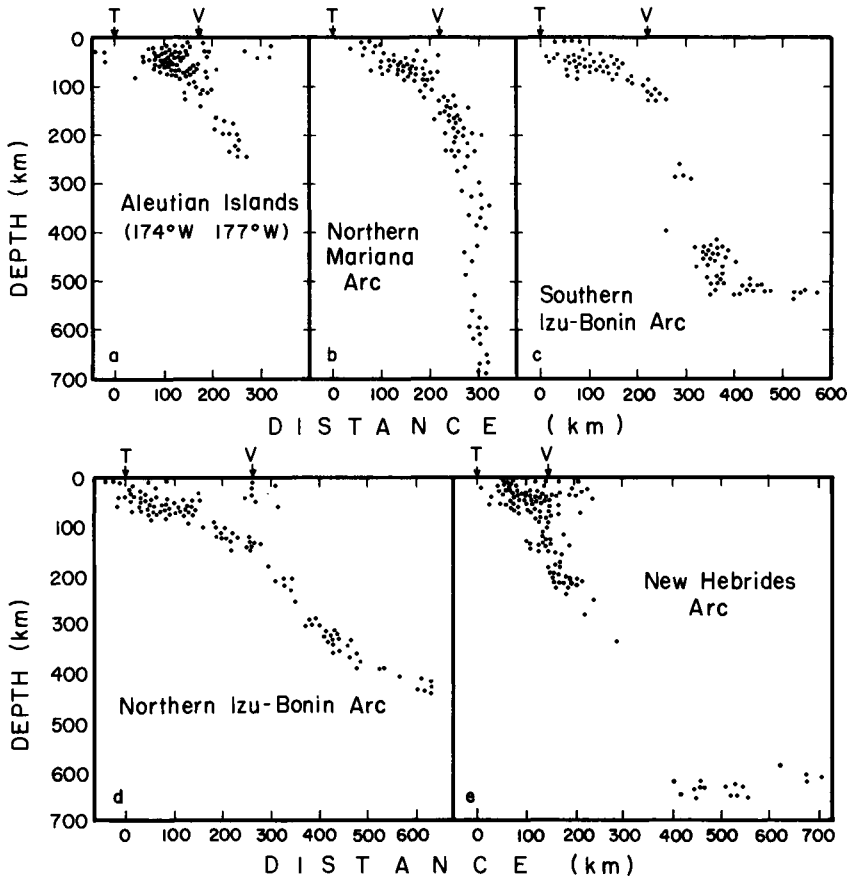


FIG 5.13 Vertical cross-sections of hypocenter distributions beneath modern arc-trench systems. Each diagram shows earthquakes for 7–10-year periods between 1954 and 1969. T = trench axis; V = recently active volcanic chain. Distance is measured horizontally from each trench axis in kilometers. Hypocenter data from many sources, and principally from National Earthquake Information Center, U.S. Coast and Geodetic Survey.

western Pacific. Hypocenters are linear and rather continuous on the northern end of this arc system (fig. 5.13d), becoming progressively more discontinuous toward the south. Near the southern end of the arc the seismic zone exhibits a pronounced gap between 150 and 400 km depth (fig. 5.13c). An excellent correlation exists between the length of seismic zones and the product of plate convergence rate and age of the downgoing slab.

First-motion studies of earthquakes in subduction zones indicate variation as a function both of lateral distance along a subduction zone and of depth. Seaward from the trench in the upper part of the lithosphere, where the plate begins to bend, shallow extensional mechanisms predominate. Because of their low strength, sediments in oceanic trenches cannot transmit stresses, and hence they are usually flat-lying and undeformed. Seismic reflection profiles indicate, however, that rocks on the landward side of trenches are intensely folded and faulted. Thrusting

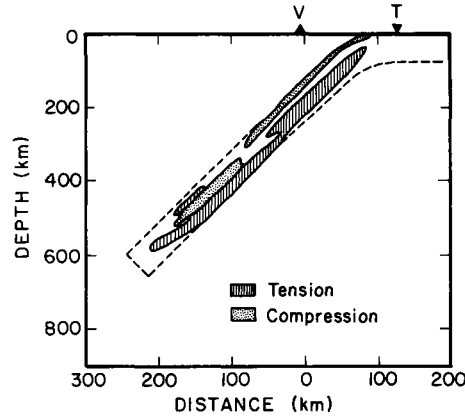


FIG 5.14 Calculated distribution of stresses in a descending slab. Subduction rate, 8 cm/a and dip = 45°. After Goto et al. (1985). V, volcanic front; T, trench.

mechanisms dominate at shallow depths in subduction zones (20–100 km). At intermediate depths (100–300 km), compressional or extensional motions dominate, while at depths >300 km, compressional motions are characteristic. At depths <25 km, descending slabs are characterized by low seismicity. Large-magnitude earthquakes are generally thrust-types and occur at depths >30 km (Shimamoto, 1985). During large earthquakes, ruptures branch off the slab and extend upwards, forming thrusts that dip away from the trench axis. Recent calculations of stress distributions at <300 km depth show that compressional stresses generally dominate in the upper parts of descending slabs while tensional stresses are more important in the central parts (Goto et al., 1985) (fig. 5.14). At 300–350 km depth compressional stresses are very small, and at depths >400 km a region of compressional stress is bounded both below and above by tensional stress regions.

The seismicity in descending slabs is strongly correlated with the degree of coupling between the slab and the overriding plate (Shimamoto, 1985). The low seismicity in descending slabs at depths <25 km may reflect relatively high water contents and the low strength of subducted clays, both of which lead to decoupling of the plates and largely ductile deformation. At greater depths, diminishing water and clay contents (due to slab devolatilization) result in greater coupling of overriding and descending slabs. In continental-margin subduction zones (e.g., Peru–Chile, Alaska), coupling is very strong resulting in large earthquakes and a relatively low dip of the descending slab (fig. 5.15). In other areas, such as the Kurile and Mariana arc systems (fig. 5.13), slabs are largely decoupled from overriding plates and extend to great depths, and earthquakes are smaller and less frequent. In some instances, subducting slabs may change dip and travel beneath the lithosphere of the overriding plate (fig. 5.15). The probable explanation for this geometry is that the subducting plate is relatively buoyant (Sacks, 1983). As the oceanic lithosphere thickens with age (fig. 3.15), it cools and becomes more dense and finally becomes negatively buoyant and descends into the asthenosphere. If the lithosphere is subducted before this time, which corresponds to about 50 Ma in age, it will tend to resist subduction and ride beneath the lithosphere of the overriding plate. It is not until the basalt–eclogite transition occurs, which is retarded due to relatively low

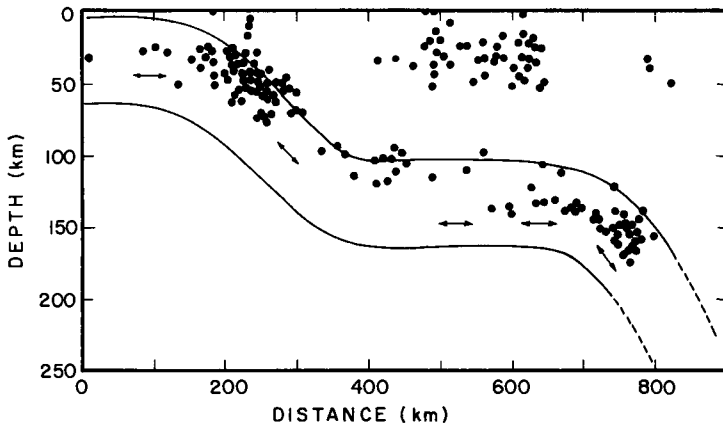


FIG 5.15 The Peru–Chile subduction zone in central Peru. Earthquake first motions shown by arrows are tensional. Modified after Sacks (1983).

temperatures, that the slab finally descends into the mantle. The large gap in hypocenters (from 200 to 600 km) in the central Peru slab may indicate that the tip of the slab broke off and settled to the 670-km discontinuity. A similar break-off may be recorded in the New Hebrides slab (fig. 5.13e). Because slabs appear to penetrate the 670-km discontinuity (Chapt. 3), the lack of earthquakes below 700 km probably reflects the depth of the brittle–ductile transition in descending slabs.

Q (the specific attenuation factor) has a rather complex distribution at subduction zones. A zone of extremely low seismicity and $Q(\leq 150)$ exists above the Tonga subduction zone and correlates with high surface heat flow in the overlying active marginal basin (the Lau–Havre trough, see fig. 4.2). A similar region occurs in the upper mantle beneath the Sea of Japan. Unlike the Tonga region, the upper mantle above the Peru–Chile subduction zone exhibits high to moderate seismicity and high $Q(1000\text{--}3000)$ (James, 1971) suggesting a thickened lithosphere. It is possible that partially molten mantle diapirs account for the low- Q regions above some descending slabs. The relatively thick lithosphere beneath the central Andes (~ 350 km), on the other hand, may characterize regions in which plates carrying continental crust are overriding plates carrying oceanic crust.

Collisional Boundaries

Deformational zones associated with collisional boundaries are widespread, as exemplified by the India–Asia boundary which extends for at least 3000 km northeast of the Himalayas (see fig. 6.24) (Molnar and Tapponier, 1975). Earthquakes are chiefly <100 km deep and first-motion studies indicate a variety of fault types. Thrust fault mechanisms generally dominate near the suture zone, such as the Indus suture in the Himalayas. Strike-slip faulting is common in the overriding plate, as illustrated by the large strike-slip faults produced in China and Tibet during the India collision. In addition, extensional faulting may extend great distances beyond the suture in the overriding plate. For instance, the Baikal rift in southern Siberia appears to have formed in response to the India collision 45 Ma ago.

A plate boundary in the early stages of an arc–continent collision is illustrated by

the Banda arc system in eastern Indonesia. Australia is beginning to collide with this arc as the Australian plate is subducted beneath the arc (Plate I). A large bend in the descending slab beneath Timor may be produced by subduction of continental crust. Numerous hypocenters at 50–100 km depth are interpreted to reflect the beginning of detachment of the descending slab as continental crust resists further subduction. Farther to the east, in New Guinea, Australia collided with the arc system earlier and has since become accreted to New Guinea. Earthquakes and active volcanism in northern New Guinea are generally interpreted to reflect initiation of a new subduction zone dipping to the south beneath New Guinea (in the opposite direction of subduction prior to collision).

Stress Distribution within Plates

Stress distributions in the lithosphere can be estimated from geological observations, first-motion studies of earthquakes, and direct measurement of *in situ* stress (Zoback and Zoback, 1980). Results indicate the existence of *stress provinces* that show similar stress orientations and earthquake magnitudes, and have linear dimensions ranging from hundreds to thousands of kilometers. Maximum compressive stresses for much of North America trend E–W to NE–SW and E–W to NW–SE for cratonic regions of South America. Western Europe is characterized by dominantly NW–SE compressive stresses, while much of Asia is more nearly N–S. Within the Indian plate compressive stresses range from N–S in India to nearly E–W in Australia. Horizontal stresses are variable in Africa but suggest a NW–SE trend for maximum compressive stresses in West Africa and an E–W trend for the minimum compressive stress in East Africa. Oceanic lithosphere away from plate boundaries is generally characterized by variable deviatoric compressive stresses. The overall pattern of stresses in both the continental and oceanic lithosphere is consistent with the present distribution of plate motions as deduced from magnetic anomaly distributions on the seafloor.

Examples of stress provinces in the United States are shown in figure 5.16. Most of the central and eastern parts of the United States are characterized by compressive stresses, ranging from NW–SE along the Atlantic Coast to dominantly NE–SW in the midcontinent area. Much of the western United States is characterized by extensional and strike-slip stress patterns, although the Colorado Plateau, the Pacific Northwest, and the area around the San Andreas fault are dominated by compressive deformation. The abrupt transitions between stress provinces in the western United States imply a crustal or uppermost mantle source for the stresses. The correlation of stress distribution and heat-flow distribution in this area indicates that widespread rifting in the Basin and Range Province (and other rift areas) is linked to thermal processes in the mantle. The broad transitions between stress provinces in the central and eastern United States reflect deeper stresses at the base of the lithosphere related to drag resistance and compressive forces transmitted from the Mid-Atlantic ridge.

Plate Motions

The motion of a plate on a sphere can be described in terms of a *pole of rotation* passing through the center of the sphere. Because all plates on the Earth are moving

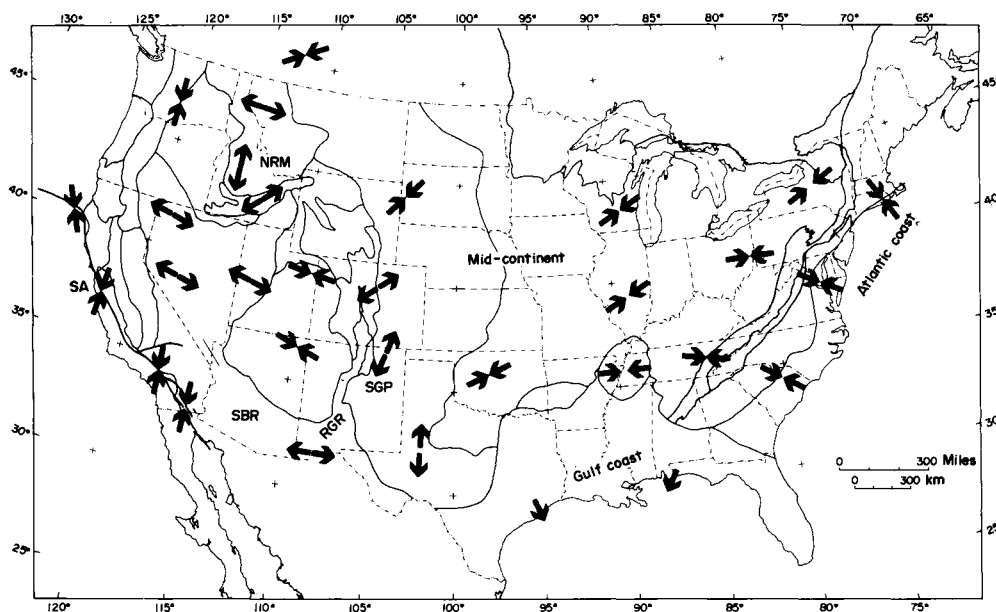


FIG 5.16 Stress provinces in the United States (from Zoback and Zoback, 1980). Arrows represent directions of least (outward-directed) or greatest (inward-directed) principal horizontal compression. Province abbreviations: PNW, Pacific Northwest; SA, San Andreas; SN, Sierra Nevada; NBR, SBR, northern and southern Basin and Range Province; RGR, Rio Grande rift; CP, Colorado Plateau; SR, Snake River Plain; NRM, northern Rocky Mountains; SGP, southern Great Plains.

relative to each other, the angular difference between a given point on a plate to the pole of rotation of that plate generally changes with time. For this reason the trajectory of a point on one plate as observed from another cannot be described by a small circle around a fixed pole of rotation. Instead, the shape of a relative motion path is that of a spherical cycloid (Cronin, 1987). The trajectory of a point on a moving plate relative to a point on another plate can be described if three variables are known for all time during the period of displacement being considered: (1) the position of the pole of rotation, (2) the direction of relative motion, and (3) the magnitude of the angular velocity. An example of cycloid motion for a hypothetical plate is given figure 5.17. Plate 2 rotates around pole P2 with a given angular velocity. P2 appears to move with time relative to an observer on plate 1, tracing the path of a small circle. When the motion of plate 2 is combined with the relative motion of P2, a reference point M on plate 2 traces a figure of rotation around two axes known as a *spherical cycloid*.

Rates of plate motion can be quantified for the last approximately 150 Ma by correlating seafloor magnetic anomalies with the geomagnetic time scale (fig. 5.18). For convenience, magnetic anomalies are numbered with one at ridge axes. If the spreading rate has been constant in one ocean basin it is possible to extend the geomagnetic time scale back to more than 4.5 Ma using the magnetic anomaly patterns. Although data indicate that a constant spreading rate is unlikely in any ocean basin, the South Atlantic most closely approaches constancy (~ 1.9 cm/a) and is commonly chosen as a reference to extrapolate the time scale. Paleontologic dates

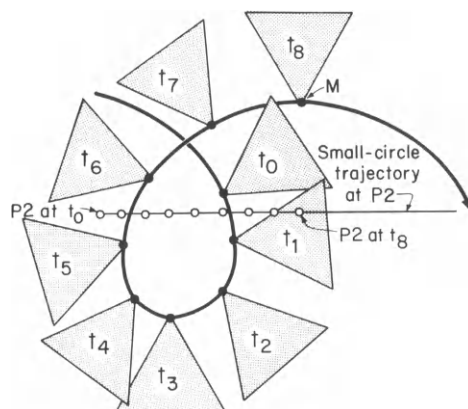


FIG 5.17 Spherical cycloid motion of plate 2 as its pole of rotation P2 processes along a line that represents a small circle around P1, the pole of rotation of plate 1. M is a point on plate 2. Modified after Cronin (1987).

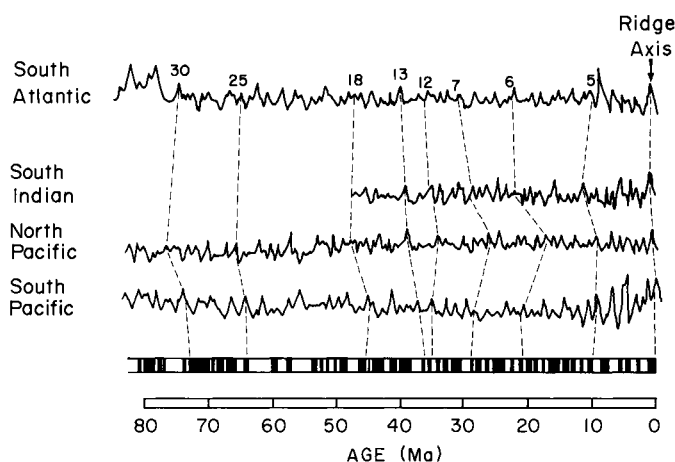


FIG 5.18 Magnetic profiles from the Atlantic, Indian, and Pacific Ocean basins. Modified after Heirtzler et al. (1968). Geomagnetic timescale is also given beneath the profiles with normal (black) and reversed (white) magnetized bands. Proposed correlations of anomalies are shown with dashed lines.

from sediment cores retrieved by the Deep Sea Drilling Project, and radiometric dates of basalts dredged from the ocean floor, substantiate an approximately constant spreading rate in the South Atlantic and allow the extension of the magnetic time scale to about 80 Ma. The correlations of magnetic anomalies with distance from ridge axes (fig. 5.18) indicate that spreading rates in the South Indian and North Pacific basins have been more variable and, on the average, faster than the spreading rate of the South Atlantic. Spreading rates also can be estimated from dislocation theory, using data derived from first-motion studies of earthquakes and from observed dip-lengths of subduction zones, if these lengths are assumed to be a measure of the amount of underthrusting during the last 10 Ma. Estimates of spreading rates, which are usually expressed as half-rates (the rate of one limb of an

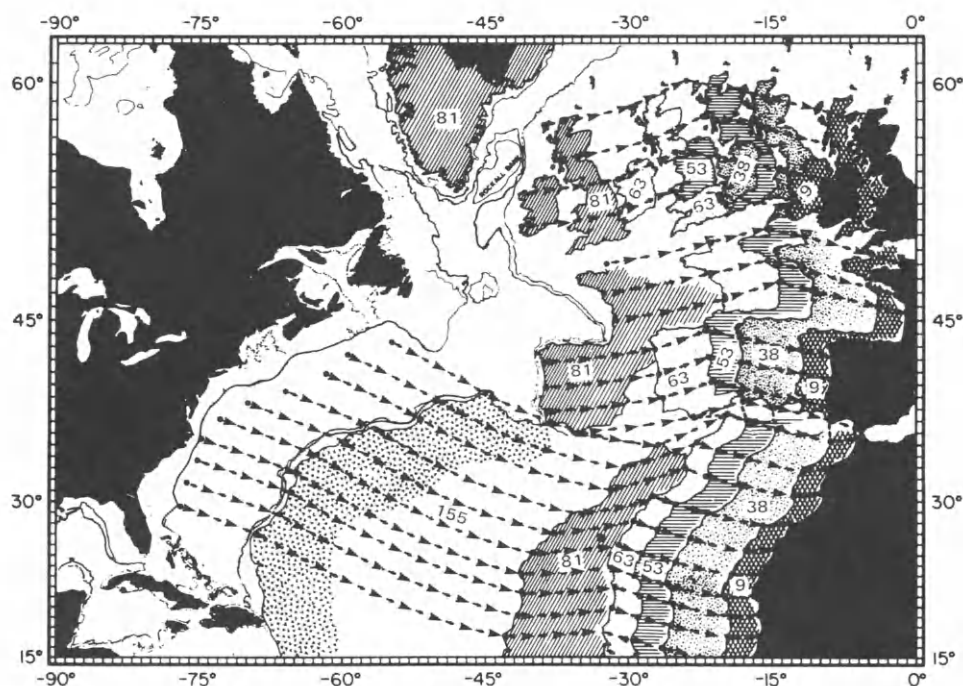


FIG 5.19 Seafloor spreading reconstruction of the opening of the North Atlantic (from Pitman and Talwani, 1972). Black continents represent present positions; dates for earlier positions are indicated in millions of years. Arrows represent flow lines.

ocean ridge), commonly range within a factor of two of one another using both methods. These rates range from about 1 to 20 cm/a, averaging a few centimeters per year, and variations occur from one segment of a ridge to another and as a function of time.

From estimated spreading rates it is possible to contour the age of the seafloor. Such contours are known as *isochrons*, and examples are shown on Plate I. Magnetic anomalies are numbered as shown in figure 5.18 (no. 30 $\approx$ 72 Ma, no. 7 $\approx$ 28 Ma). The rate of spreading has varied between crustal segments bounded by E-W trending transform faults for the last 90 Ma. Isochrons can be used to construct age maps of the seafloor. The oldest oceanic crust (Jurassic) occurs immediately adjacent to the Izu-Bonin subduction zone south of Japan. Since the rate at which oceanic crust has been produced at ridges during the past several hundred million years is of the order of a few centimeters per year, it is unlikely that crust much older than Jurassic will be found on the ocean floors today. The average age of oceanic crust is about 60 Ma. Fragments of oceanic crust older than Jurassic (i.e. ophiolites) are found in continental orogenic belts where they were tectonically emplaced during orogeny.

From calculated seafloor spreading directions and rates, it is possible to reconstruct plate positions and to estimate rates of plate separation for the last 200 Ma. One way of illustrating such reconstructions is by the use of flow or drift lines, as shown in figure 5.19 for the opening of the North Atlantic. The arrows indicate the relative directions of movement. Earlier positions of Africa and Europe relative

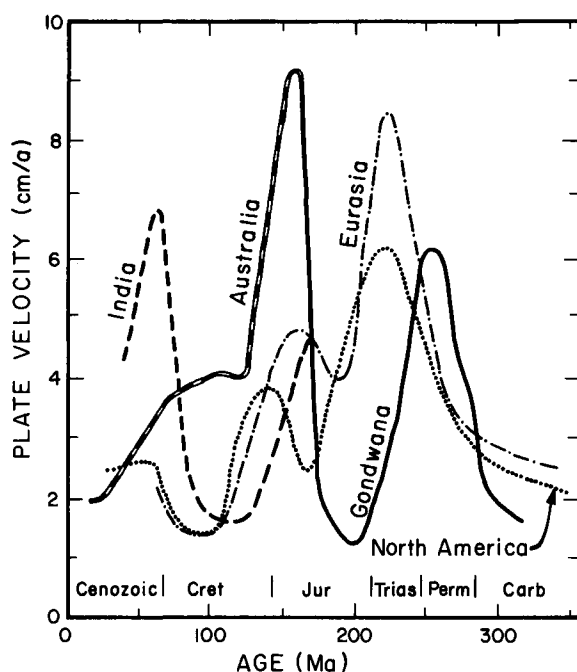


FIG 5.20 Velocities (root mean square) of continental plates calculated from apparent polar wandering paths for the last 300 Ma. Modified after Piper (1987).

to North America are also shown, with the corresponding ages in millions of years. The reconstruction agrees well with proposed geometric fits across the North Atlantic (fig. 1.3). It is clear from the lines of motion that Europe and Africa have been on two different plates for the last 160 Ma. Changes in spreading rates and directions, however, occur on both the African and Eurasian plates at the same time, at about 60 and 80 Ma ago. The separation of Africa and North America occurred primarily between 80 and 180 Ma, whereas separation of Eurasia from North America occurred chiefly in the last 80 Ma.

If true polar wander has been small compared to plate motions in the geologic past, it is possible to estimate minimum plate velocities from calculations of continental drift rates (Gordon et al., 1979; Bryan and Gordon, 1986). Also, if at least some hotspots have remained relatively fixed, it is possible to estimate plate motions relative to these hotspots (Hartnady and le Roex, 1985). A summary of estimates of continental drift rates is given in figure 5.20. Compared to average modern continental plate velocities of 10–20 cm/a relative to assumed fixed hotspots, it would appear that during the last 300 Ma most continental plates have moved faster than modern plates. Episodes of rapid plate motion are recorded during the Triassic–Early Jurassic (200–250 Ma) on most continents. Australia and India show peak velocities at about 150 and 50 Ma, respectively, after they fragmented from Gondwana. Results indicate that, in the past, continental plates for intervals of 50 ± 20 Ma have moved as fast as modern oceanic plates. It is noteworthy that maxima in continental plate velocities often occur just after fragmentation from a

supercontinent. For instance, the velocity maxima in the early Mesozoic follow the beginning of the breakup of Pangaea, and the peak velocities for Australia and India follow rifting of these continents from Gondwana.

Aseismic Ridges and Ocean Plateaus

About 10% of the ocean floors are covered by ridges or plateaus that show little if any seismic activity and variable (often none) volcanic activity. These features, which appear to have diverse origins, are known as *aseismic ridges* and *ocean plateaus*. More than 100 are known and they are particularly widespread in the western Pacific (Ben-Avraham et al., 1981) (see fig. 6.26). Aseismic ridges and oceanic plateaus (hereafter, ROP) rise thousands of meters above the seafloor and some, such as the Seychelles Bank in the Indian Ocean, rise above sea level. Seismic data suggest crustal thicknesses of 20–40 km and most ROP do not have magnetic lineations indicating they are not typical oceanic crust. Drilling and dredging suggest diverse origins. Some have granitic basement (Seychelles Bank, Parcel Islands in the South China Sea, the Lord Howe Rise north of New Zealand and the Agulhas Plateau south of Africa), suggesting they are rifted fragments of continental crust. Others, such as the Cocos and Carnegie ridges west of northern South America (Plate I), are of volcanic origin, related probably to hotspot activity. Some, such as the Palau–Kyushu ridge south of Japan, are extinct oceanic arcs. One of the most exciting aspects of ROP is that they are moving on oceanic plates and should eventually end up in subduction zones. Because they are more buoyant than oceanic crust they may be scraped off and accreted to active arc or active continental margins, and continental growth may occur by accretion of ROP (Chapt. 10).

What happens when a ROP enters a subduction zone and collides with an arc? Such collisions are characterized by changes in the arc system as follows:

(1) Because ROP resist subduction, they may produce a cusp in the arc system, as illustrated for instance by the intersection of the Caroline and Marcus ridges with the Mariana arc south of Japan (Plate I). Paleomagnetic and structural geologic data from the Mariana arc support this interpretation, indicating that the arc was rotated at its ends by collision of these ridges between 10 and 30 Ma ago (McCabe, 1984).

(2) Volcanic gaps in arc systems commonly occur at points of collision between ROP and arcs (McGeary et al., 1985). Examples are the Tehuantepec, Cocos, Carnegie, Nazca, and Juan Fernandez ridges along the Middle American and Peru–Chile subduction systems (Plate I). In some instances ROP appear to accrete to arcs, while in others they may subduct beneath arcs. If a ROP completely resists subduction and becomes accreted to an arc, subduction may stop, producing a volcanic gap. Volcanic gaps also characterize some collisions in which the ROP appear to subduct. Cessation of volcanism in these cases may be due to buoyant subduction such that the descending slab slides beneath the overlying lithosphere without a mantle wedge as in the case of central Peru (fig. 5.15). Magmas may not be produced due to elimination of the mantle wedge which plays a major role in the production of subduction-related magmas (Chapt. 7).

(3) Seismic gaps also occur at some ROP–arc collision zones, indicating that the ROP did not subduct.

Hotspots and Plumes

General Characteristics

The hotspot model (Wilson, 1963) that suggests linear volcanic chains and aseismic ridges form as oceanic crust moves over relatively stationary magma sources, has been widely accepted in the geological community. *Hotspots* are generally thought to form in response to mantle plumes, which rise like salt domes in sediments, through the mantle to the base of the lithosphere. Progressive adiabatic melting in the upper mantle leads to large volumes of magma which are partially erupted (or intruded) at the Earth's surface. Hotspots are characterized by the following features:

(1) A relative topographic high of 500–1200 m and typical widths of 1000–1500 km. These highs are supported by asthenosphere or reheated lithosphere, and persist beneath hotspot traces, implying that they become fixed to the lithosphere (Crough, 1983). Highs are probably indirect manifestations of ascending mantle plumes.

(2) Uplifts, especially in oceanic areas, are capped by active or recently active volcanoes. Examples are Hawaii and Yellowstone Park in the western United States.

(3) Most hotspots are characterized by gravity highs reflecting the rise of more dense mantle material. Some, however, have gravity lows.

(4) One of two aseismic ridges of mostly extinct volcanic chains lead away from hotspots. These are known as *hotspot traces* and examples are illustrated in figure 5.21.

(5) High heat flow, probably reflecting a mantle plume, characterizes all modern hotspots.

Somewhere between 40 and 150 active hotspots have been described on the Earth. The best-documented hotspots (42 in number) have a rather irregular distribution occurring in both oceanic and continental areas (fig. 5.21). Some occur on or near ocean ridges, such as Iceland, St Helena, and Tristan, while others occur near the centers of plates. An important correlation seems to exist (even when a larger number of hotspots is considered) between hotspot distribution and geoid height at the core–mantle boundary. The large number of hotspots in and around Africa and in the Pacific basin correspond to the two major geoid highs (fig. 3.11) (Stefanick and Jurdy, 1984). The fact that the geoid highs appear to reflect processes in the deep mantle or core supports the idea that hotspots are caused by mantle plumes rising from the deep mantle.

A major problem that has stimulated a great deal of controversy is whether hotspots remain fixed relative to plates. The almost perfect fit of volcanic chains in the Pacific plate with a pole of rotation at 70°N, 101°W and a rate of rotation about this pole of about 1 deg/Ma for the last 10 Ma, suggests that Pacific hotspots have remained fixed relative to each other over this period of time. Also consistent with relatively fixed positions of these hotspots is the fact that the sedimentation rate determined in cores of deep-sea sediments just north of the equator has its highest rate when the model predicts that the sediments were at the equator, where maximum sedimentation rates are observed today. If all hotspots have remained fixed with respect to each other, it should be possible to superimpose the same

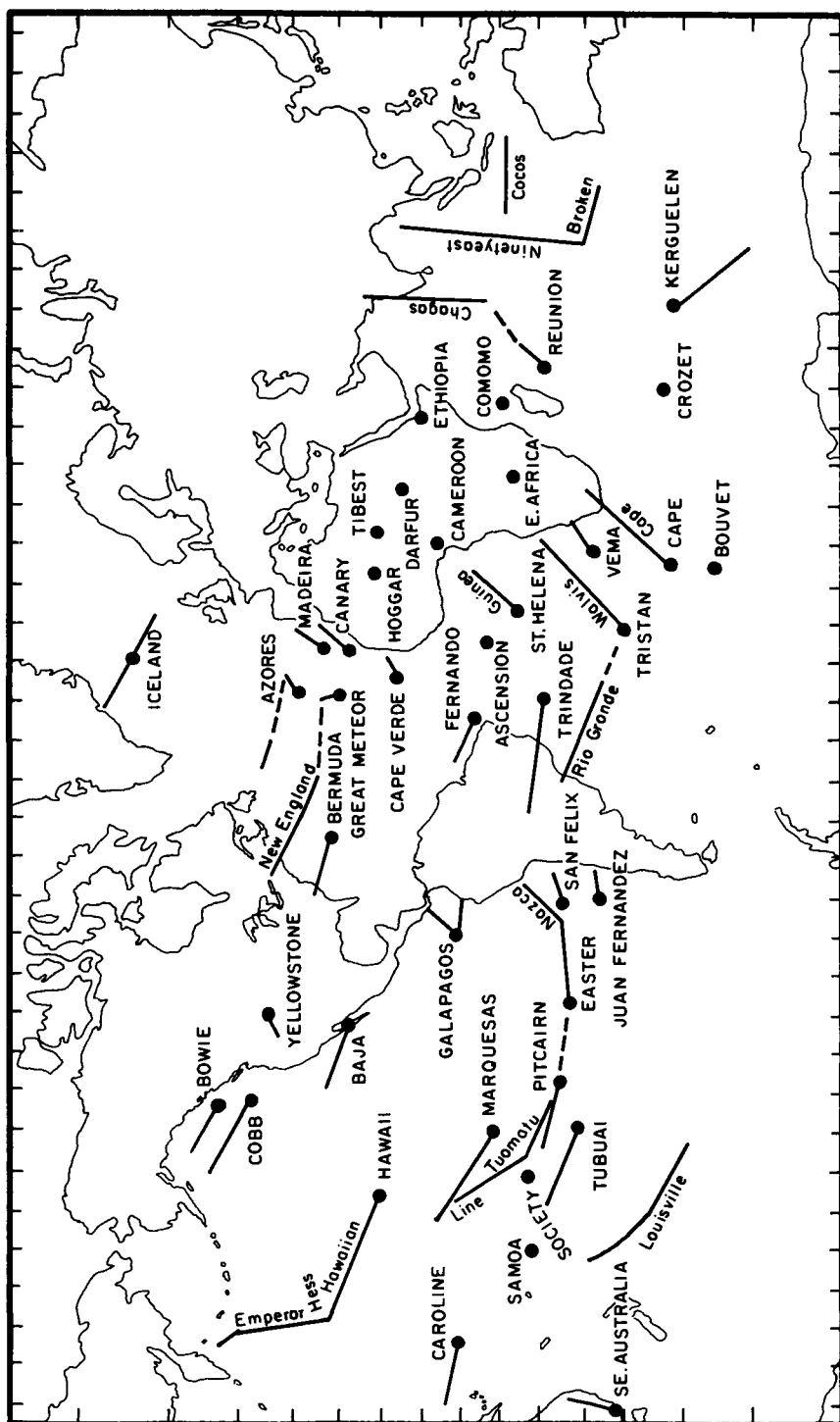


FIG 5.21 Distribution of hotspots (dots) and hotspot traces (lines). Modified after Crough (1983).

hotspots in their present positions on their predicted positions at other times in the last 150–200 Ma. Except for hotspots in the near proximity of each other, however, it is generally not possible to do this, suggesting that hotspots move in the upper mantle (Molnar and Stock, 1987). Rates of hotspot motion are more than an order of magnitude less than seafloor spreading rates. Data suggest that hotspots in the Pacific, Atlantic, and Indian Ocean basins move relative to each other at rates of 1–2 cm/a. Absolute plate motions calculated relative to the Hawaiian hotspot, and assuming a rate of volcanic migration along the Hawaiian chain of 10 cm/a, indicate that the North American plate is moving SW at 3–4 cm/a, the South American plate WNW at 3–3.5 cm/a and the African plate N to NNE at 1–2 cm/a (McDougall and Duncan, 1980). Several lines of evidence suggest that the African and Antarctic plates have been relatively stationary for the last 25 Ma.

Hotspot Traces

Chains of seamounts and volcanic islands interpreted in terms of the hotspot model are common in the Pacific basin. Major Pacific chains include the Hawaiian–Emperor, Line, Society, and Austral islands, all of which are subparallel to either the Emperor or Hawaiian chains and approximately perpendicular to the axis of the East Pacific Rise (Plate I). Closely spaced volcanoes form aseismic ridges such as the Ninetyeast ridge in the Indian Ocean and the Walvis and Rio Grande ridges in the South Atlantic (Plate I). Radiometric dates demonstrate that the focus of volcanism in the Hawaiian chain has migrated to the southeast at a linear rate of about 10 cm/a for the last 30 Ma (fig. 5.22) (McDougall and Duncan, 1980). A major change in spreading direction of the Pacific plate is recorded by the bend in the Hawaiian–Emperor chain at about 43 Ma, and this change is also demonstrated by magnetic anomaly distribution in the northern Pacific plate. Similar linear decreases in the age of volcanism occur toward the southeast in the Marquesas, Society, and Austral islands in the South Pacific, with rates of migration of the order of 11 cm/a and in the Pratt-Welker seamount chain in the Gulf of Alaska at a rate of about 4 cm/a. The life spans of hotspots vary, and probably depend on such parameters as plume size and the tectonic environment into which a plume is emplaced. On the Pacific plate three volcanic chains were generated by hotspots between 25 and 70 Ma, whereas 12 chains have been generated in the last 25 Ma.

Hotspots may interact with lithospheric plates in a variety of ways, some of which are illustrated in figure 5.23 (Epp, 1984). If oceanic plate motions relative to hotspots are small, large amounts of magma are erupted (or intruded) forming a large island with thick crust as exemplified by Iceland and oceanic plateaus such as the Shatsky rise (a) (Plate I). If plate hotspot motion is erratic in direction and rate, irregular clusters of volcanoes may form (b) as characterized perhaps by the Mid-Pacific Mountains (Plate I). Volcanic chains that cross transform faults often show cross-trends of volcanoes parallel to the faults. Such is the case along the Hawaiian ridge where it crosses the Molokai and Murray fracture zones. Transform faults are zones of weakness, and magma may be injected into them as they pass over hotspots (c). Another interaction is possible if a plume breaks out at the base of the lithosphere and flows from the hotspot back along the weakened hotspot trace (d). This could result in irregularities in the age of volcanism along the hotspot trace as

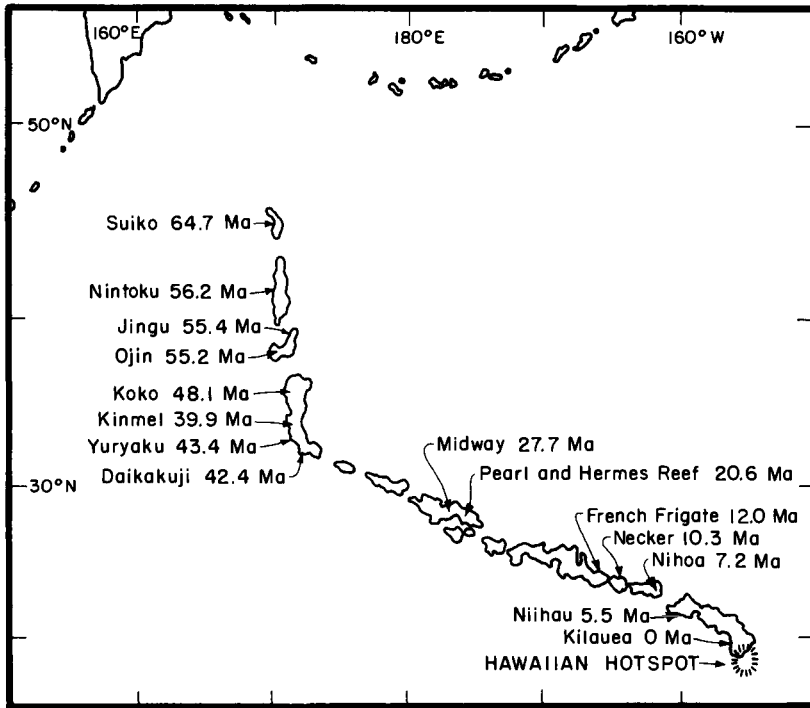


FIG 5.22 Map of the Hawaiian-Emperor volcanic chain showing locations and ages of last volcanism. The Hawaiian hotspot today is centered just south of the island of Hawaii. Modified after Molnar and Stock (1987).

observed in some island chains. Because the lower lithosphere is at near-melting temperatures, magmas may be stored here for considerable lengths of time. This can lead to eruption of magma (generally small amounts) along the volcanic chain after passing over the hotspot (e). Also hotspot conduit systems may be bent by the plate motion. When the conduits are bent to some critical angle the plume rises and forms a new conduit (f). If an individual conduit is not disconnected from the plume, eruption may continue in volcanoes that have passed over the hotspot.

Hotspot traces also have manifestations in the continents, although less well-defined than in ocean basins. For example, North America moved NW over the Great Meteor hotspot in the Atlantic Basin between 100 and 150 Ma ago. The trajectory of the hotspot is defined by the New England seamount chain (Plate I) and Cretaceous kimberlites and alkalic complexes in New England. Dated igneous rocks fall near the calculated position of the hotspot at the time they formed (Crough et al., 1980). Geologic data and paleotemperatures indicate that this region was elevated at least 4 km as it passed over the hotspot. When post-Triassic kimberlites from North and South America and Africa are rotated to their position of origin relative to present Atlantic hotspots, the majority appear to have formed within 5 degrees of a mantle hotspot. As an example, the calculated trajectory of the Trindade hotspot (east of Brazil, see Plate I) matches the locations of three dated kimberlites from Brazil and also roughly coincides with the distribution of alluvial diamond deposits (which are derived from kimberlites) (fig. 5.24). High heat flow, low seismic

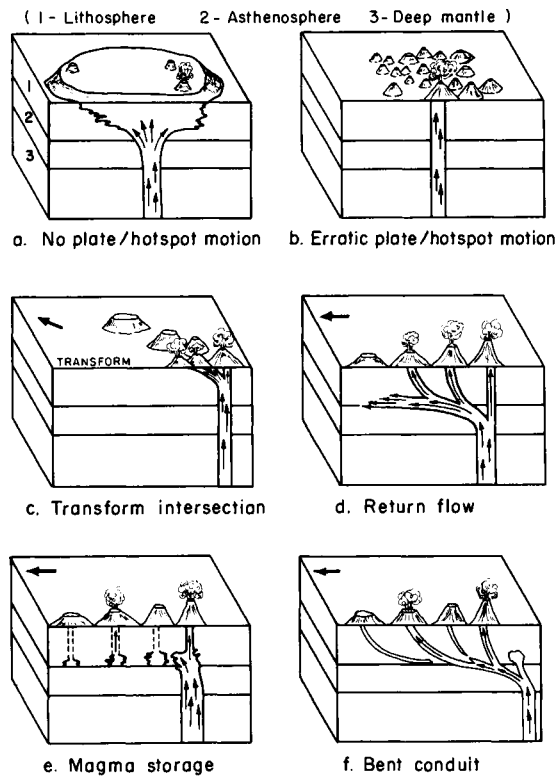


FIG 5.23 Some possible interactions of lithospheric plates and hotspots. Modified after Epp (1984).

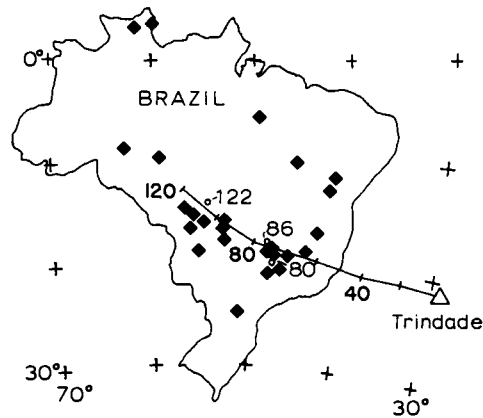


FIG 5.24 Locations of dated kimberlites (○) and major alluvial diamond deposits (◆) in Brazil compared to calculated path of the Trindade hotspot. After Crough et al. (1980).

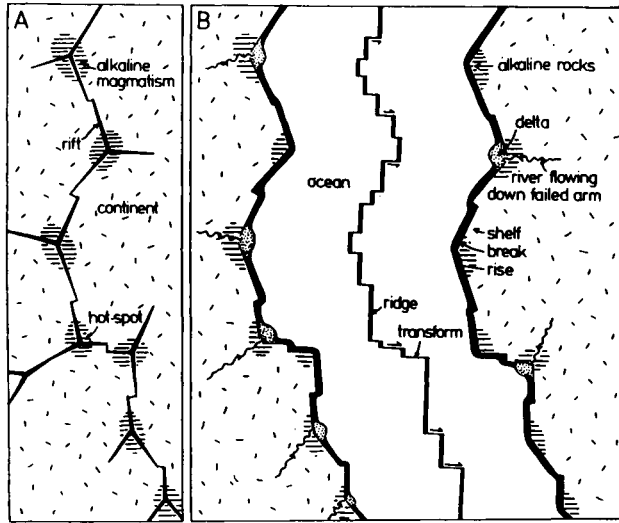


FIG 5.25 Early stages in the opening of an ocean basin, showing the possible influence of hotspots. From Dewey and Burke (1974).

velocities and densities at shallow depth, and high electrical conductivity at shallow depth beneath Yellowstone National Park in Wyoming are interpreted to reflect a mantle hotspot at this locality (Smith and Christensen, 1980). The movement of the North American plate over this hotspot during the past 15 Ma has been accompanied by the development of the Snake River volcanic plain (Plate I), with the oldest volcanics occurring at the southwest extremity of the plain in southwest Idaho. As the plate moved southwest, volcanism migrated northeast. Other examples of hotspot traces, in part in continental areas, are the Mesozoic granites in Nigeria; the Thulian volcanic chain extending from Iceland to Ireland and recording the opening of the North Atlantic; and the central European volcanic province extending from western Germany to Poland.

Hotspots also may be important in the break-up of continents (Dewey and Burke, 1974). Mantle plumes, for instance, may rupture continents producing RRR triple junctions (fig. 5.25A). Major irregularities along rifted continental margins may be inherited from original plume distributions (B). Major drainage systems are commonly localized in the failed arms of plume triple junctions (aulacogens) resulting in delta progradation at their mouths. Examples are the major river systems entering the Atlantic Ocean from South America and Africa.

Plume Generation

As mentioned in Chapter 3, seismic velocity variation in the D'' layer at the base of the mantle may reflect instability above the molten core, and as a result hot plumes of lower mantle may be generated in this layer. Numerical simulations of the D'' layer support this possibility, and show that plumes may be formed by coalescence of small-scale convective instabilities (Olson et al., 1987). The simulated sequence of events for a layer with a permeable top is shown in figure 5.26. The thermal structure about 130 Ma after plume initiation suggests a 50–100-km-wide plume

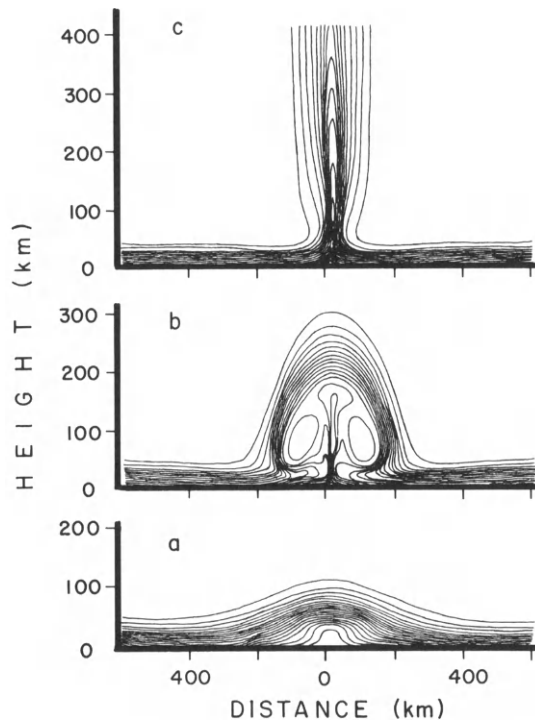


FIG 5.26 Temperature contours of simulated plume derived from the D'' layer at 111 (a), 127 (b) and 143 (c) Ma after thermal perturbation from below. Modified after Olson et al. (1987). Contour interval is 40°C.

surrounded by boundary zones about 50 km thick (b). By about 140 Ma the plume has ascended, leaving a trailing conduit (c). These simulated plumes exhibit a time-dependent behavior including upward propagating conduits that contribute to episodicity of hotspot magmatism. Results further suggest that successful plumes (i.e. those that rise out of the D'' layer) should be produced if a temperature perturbation in the source layer exceeds 200°C. Plumes produced in the deep mantle in such a manner seem to require about 50–100 Ma to rise to the base of the lithosphere, and this is much shorter than the estimated times of convective overturn in the mantle (Chapt. 3).

Summary Statements

1. The Earth's magnetic field has reversed its polarity many times in the geologic past. Normal and reverse polarity intervals in the stratigraphic record allow construction of a geomagnetic time scale. Magnetic reversals show periodicity on several scales, and evidence of reversals exists in rocks as old as 3.5 Ga. Reversals appear to be caused by changes in fluid motions in the outer core.
2. Polarity intervals correlate with magnetic anomaly distributions on the seafloor allowing seafloor spreading rates to be estimated. The magnetic anomalies are caused by magnetized basalt injected into axial zones of ocean ridges during normal and reversed polarity intervals.

3. During a reversal the Earth's dipole field decreases and rapid changes occur in declination and inclination.
4. For apparent polar wandering paths calculated from paleomagnetic data to have significance in terms of ancient plate motions, each of the following problems must be considered: secular variation in the magnetic field, the existence of a dipole field in the past, the magnitude of true polar wander, identification of normal and reversed poles, deformation of the rocks, dating of magnetization and separation of various remanent magnetizations.
5. Apparent polar wandering paths show distinct characteristics for various plate tectonic scenarios. Results indicate that continents have grown by microcontinent or arc collisions.
6. Plate boundaries are of four types: ocean ridges where new lithosphere is created; subduction zones where lithosphere descends into the mantle; transform faults where plates slide by each other; and collisional zones, where continents or arcs have collided. Plate boundaries evolve with time and may change from one type to another or become extinct, and new boundaries are continually being created.
7. Divergent plate boundaries (ocean ridges) are characterized by small-magnitude, shallow earthquakes with vertical motions reflecting injection of new lithosphere. Topography in axial rifts varies from high relief to little if any relief in going from slow- to fast-spreading ridges. Ocean ridges grow by lateral propagation.
8. Transform faults are characterized by shallow, variable-magnitude earthquakes exhibiting strike-slip motion. Ridge–ridge transforms retain a constant length with time, whereas other transforms lengthen or shorten with time.
9. Convergent boundaries (subduction zones) are characterized by a dipping seismic zone with variable-magnitude earthquakes and, in some instances, seismic gaps suggestive of plate fragmentation. Fault motions vary with depth in the seismic zone and seismicity is strongly correlated with degree of coupling of descending slab and mantle wedge. Plates <50 Ma in age may be buoyantly subducted and slide beneath the overriding plate.
10. Collisional boundaries are characterized by wide (≤ 3000 km) zones of lateral deformation with compressive fault motions dominating near sutures and strike-slip or vertical motions in the plateau areas overlying partially subducted plates. Arc–arc (or arc–continent) collisions may result in a subduction zone reversal.
11. The motion of one plate relative to another is that of a spherical cycloid, and is a function of the position of the pole of rotation, the direction of relative motion, and the angular velocity of the plate. Average relative velocities of oceanic plates range from 1 to 20 cm/a and the oldest surviving seafloor is about 150 Ma.
12. Continental plates appear to have moved faster than at present during the last 300 Ma, as deduced from apparent polar wandering paths. Maxima in ancient continental plate velocities often occur after fragmentation of a supercontinent.
13. Collision of aseismic ridges and oceanic plateaus with arc systems may cause bends in subducting plates as well as volcanic and seismic gaps in the arc. These ridges and plateaus may be accreted to the arc or subducted into the mantle.
14. Chains of volcanic islands and aseismic ridges on the seafloor appear to have formed as oceanic plates move over hotspots in the upper mantle. Similar,

although not as well defined, trajectories are formed when continents move over hotspots. Lifespans of hotspots appear to be ≤ 70 Ma.

15. Hotspots are probably caused by mantle plumes that rise rapidly (50–100 Ma) from the lower mantle, perhaps from the D'' layer above the core–mantle boundary. A correlation of hotspot distribution and topographic highs on the core–mantle boundary supports a deep origin for plumes.

Suggestions for Further Reading

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CHAPTER 6

Tectonic Settings

Introduction

To identify specific tectonic environments in the geologic past it is necessary to equate lithologic associations and other geologic features preserved in the geologic record to tectonic setting. From studying modern plate settings it is possible to learn more about what to look for in ancient rocks to be able to identify, or at least constrain, the tectonic setting in which they formed. It also may be possible with such information to evaluate the intriguing question of just how far back in geologic time plate tectonics has operated. Characteristic features of tectonic setting include lithologic assemblages (both supracrustal rocks and intrusive rocks), deformational styles and histories, metamorphism and P–T paths, as well as mineral and energy deposits. Although we can sample and study sediments forming in modern plate settings, it is more difficult to study young plutonic and metamorphic rocks and deep-seated deformation. One approach is to examine deep canyons in mountain ranges where relatively young (<50 Ma) deep-seated rocks are exposed, to which a tectonic setting can be assigned with some degree of confidence. Seismic reflection profiles provide a means to study deep-seated deformation. These studies have been particularly useful in modern arc systems. We can also study recently elevated blocks that expose lower crustal rocks as well as xenolith populations brought up in young volcanoes (as discussed in Chapt. 4). Finally, it is possible to simulate P–T conditions that exist in modern plates in the laboratory and in computer models, from which constraints can be placed on geologic processes occurring at depth.

This chapter will examine some of the processes that leave tectonic imprints in rocks. It is only when results from a variety of studies converge on the same interpretation, that ancient tectonic settings can be identified with confidence. Even then, one must use care in that one or more tectonic settings that no longer exist on the Earth may have existed in the past, and they may have left imprints in the geologic record similar to those of modern plate settings. This is especially true for the Archean.

Petrotectonic Assemblages

Rock assemblages that characterize particular modern plate tectonic settings are known as *petrotectonic assemblages*. Such assemblages include both supracrustal rocks (sediments and volcanics) and intrusive rocks. Petrotectonic assemblages are well

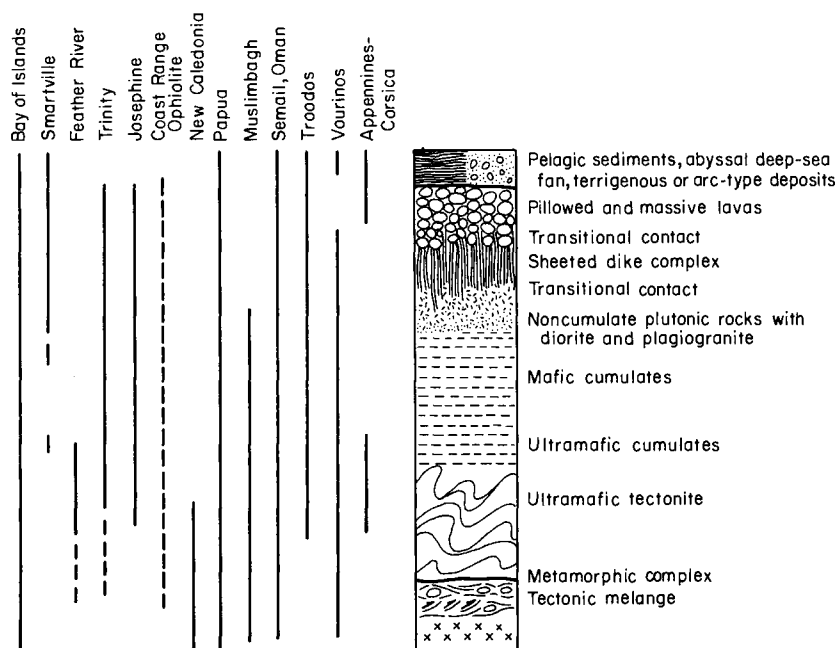


FIG 6.1 An idealized ophiolite succession compared with various exposed ophiolites. Modified after Moores (1982).

known for ocean ridges and basins, subduction zones, continental rifts (including hotspots and aulacogens), cratons, and collisional orogens. Lithologic assemblages formed of each of these environments are described in the following sections.

Ophiolites

Ophiolites are tectonically emplaced successions of mafic and ultramafic rocks that are generally considered to represent fragments of oceanic crust (Moores, 1982). An ideal ophiolite includes from bottom to top the following units (fig. 6.1): (1) ultramafic tectonite; (2) layered cumulate gabbros and ultramafic rocks; (3) non-cumulate gabbros, diorites, and plagiogranites; (4) sheeted diabase dikes; and (5) pillowed basalts. Overlying this succession in many ophiolites are abyssal or/and pelagic sediments or arc-related volcanoclastic sediments. Due to faulting or other causes the idealized ophiolite succession is rarely found in the geologic record. Instead, one or more of the ophiolite units are missing or they have been dismembered by faulting and occur as blocks, perhaps in a melange.

Some ophiolites are in fault contact with underlying shallow marine cratonic sediments, while others occur as tectonic slivers in an accretionary prism composed of graywackes and other arc-related rocks. These ophiolites appear to be emplaced along passive and active continental margins, respectively. The basal ophiolite melange (fig. 6.1) consists of a chaotic mixture of diverse rocks in a highly sheared matrix. Clast lithologies include ophiolite-derived materials, pelagic and abyssal sediments, graywackes, and various metamorphic and volcanic rocks. Matrices are commonly sheared serpentinite. Most basal ophiolite melanges appear to be of tectonic origin formed during ophiolite emplacement. Overlying the melange, and in

some cases gradational with it, is a thin layer of metamorphic rocks, which may be zoned with the highest metamorphic grade at the top (near the ultramafic). Mineral assemblages reflect temperatures of $\geq 500^{\circ}\text{C}$ and burial depths of < 20 km. Most data suggest the metamorphic rocks formed in (or at the base of) the oceanic crust prior to ophiolite emplacement, perhaps along a transform fault.

Most ophiolites contain sheared and serpentinized ultramafic rocks above the metamorphic rocks. These ultramafic tectonites are composed chiefly of harzburgite with pronounced foliation and, generally, compositional banding. Lenses of dunite and chromite occur within the harzburgites. Overlying the tectonites are cumulate ultramafic and gabbroic rocks that have formed by fractional crystallization. These rocks have cumulus textures and well-developed compositional banding. Some ophiolites contain noncumulate gabbro, diorite, and plagiogranite in the upper part of the noncumulate zone. *Plagiogranites* are tonalites composed of quartz and sodic plagioclase with minor hornblende. They typically have granophyric intergrowths and may be intrusive into layered gabbros.

Above the noncumulate unit in an idealized ophiolite is a sheeted dike complex (fig. 6.1). Dikes may crosscut or be gradational with the noncumulate rocks. Although dominantly diabases, dikes range from diorite to pyroxenite in composition. Dike thickness is variable, commonly from 1 to 3 m. One-way chilled margins are a characteristic feature generally interpreted to reflect vertical intrusion in an oceanic axial rift zone where one dike is intruded in the center of another as the lithosphere spreads. The transition from sheeted dikes into pillow basalts generally occurs over an interval of 50–100 m where screens of basalt between dikes become more abundant upwards. The uppermost unit of ophiolites is comprised of extrusive ocean-ridge basalts as pillowed flows or hyaloclastic breccias. Thickness of this unit varies from a few meters to 2 km, and pillows form a honeycomb network with individual pillows ranging up to 1 m across. A few dikes cut the pillowed basalt unit.

Many ophiolites are overlain by sediments reflecting pelagic, abyssal, or arc depositional environments. Pelagic sediments include radiolarian cherts, red fossiliferous limestones, metalliferous sediments, and abyssal sediments. Abyssal sediments are chiefly pelites and siltstones that were deposited on abyssal plains, and often show evidence of both volcanic and continental provenance. Modern abyssal plains are most frequent adjacent to passive continental margins, and hence have significant input of sediments from continental shelves. Some ophiolites are overlain by graywackes and volcanoclastic sediments of arc provenance.

Subduction-Related Assemblages

Numerous geologic environments are associated with subduction. In an idealized arc system three zones are recognized: the arc–trench gap, the arc, and the arc–rear area (fig. 6.2). From the oceanside landwards, a continental-margin arc is characterized by the trench; an accretionary prism with overlying forearc basins; the volcanic arc with intra-arc basins; a fold-thrust belt; and a retroarc foreland basin (fig. 6.2a). An oceanic arc differs from a continental-margin arc primarily in the arc rear area where it is composed of an active back-arc basin, a remnant arc, and an inactive back-arc basin (fig. 6.2b). Each of these arc environments is characterized by a lithologic assemblage.

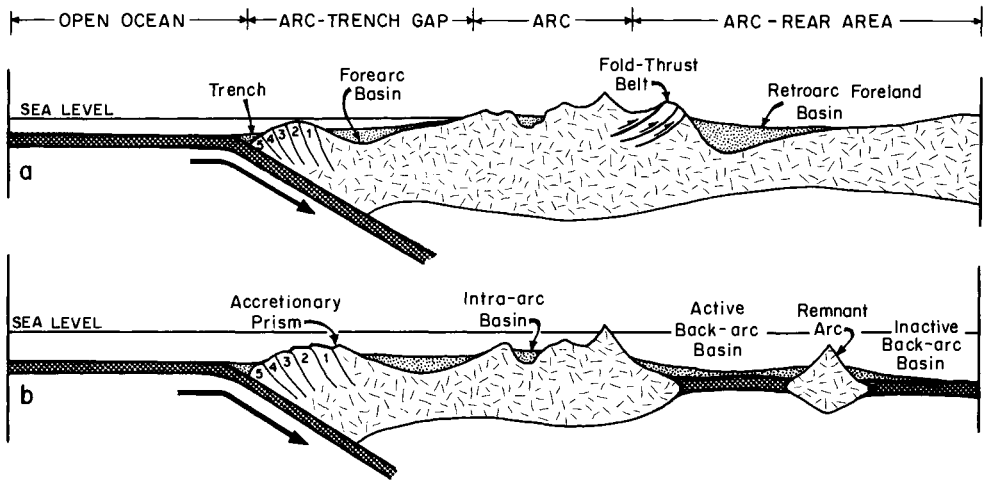


FIG 6.2 Diagrammatic cross-sections of a continental margin arc (a) and an island arc (b) showing major tectonic divisions. Numbers in accretionary prism indicate relative age of slabs (1 being the oldest).

The Trench

Trench sediments are dominantly graywacke turbidites with minor pelagic components. Turbidity currents generally enter trenches at submarine canyons and flow along trench axes. Sediments can be transported along trenches for up to 300 km, as for example in the Sunda trench south of Sumatra, where detritus from the Himalayas enters the trench on the north from the Bengal submarine fan (Moore et al. 1982). Although most detrital sediment is clay or silt, sand and coarser sediments may be deposited in proximal facies. Down-axis changes in facies suggest that trenches are filled by a succession of radiating submarine fans. This longitudinal filling is supplemented by lateral infilling from side canyons. Provenance of trench turbidites is variable. If a trench heads near a continent, continental sources generally dominate with volcanic arc sources of minor importance. Seismic reflection studies of trench sediments show hummocky topography in places suggestive of buried debris flows or submarine slumps. Trench sediments may be preserved in the geologic record if they are added to the arc as faulted slabs in the accretionary prism.

The Accretionary Prism

The *accretionary prism* (or subduction zone complex) consists of a series of steeply inclined, fault-bounded wedges of sediment and volcanic rocks above a descending slab. These wedges represent oceanic crust and trench sediments that have been accreted to the front of the arc. Individual wedges in the accretionary prism decrease in age as the trench is approached (fig. 6.2). In addition to fragments of oceanic crust (ophiolites, pelagic sediments, etc.) and trench sediments, the characteristic rock type of the accretionary prism is melange. Because melanges have different characteristics and origins a widely accepted definition of melange has not yet emerged. One common definition is that *melange* is a mappable body of rock characterized by the lack of continuous bedding and the inclusion of fragments of

rocks of all sizes (up to more than a kilometer) contained in a fine-grained, often sheared matrix (Raymond, 1984). Both sedimentary and tectonic processes can give rise to such units. *Olistostromes* are melanges produced by gravitational sliding and accumulate as semi-fluid bodies that do not have bedding, but include associated turbidites. Clasts within melanges may be exotic (derived from another environment) or native (reworked from the immediate environment) and are generally matrix supported. Clast lithologies commonly include graywacke, mudstone, chert, basalt (greenstone), and other ophiolite lithologies, arc volcanics and rare granitic rocks. Melanges are usually folded and may contain more than one cleavage or foliation. Sheared matrices are usually comprised of serpentine and fine-grained rock and mineral fragments.

Although melanges are typical of accretionary prisms they may be formed by several processes (Cowan, 1986). Tectonic melanges are produced by compressive forces along the upper part of descending slabs at shallow depths. Fragmentation and mixing of largely lithified rocks may occur along a migrating megashear zone subparallel to a subducting slab. Fragments of oceanic crust and trench sediment are scraped off the descending plate and accreted to the overriding plate. *Olistostromes* may be produced by gravitational slumping or debris flows on the oversteepened trench walls or along the margins of a forearc basin. Debris flows in which clay minerals and water form a single fluid possessing cohesion are probably the most important transport mechanism of *olistostromes*.

Forearc Basins

Forearc basins are marine depositional basins on the trench side of arcs. They vary in size and abundance with the evolutionary stage of an arc. In continental-margin arcs such as the Sundra arc, forearc basins range up to 700 km in strike length (Moore et al., 1982). They overlie the accretionary prism, which may be exposed as submarine hills within and between forearc basins. Sediments in forearc basins, which are chiefly turbidites with sources in the adjacent arc system, range up to many kilometers in thickness. Hemipelagic sediments are also of importance in some basins, such as those in the Mariana arc. *Olistostromes* can form in forearc basins by sliding and slumping from locally steepened slopes. Forearc-basin clastic sediments may record progressive unroofing of adjoining arcs, as for instance reflected by the Great Valley Sequence (Jur-Cret) in California (Dickinson and Seely, 1986). Early sediments in this sequence are chiefly volcanic detritus from active volcanics and later sediments reflect progressive unroofing of the Sierra Nevada batholith. Volcanism is rare in modern forearc regions and neither volcanic nor intrusive rocks are common in older forearc successions.

Volcanic Arcs

Volcanic arcs range from entirely subaerial, such as the Andean and Middle America arcs, to mostly or completely submarine, such as many of the immature oceanic arcs in the SW Pacific. Other arcs, such as the Aleutians, change from subaerial to partly submarine along strike (Plate I). Subaerial arcs are composed chiefly of flows and associated pyroclastic deposits, which often occur in large stratovolcanoes. Submarine arcs are composed of pillowed flows and large volumes of hyaloclastic tuffs and

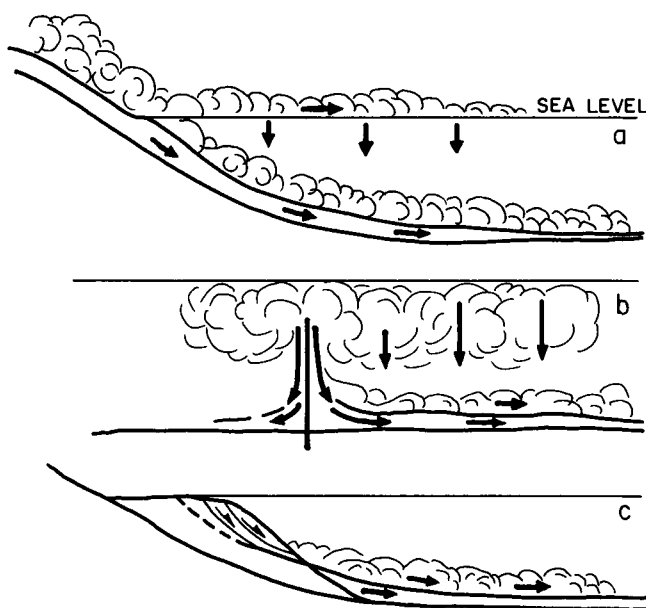


FIG 6.3 Mechanisms for the origin of subaqueous ash-flows (from Fisher, 1984). (a) Hot ash-flow flows from land into water; (b) ash-flow forms from column collapse; (c) ash-flow develops from slumping of pyroclastic debris.

breccias. Volcanism begins rather abruptly in arc systems at the *volcanic front*, which occurs 80-150 km inland from the trench. Both tholeiitic and calc-alkaline magmas characterize arcs (Chapt.7) with andesites and basaltic andesites often dominating. Felsic magmas are generally emplaced as batholiths, although felsic volcanism is common in most continental-margin arcs.

Active Back-arc Basins

Active back-arc basins (ABAB) occur over descending slabs behind arc systems and are characterized by high heat flow, relatively thin lithosphere, and in many instances an active ocean ridge, which is enlarging the size of the basin. Sediments are extremely varied depending on basin size and nearness to an arc. Proximal to the arc (and to the remnant arc), volcanoclastic sediments generally dominate, whereas in more distal regions pelagic, hemipelagic and biogenic oozes may predominate. During the early stages of basin opening, thick epiclastic deposits largely representing gravity flows are important. With continued opening of an ABAB, these deposits pass laterally into turbidites, which are succeeded distally by pelagic and biogenic sediments (Leitch, 1984). Discrete layers of air-fall tuff may be widely distributed in ABAB. Early stages of basin opening are accompanied by diverse magmatic activity including felsic volcanism, whereas later evolutionary stages are characterized by an active ocean ridge. Most ophiolites carry a subduction zone geochemical signature (Chapt. 7), and thus appear to have formed in ABAB.

Subaqueous ash flows may erupt or flow into ABAB and such flows can develop in three principle ways (Fisher, 1984) (fig. 6.3). The occurrence of welded ash-flow tuffs in some ancient ABAB successions suggests that hot ash flows enter water

without mixing and retain enough heat to weld (case a). Several mechanisms have been suggested by which ash flows retain heat. During an eruption a cloud of turbulent ash may flow across (as well as beneath) the water, as observed in some modern eruptions (a). Submarine eruptions of felsic magma may eject large amounts of ash into the sea which fall onto the seafloor forming a dense, water-rich debris flow (case b). In the closing stages of eruption, small turbidity currents are deposited as thin graded beds above these debris flows. Finally, submarine ash flows may develop by slumping of unstable slopes composed of pyroclastic debris (case c). Turbidites in ABAB are deposited in large coalesced submarine fans (Carey and Sigurdsson, 1984). These fans receive sediment from multiple volcanic sources along the arc, and massive influxes of sediment are related to arc volcanism and/or to uplift of the arc (Klein, 1985). Results from modern arcs suggest that rates of arc uplift must be >400 m/Ma to generate the requisite sediment yield to form submarine fans.

Processes controlling ABAB deposition include latitude-dependent biogenic productivity and oceanic circulation, climate, volume of arc volcanism, rate of arc uplift and regional wind patterns. Because of the highly varied nature of modern ABAB sediments, and the lack of a direct link between sediment type and tectonic setting, one cannot assign a distinct sediment assemblage to ABAB. It is only when a relatively complete stratigraphic succession is preserved, and that detailed sedimentological and geochemical data are available, that ancient ABAB successions may be identified.

Remnant Arcs

Remnant arcs are submarine ridges that represent extinct portions of arcs that have been rifted away by the opening of a back-arc basin. They are composed chiefly of subaqueous volcanic rocks similar to those formed in submarine arcs. Once isolated by rifting, remnant arcs subside and are blanketed by progressively deeper water pelagic and biogenic deposits and distal ash showers.

Inactive Back-arc Basins

Inactive back-arc basins (IBAB), such as the western part of the Philippine plate (Plate I), are greater in age, have a thicker pelagic sediment blanket, and lack evidence for recent seafloor spreading, when compared to ABAB. These basins, which form by an early period of arc rifting, have similar lithologic assemblages to ABAB.

Retroarc Foreland Basins

Retroarc foreland basins form behind continental-margin arc systems (fig. 6.2a). They are filled largely with clastic terrigenous sediments derived from the *fold-thrust belt* behind the arc. A key element in foreland basin development is its syntectonic character (Graham et al. 1986). The greatest thickness of foreland basin sediment borders the fold-thrust belt reflecting enhanced subsidence caused by thrust-sheet loading and deposition of sediments. Another characteristic feature of retroarc foreland basins is that the proximal basin margin progressively becomes involved with the propagating fold-thrust belt (fig. 6.4). Sediments shed from the rising

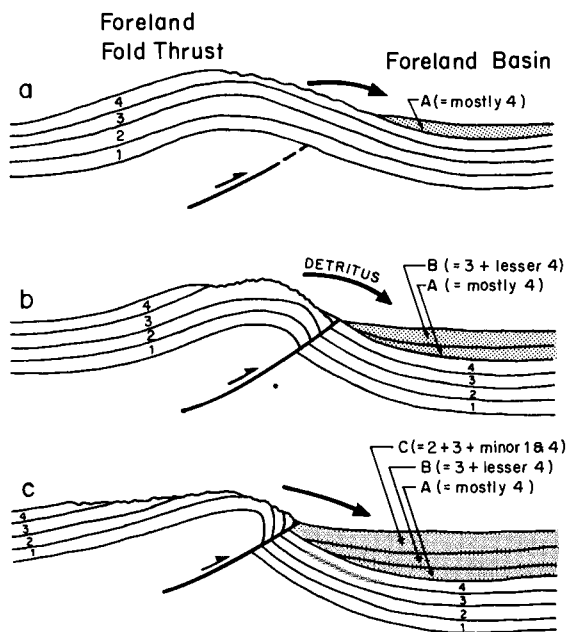


FIG 6.4 Progressive unroofing of an advancing foreland thrust sheet. After Graham et al. (1986).

fold-thrust belt are eroded and redeposited in the foreland basin only to be recycled again with basinward propagation of this belt. Proximal regions of foreland basins are characterized by coarse, arkosic alluvial fan sediments, and distal facies by fine-grained sediments and variable amounts of marine carbonates. Progressive unroofing in the fold-thrust belt should lead to an 'inverse' stratigraphic sampling of the source in foreland basin sediments as illustrated in figure 6.4. Such a pattern is well developed in the Cretaceous foreland basin deposits in eastern Utah (Lawton, 1986). In this basin, early stages of uplift and erosion resulted in deposition of Paleozoic carbonate-rich clastic sediments followed later by quartz-feldspar-rich detritus from the elevated Precambrian basement. Foreland basin successions are also characterized by upward coarsening and thickening terrigenous sediments, a feature which reflects progressive propagation of the fold-thrust belt into the basin.

Continental Rift Assemblages

Continental rifts may be single or multiple, as in the case of the Basin and Range Province. Also included in this category are aulacogens and continental hotspots such as Yellowstone Park in Wyoming. Continental rifts are characterized by immature terrigenous clastic sediments and bimodal volcanics. The sediments are chiefly arkoses, feldspathic quartzites, and conglomerates derived from rapidly uplifted fault blocks composed of granitic and quartzo-feldspathic metamorphic rocks. Evaporites also characterize the early stages of rift formation. If a rift is inundated with seawater, as exemplified by the Rhine graben, marine sandstones, shales, and carbonates may also be deposited. Bimodal volcanics are generally

subaerial basalt flows and associated rhyolites and trachytes. The felsic volcanics are emplaced either as ash flow tuffs or domes. In some rifts, alkali basalt and phonolite end members are characteristic, and calc-alkaline volcanics are important in only a few rifts. Deeply eroded rifts expose large volumes of granitic rocks, as for instance found in the Permian Oslo graben in Norway. Granitic rocks range from granite to monzodiorite in composition with monzonites and syenites usually dominating (Williams, 1982). Rift assemblages may also be deposited in rifted basins associated with arc systems, collisional boundaries, and faulted continental margins if the sources are eroded deep enough such that provenance is chiefly granitic rocks. Hence, the rift assemblage must be used in conjunction with other geologic data to accurately identify ancient continental rifts.

Craton Assemblages

Craton assemblages are composed chiefly of mature clastic sediments (quartzites and shales) and shallow marine carbonates. They occur in three tectonic settings; (1) passive continental margins, (2) platform basins and shelves, and (3) cratonic margins of back-arc basins. Passive continental margins represent an advanced stage of craton rifting, and hence the rift assemblages described above generally underlie passive continental margin successions. Cratonic assemblages may interfinger with arc successions in back-arc basins, such as in the Sea of Japan. Cratonic sandstones are relatively pure quartzose sands reflecting intense weathering, low relief in source areas, and prolonged transport across subdued continental surfaces. Marine carbonates are deposited as blankets and as bioherms and biostromes along stable-basin margins.

Depositional systems in cratonic successions vary depending on the relative roles of fluvial, eolian, deltaic, wave, storm, and tidal processes. Spatial and temporal distribution of sediments is controlled by epeirogenic uplift, the area over which shallow seas cover cratons, and climate (Klein, 1982). If tectonic uplift is important during deposition, continental shelves are narrower and sedimentation is dominated by wave and storm systems. However, if uplift is confined chiefly to craton margins, sediment yield increases into the craton and fluvial and deltaic systems may dominate. For transgressive marine clastic sequences, shallow seas are extensive and subtidal, storm-dominated, and wave-dominated environments are important. During regression, fluvial and eolian depositional systems become dominant.

Collision-related Assemblages

Collision-related assemblages develop along collisional boundaries where continental crust resists subduction (fig. 6.5). Rock assemblages change both with time and space as the collision progresses and along strike. Some sediments accumulate in *peripheral foreland basins* which develop in response to uplift and erosion of a collisional zone. These basins and the sediments therein evolve in a manner similar to retroarc foreland basins. Classic examples of peripheral foreland basins developed adjacent to the Alps and the Himalayas during the Alpine–Himalayan collisions in the Tertiary (Chap. 8). During the Alpine collision, up to 6 km of alluvial fan deposits known as *molasse* were deposited in foreland basins (Homewood et al.,

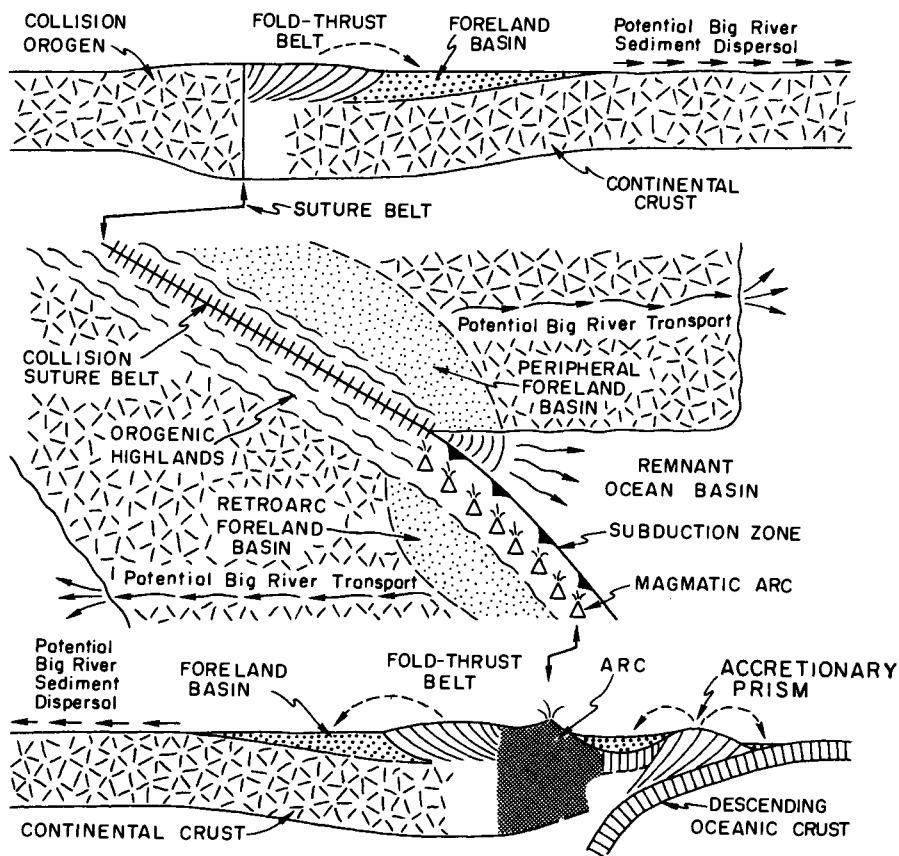


FIG 6.5 Cross-sections and plan view of an evolving collisional boundary. Modified after Dickinson and Suczek (1979). Dashed arrows indicate sediment transport directions.

1986). Individual alluvial fans up to 1 km thick and 40 km wide have been recognized in the Alps. Coarsening upward cycles and intraformational unconformities characterize collisional molasse deposits, both of which reflect uplift of the orogen and propagation of thrusts and nappes into foreland basins. Collision-derived sediment also may be shed longitudinally from the orogen and enter remnant ocean basins as turbidite fans (fig. 6.5). Sediments and volcanics may accumulate in fault-bounded basins in the thickened overriding continental plate during a continental collision. These are generally similar to rift assemblages with abundant felsic ash flow tuffs.

At deeper exposure levels (10–20 km) in collisional orogens, leucogranites are common and appear to be produced by partial melting of the crust during a collision (Chapt. 7). Thickening of continental crust, both in descending and overriding plates, leads to the production of granulites at depths >20 km. Anorthosites may form as cumulates from fractional crystallization of basalt in the lower and middle crust. Basaltic magma also may underplate the crust, and occurs as gabbro or mafic granulites in uplifted crustal sections (Chapt. 4).

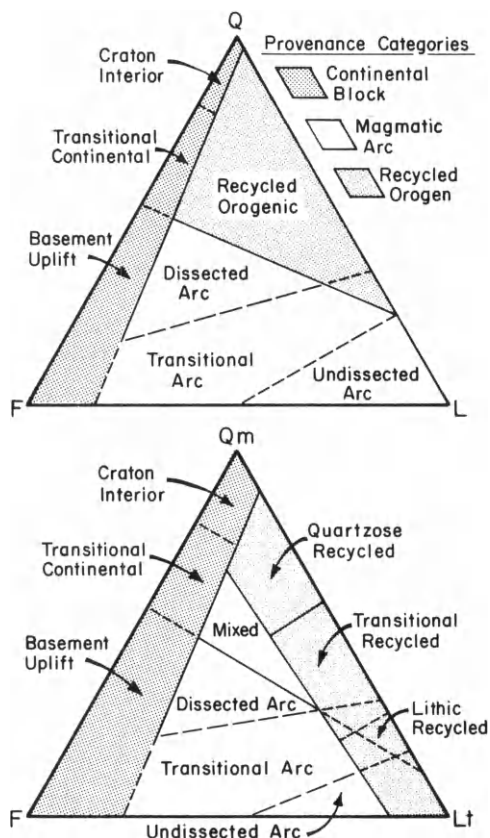


FIG 6.6 QFL and QmFLt detrital mode diagrams for matrix-poor sandstones in relation to tectonic setting (after Dickinson et al., 1983). Grain types: Q, quartzose; Qm, monocrystalline quartz; F, feldspar; L, lithic fragments; Lt, lithic fragments including chert.

Provenance of Clastic Sediments

The mineralogical and chemical compositions of clastic sediments, and especially of sandstones, are sensitive to provenance and tectonic setting. However, because other factors such as weathering, diagenesis, climate, and hydraulic sorting also affect sediment composition, it may be difficult to sort out the effect of tectonic setting. Provided the relative effects of non-provenance contributions to sediment composition can be minimized, detrital modes and both major and trace element distributions can be useful in constraining the tectonic setting of ancient clastic sediments. An example is the detrital modes of matrix-poor sandstones as suggested by Dickinson and Suczek (1979) (fig. 6.6). Sandstones from cratons are high in quartz (polycrystalline and monocrystalline) and those from continental rifts are relatively high in feldspars (chiefly K-feldspar). Arc sandstones contain a greater proportion of lithic fragments increasing in quartz and feldspar content with depth of arc dissection. Recycled orogens, which include foreland basins and some accretionary prisms and continental rifts, contain sandstones with varied detrital components, and do not always fall within the recycled orogen field (Schwab, 1986).

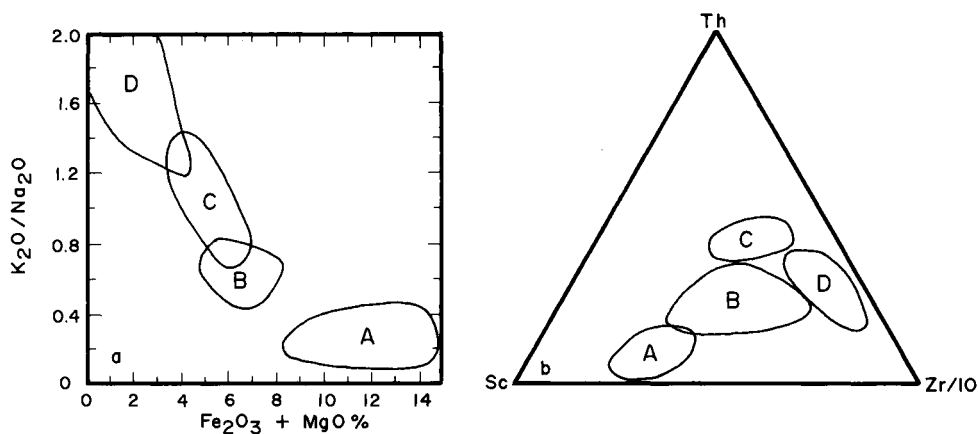


FIG 6.7 Examples of major element (a) and trace element (b) tectonic discriminant diagrams for sandstones. Modified after Bhatia (1983) and Bhatia and Crook (1986). A, Oceanic island arc; B, continental margin arc; C, foreland basins; D, passive continental margin.

Major and trace element compositions of sandstones are also sensitive to provenance and tectonic setting. Island-arc sandstones are characterized by relatively high $Fe_2O_3 + MgO$ and TiO_2 and low Al_2O_3/SiO_2 and K_2O/Na_2O ratios, whereas passive-margin sandstones show the opposite extremes of these oxides and ratios (Bhatia, 1983) (fig. 6.7a). Sandstones from continental margin arcs and foreland basins fall in between the extremes. On a K_2O/Na_2O versus SiO_2 diagram, arc sediments have relatively low K_2O/Na_2O and SiO_2 , passive-margin sediments have high values and other sediments fall at intermediate positions (Roser and Korsch, 1986). Graywackes from different tectonic settings can be distinguished by the distributions of relatively immobile trace elements such as La, Th, Zr, Nb, Y, Sc, and Co (Bhatia and Crook, 1986). Arc graywackes are relatively high in Co and Sc and passive-margin graywackes are very low in these elements. The Th–Sc–Zr diagram is particularly useful in separating graywacke tectonic groups (fig. 6.7b). The geochemical trends in sandstones reflect the input of basalt and basaltic andesite (high Fe, Mg, Co, Sc) with oceanic arcs at one extreme and passive-margin sandstones (high Al_2O_3/SiO_2 , K_2O/Na_2O , Zr, Th) at the other.

Tectonics and Metamorphism

Modelling of the thermal history of the crust during and after periods of deformation has reinforced the view that regional metamorphism and orogeny are related (Thompson and Ridley 1987). Details of how metamorphism is related to specific deformational events, however, are not well understood. Metamorphic textures and the distribution of metamorphic isograds in a given geographic area are controlled by small-scale deformational patterns that are not easily related to overall patterns in orogenic belts. However, comparison of P–T regimes in orogenic belts with calculated P–T–t paths (Chapt. 4) suggests that facies-series of regional metamorphism may be characteristic of tectonic settings. For instance, blueschists are characteristic of subduction zones and appear to form in accretionary prisms where

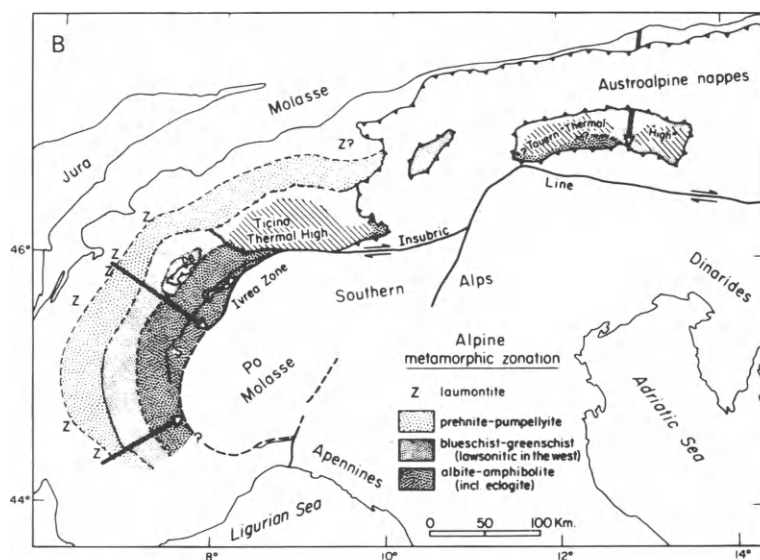


FIG 6.8 Paired metamorphic belts in the Alps. After Ernst (1971). Arrows indicate direction of former subduction.

rocks are subjected to high stress pressures at relatively low temperatures. *Paired metamorphic belts*, which are juxtaposed belts of blueschist-greenschist and amphibolite-facies terranes, are also characteristic of uplifted and eroded portions of forearc regions (Miyashiro, 1973). Low-grade belts occur on what was the seaward side of an arc system as illustrated by the paired Alpine belts in Switzerland (fig. 6.8). The mechanism by which metamorphic belts in uplifted arcs are juxtaposed is poorly known. It may involve buoyant rebound in which slabs of the accretionary prism that are dragged down farthest (to higher P–T regimes) along the top of the descending slab, isostatically rebound the most.

P–T paths in collisional terranes are complex and can vary widely within a single metamorphic belt. This results from advection of heat by folding and thrusting, conductive heat transfer during and after deformation, and the relatively short times of thermal relaxation ($<10^6$ a) within folds and thrusts (Chamberlain and Karabinos, 1987). Hence, no single P–T–t path seems to characterize collisional orogens. The relationship between thermal and deformational features in continental rifts is also poorly understood, but is thought to be characterized by near-isobaric cooling for at least some of the P–T history. Many more points need to be determined on P–T–t paths before a characteristic sets of paths can be formulated for various tectonic settings.

Ophiolites

Oceanic environments that may be reflected in ophiolites include mid-ocean ridges, back-arc basins, and immature island arcs. However, compared to seismic sections of oceanic or arc crust, ophiolites are considerably thinner (chiefly <5 km compared to 7–10 km for oceanic crust and 10–15 km for arcs). In some highly deformed

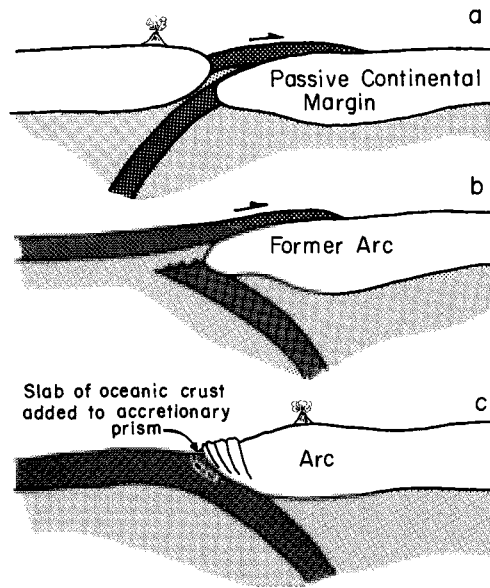


FIG 6.9 Three mechanisms of ophiolite emplacement. (a) Obduction at a passive continental margin, (b) obduction of a flake of oceanic lithosphere, and (c) addition of slab of oceanic lithosphere to an accretionary prism.

ophiolites the thickness may be controlled by tectonic thinning during emplacement. Most ophiolites, however, are not severely deformed and may have been thickened during emplacement (Moore, 1982). Hence, any model for the origin and emplacement of ophiolites must account for these large discrepancies in thickness compared to oceanic or immature arc crust.

Ophiolites appear to be emplaced in arcs or collisional orogens by three major mechanisms: (1) *obduction* or thrusting of oceanic lithosphere onto a passive continental margin during a continental collision; (2) splitting of the upper part of the descending slab and obduction of a flake onto the continent; and (3) addition of a slab of oceanic crust to an accretionary prism in an arc system (fig. 6.9). Although gravity sliding has also been proposed as a way to emplace ophiolites, this requires both an elevated source area and a slope sufficient to allow sliding of 5 km-thick slabs of oceanic crust onto the continent, a highly unlikely situation. The fact that basalts from many and perhaps most ophiolites have a subduction zone geochemical signature (Chapt. 7) suggests that they represent fragments of back-arc oceanic crust and perhaps, in a few instances, immature arc crust. Ophiolites that are fragments of true mid-ocean ridges, such as the Semail ophiolite in Oman, appear to be few in number.

The metamorphic complex at the base of ophiolites (fig. 6.1) may play a major role in ophiolite emplacement (Spray, 1984). These metamorphic soles, as they are often called, have many features in common: (1) thicknesses generally range from 10 to 500 m; (2) they extend laterally for tens to hundreds of kilometers; (3) most show a sharp decrease in metamorphic grade from top to bottom; (4) they are highly deformed and have a pronounced tectonic foliation; and (5) they are composed of mixed amphibolites (originally basalt) and metasediments. These data are consistent

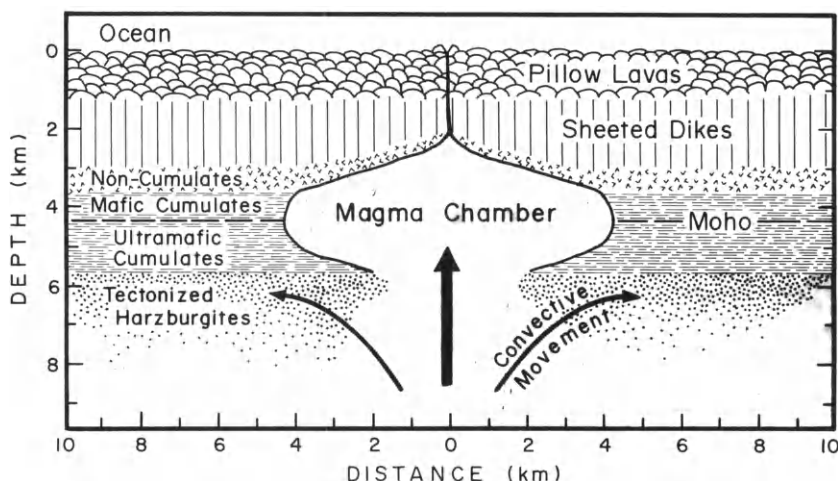


FIG 6.10 Schematic cross-section of oceanic crust near an ocean ridge showing possible relationships to an ophiolite succession. Modified after Gass (1982).

with an origin for soles by successive underplating and welding onto the base to an ophiolite as it moves from the upper mantle to the surface. Metamorphic facies boundaries within soles may represent tectonic contacts. The heat source for sole metamorphism is probably a combination of residual heat from the mantle and shear heating during ophiolite emplacement. If this interpretation is correct, ophiolites may represent tectonically decoupled slices of hot oceanic lithosphere. The fact that ophiolites are thin compared to normal oceanic crust may be explained if they are decoupled from the mantle at the asthenosphere–lithosphere boundary in the vicinity of spreading centers where oceanic lithosphere is thin. In a closing ocean basin, lithosphere decoupling may occur when an ocean ridge enters a subduction zone or during a continent–continent collision. If decoupling occurs only at the asthenosphere–lithosphere boundary it may not be possible to obduct thicker segments of oceanic lithosphere (up to 150 km thick at the time of subduction), thus accounting for the absence of ophiolites much thicker than 5 km.

The mechanism by which a complete ophiolite succession forms is not well understood. A possible model is illustrated in figure 6.10. As pressure decreases in a convective upcurrent, garnet lherzolite partially melts to produce ocean-ridge basalt (see Chapt. 7). These magmas collect in shallow chambers (3–6 km deep) and undergo fractional crystallization. Layered ultramafic and gabbroic rocks accumulate in the magma chamber forming the cumulate section of ophiolites. The tectonized harzburgites represent residue left after melting, which is highly deformed and sheared due to lateral advective motion of the asthenosphere. Dikes are ejected from the magma chamber forming the sheeted dike complex, and many are erupted forming pillow lavas in the axial rift. Fractional crystallization in the magma chamber also gives rise locally to diorites and plagiogranites which are emplaced at the top of the layered cumulate series.

Arc Systems

An important question related to the origin of arc systems is that of how and where new subduction zones begin. Because of the very high stress levels necessary for the

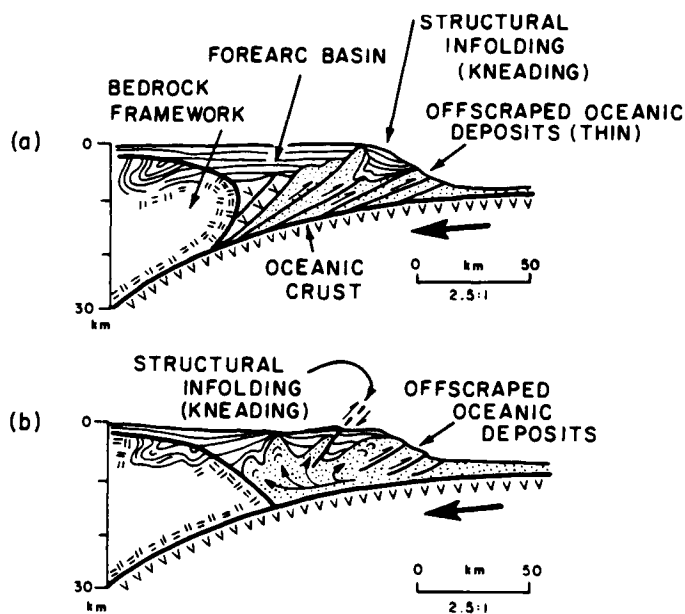


FIG 6.11 Accretion and kneading in an arc system. After Scholl et al. (1980).

oceanic lithosphere to rupture it is likely that pre-existing zones of weakness in the lithosphere provide sites for new subduction zones. A survey of young arcs indicates that both interfaces of crustal types and extinct or active plate boundaries are zones of weakness where new subduction zones have formed (Karig, 1982). Passive continental margins adjacent to either large ocean basins or back-arc basins are possible sites of new subduction. Not only is it likely that the lithosphere is weak at such interfaces, but the ocean basin is generally downflexed. Subduction can also be initiated along extinct convergent margins and ocean ridges or fracture zones, which are zones of lithosphere weakness, and changes in plate motions can change an active transform fault into a subduction zone.

Seismic reflection profiling and geological studies of uplifted and eroded arc systems have led to a greater understanding of arc evolution and of the development of accretionary prisms and forearc basins. Widths of modern arc-trench gaps (75–250 km) are proportional to the ages of the oldest igneous rocks exposed in adjacent arcs (Dickinson, 1973). As examples, the arc-trench gap width in the Solomon Islands is about 50 km, with the oldest igneous rocks about 25 Ma, and the arc-trench gap width in northern Japan (Honshu) is about 225 km, with the oldest igneous rocks about 125 Ma. The correlation suggests progressive growth in the width of arc-trench gaps with time. Such growth appears to reflect some combination of outward migration of the subduction zone by accretionary processes and inward migration of the zone of maximum magmatic activity. *Subduction zone accretion* involves the addition of masses of sediment ($\pm$ oceanic crust) to the margin of an arc or continent in the accretionary prism. Seismic profiling suggests that accretionary prisms are composed of sediment wedges separated by high-angle thrust faults produced by offscraping of oceanic sediments (fig. 6.11a). *Offscraping accretion* involves the tectonic skimming of oceanic sediments and of mafic and

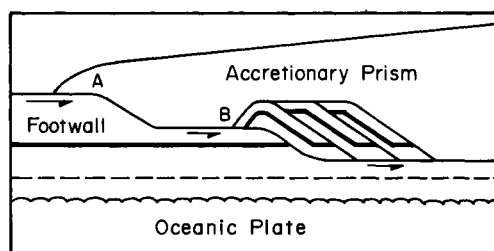


FIG 6.12 Idealized cross-section showing duplex accretion. After Sample and Fisher (1986). A, Zone of offscraping and B, zone of duplex accretion.

ultramafic oceanic lithosphere and their addition to the continental (or arc) bedrock framework (Scholl et al., 1980). Offscraping results in outward growth of the accretionary prism and also controls the location and evolutionary patterns of overlying forearc basins. Reflection profiles suggest more complex deformational patterns than simple thrust edges. Deformation may include large-scale structural mixing and infolding of forearc basin sediments, a process referred to as subduction kneading (fig. 6.11b). Geological evidence for subduction kneading comes from uplifted and exposed accretionary prisms, exemplified by parts of the Franciscan Complex in California.

One possible mechanism of accretion is duplex accretion. A *duplex* is an imbricate package of isolated thrust slices bounded on top by a thrust and below by a low-angle detachment fault (Sample and Fisher, 1986). During transfer of displacement from an upper to a lower detachment horizon, slices of footwall are accreted to the hanging wall (accretionary prism) and rotated by bending of the frontal ramp (fig. 6.12). Observations from seismic reflection profiles, as well as exposed accretionary prisms, indicate that duplex accretion occurs at greater depths than offscraping accretion.

Although some arcs, such as the Middle America and Sunda arcs, appear to have grown appreciably by accretionary processes, others such as the New Hebrides arc have little if any accretionary prism. In these latter arcs, either very little sediment is deposited in the trench or most sediments are subducted (fig. 6.13a). One possible way to subduct sediments is in grabens in the descending slab. If sediments are subducted in large amounts beneath some arcs, they cannot contribute substantially to arc magma production as constrained by isotopic and trace element distributions in arc volcanics (Chapt. 7). *Subduction erosion* is another process that has been proposed for arcs with insignificant accretionary prisms. It involves mechanical plucking and abrasion along the top of a descending slab, which causes a trench's landward slope to retreat shoreward (fig. 6.13b,c). Subduction erosion may occur either along the top of the descending slab (b) or at the leading edge of the overriding plate (c). Accretion, kneading, erosion, and sediment subduction are all potentially important processes in subduction zones, and any one of them may dominate at a given place and evolutionary stage.

Another question related to convergent plate margins is that of what happens when an ocean ridge collides with an arc. Most studies, which are largely of ancient collisions, conclude that ridges are subducted. Ridge crest subduction has been described beneath Japan, western North America (see Chapt. 8), the Antarctic

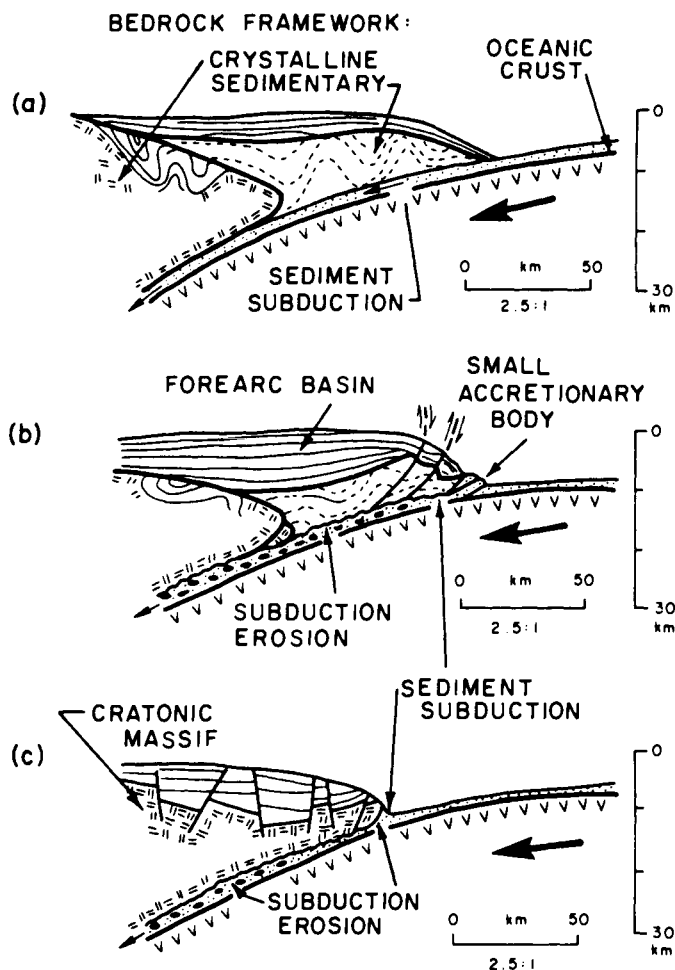


FIG 6.13 Sediment subduction and sediment erosion at a convergent plate boundary. After Scholl et al. (1980).

Peninsula and the Aleutian Islands. If a ridge is subducted the arc should move “uphill” and become emergent as the ridge crest approaches, and it should move “downhill” becoming submerged, as the ridge passes down the subduction zone (DeLong and Fox, 1977). Corresponding changes in sedimentation should accompany this emergence–submergence sequence of the arc. Ridge subduction may also lead to cessation of subduction-related magmatism as the hot ridge is subducted. This could be caused by reduced frictional heating in the subduction zone or progressive loss of volatiles from a descending slab as a ridge approaches. Also, the outer arc should undergo regional metamorphism as a hot ridge crest is subducted. All three of these phenomena are recorded in the Aleutian arc and hence support the subducted ridge model. Ridge subduction may also result in a change in stress regime in the overriding plate from dominantly compressional to extensional, and in the opening of a back-arc basin (Uyeda and Miyashiro, 1974). The subduction of active ridges may lead to formation of new ridges in the descending plate at great

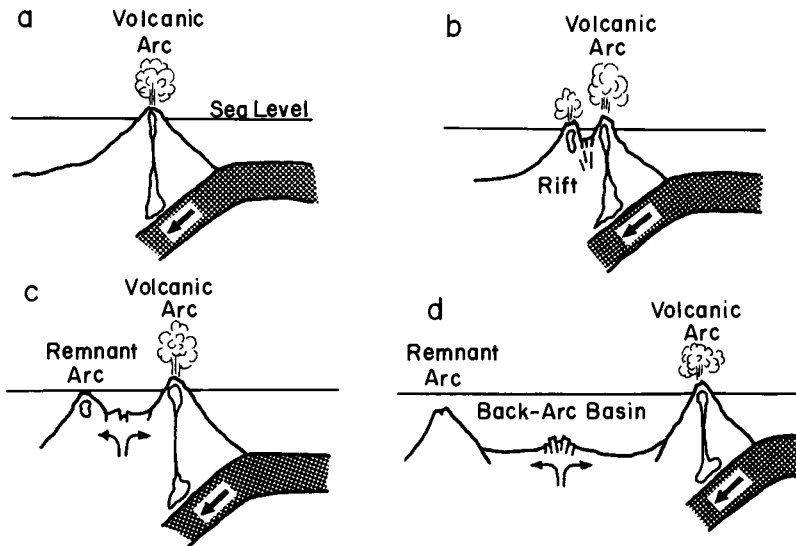


FIG 6.14 Generalized model for the evolution of a back-arc basin. After Karig (1971).

distances from the convergent margin. For instance, the rifting of Antarctica from Australia, which began almost 50 Ma ago, coincided with the subduction of a ridge system along the northern edge of the Australian–Antarctica plate.

The effects of subducting an active ocean ridge have been studied where the Chile ridge enters the Peru–Chile trench (Plate I). Surprisingly, very few effects of this collision are reflected by the Chile ridge (Cande et al., 1987). There is no measurable rotation of the ridge axis and only a change in spreading rates is observed. On the other hand, the continental margin responds both mechanically and thermally to the collision. The landward trench slope steepens and narrows near the collision zone, the accretionary prism is reduced greatly in size (by >75%) and heat flow increases. Two factors seem important in decreasing the volume of the accretionary prism: (1) the topographic relief on the ridge may increase the rate of subduction erosion carrying material away from the bottom of the accretionary prism, and (2) subduction of basement ridges caused by transform faults may mechanically weaken the base of the accretionary prism, making it more susceptible to removal by subduction erosion. The more intriguing question of what happens to the convective upcurrent beneath the Chile ridge when it is subducted remains unanswered.

Perhaps no other topic related to convergent plate boundaries has received more attention than that of the origin and evolution of back-arc basins. An idealized model for the opening of a back-arc basin is shown in fig. 6.14 based on studies in the Philippine plate (Karig, 1971). Beginning with a volcanic arc (a), rifting is initiated near the volcanic front as a result of the subduction process (b). The arc is split along zones of weakness such as magma conduit systems and steep normal faults that define boundaries of the rift. New oceanic crust is eventually produced in the rift as the remnant arc migrates away from the active arc (c). As the back-arc basin grows, the flanks of the newly formed ridge subside in a manner similar to mid-ocean ridge subsidence (d). This cycle may happen more than once for the same arc, resulting in the formation of several back-arc basins and remnant arcs as observed in the Philippine basin today.

TABLE 6.1 *Models for the Origin of Back-Arc Basins*

- I. *Active spreading*
 - (a) Rise of a diapir
 - (b) Induced convection in mantle wedge

- II. *Passive spreading*
 - (a) Retreat of descending plate
 - (1) slab anchored to asthenosphere
 - (2) gravitational collapse of slab
 - (3) buckling of spherical shell
 - b) Retreat of overriding plate
 - (1) slab anchored to stationary mantle
 - (2) collision of buoyant feature

With exception of those back-arc basins that represent entrapped fragments of previously existing ocean basins (such as the Bering Sea and the West Philippine basin), it is generally agreed that back-arc basins form by spreading of the seafloor (Uyeda, 1986). This is supported by dated magnetic anomalies in many back-arc basins. Spreading models are divided into active and passive categories (table 6.1). Two active spreading models have been described, the diapir and induced convection models. In the diapir model, extension is caused by injection and spreading of mantle diapirs that rise buoyantly from the subducted slab or lower part of the mantle wedge. The diapirs are produced by partial melting in or above the descending slab. In the induced convection model, spreading is caused by convective flow of the asthenosphere, which is induced by the subducting slab. This convection in the mantle wedge directs hot mantle flow at the base of the lithosphere behind the arc, resulting in opening of a back-arc basin. Active spreading models, however, are faced with several problems. In general, they fail to explain the temporal and spatial distributions of back-arc basins as well as the absence of back-arc basins behind most arc systems. The diapir model cannot readily account for the migration of back-arc basins away from an arc with time. The convection model, which predicts that back-arc spreading should be initiated at >100 km behind the junction of the subducted slab and the overlying lithosphere, cannot explain why back-arc rifting usually begins at the volcanic front almost directly above the slab-lithosphere junction.

Passive models for back-arc spreading involve progressive separation of descending and overriding plates by forces other than those associated with active spreading. If a subducted slab is anchored to the asthenosphere, back-arc spreading occurs when the overriding plate has a landward velocity component (V_O) relative to the rollback velocity (V_R), which is the retreat velocity of the trench hingeline (fig. 6.15). Supporting this model for some back-arc basins is the fact that in subduction zones without back-arc spreading, overriding plates have oceanward rather than landward velocity components. The same effect occurs if a slab gravitationally collapses (i.e. when $V_S > V_N$) or if a slab is anchored to stationary mantle and the overriding plate retreats. Collision of aseismic ridges or other buoyant features on the seafloor with a subduction zone has also been suggested as a mechanism to initiate back-arc spreading. The resistance of these features to subduction may cause extensional stresses in the arc resulting in rifting. Still another possible passive model involves the retreat of the trench hingeline accompanying buckling of a

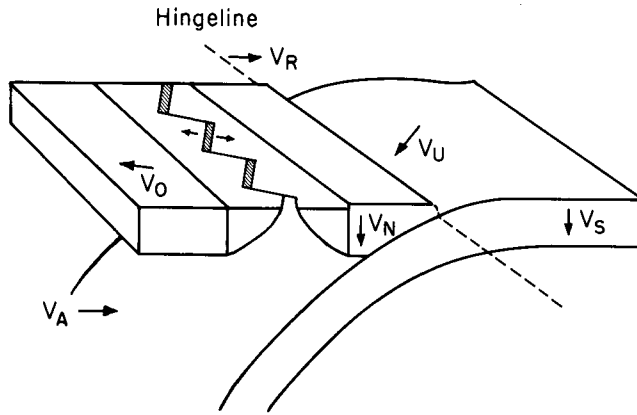


FIG 6.15 Velocity components acting at a convergent plate margin. After Uyeda (1983). Velocity symbols: V_O , overriding plate; V_U , descending slab; V_R hingeline rollback; V_N sinking of overriding plate; V_S sinking of descending plate.

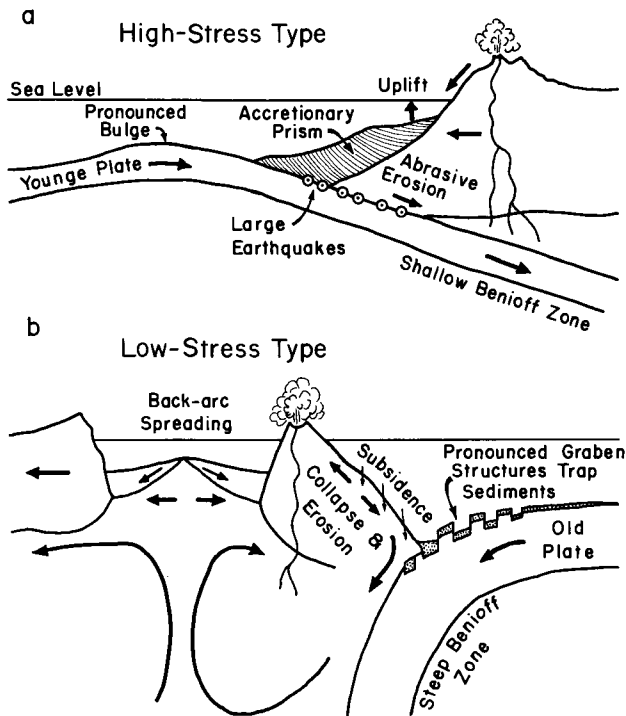


FIG 6.16 Idealized cross-sections of high- and low-stress subduction zones. After Uyeda (1983).

spherical shell lithosphere. On a spherical body such as the Earth, the mode of buckling of a descending slab changes along strike so that as the arc lengthens, it becomes more arcuate in plan view and back-arc basins may develop (Yamaoka et al., 1986.)

Uyeda (1983) has suggested that subduction zones are of two major types, each representing an end member in a continuum of types (fig. 6.16). The relatively high-

stress type, as exemplified by the Peru–Chile arc, is characterized by a pronounced bulge in the descending slab, a large accretionary prism, relatively large shallow earthquakes, a shallow dip of the subduction zone or buoyant subduction, a relatively young descending slab, and a wide range in composition of calc-alkaline and tholeiitic igneous rocks (a). The low-stress type, of which the Mariana arc is an example, has little or no accretionary prism, few large earthquakes, a steep dip of the descending plate which is relatively old, a dominance of basaltic igneous rocks and a back-arc basin (b). In the high-stress type, the descending and overriding plates are more strongly coupled than in the low-stress type, explaining the importance of large earthquakes and accretionary growth. This stronger coupling, in turn, may result from a young relatively warm plate subducting at a shallow angle. In the low-stress type the overriding plate is retreating from the descending plate, resulting in the opening of a back-arc basin. In the high-stress type, however, the overriding plate is either retreating very slowly compared to the descending plate, or perhaps converging against the descending plate. Thus, the two major factors contributing to differences in subduction zones appear to be: (1) relative motions of descending and overriding plates, and (2) the age and temperature of the descending plates.

Continental Rifts

Introduction

Rift systems may be narrow and long, such as the East African system; narrow and short, such as the Baikal and Rhine rifts; or very broad multiple rifts, such as the Basin and Range Province. Those that form in cratons, such as the East African rift system, are commonly associated with domal uplifts, although the timing of doming relative to rifting may vary (Mohr, 1982). Geophysical data indicate that both the crust and lithosphere are thinned beneath rifts and that most or all of the crustal thinning occurs in the lower crust (Chapt. 4). Rifts are characterized by high heat flow, generally regarded to reflect hot asthenosphere at relatively shallow depths. The amount of separation of young rifts can be estimated from offset on faults and from gravity profiles. Results range from as little as 10 km for the Baikal rift to ≥ 50 km for the Rio Grande rift. Some rifts, such as the Kenya rift, have a prominent gravity anomaly running along the axis, generally interpreted to reflect emplacement of dikes or sills at shallow depth. Magma chambers have been described in some rifts based on the seismic data of which the best-documented cases are magma chambers in central New Mexico beneath the Rio Grande rift and in Iceland (Sanford and Einarsson, 1982). Although in some instances tectonic and volcanic activity in rifts are related in space and time, in many cases they are not. Zones of major eruption seldom coincide with the main rift faults.

Rifts range from those with abundant volcanics, like the East African rift, to those like the Baikal rift with only minor volcanics. Igneous rocks in rifts are often bimodal in composition and may be alkaline, as in the case of the East African rift and Oslo graben, or tholeiitic, as in the northern Rio Grande rift. In a few cases they are calc-alkaline. Alkali basalts, basanites, and tephrites dominate in the East African rift followed in abundance by phonolites and trachytes (Williams, 1982). In the northern Rio Grande rift, tholeiites and basaltic andesites are most important followed by rhyolite domes. Farther south in the Rio Grande rift alkali basalts dominate. The

Kenya and Oslo rifts show a decrease in magma alkalinity with time, a feature that probably reflects a decrease in depth of magma generation. Many rifts show an increase in magma alkalinity from the center to the edge of the rift (East African, Rio Grande) suggesting a deepening of magma sources in this direction.

Although continental rifts may form in a number of tectonic settings, they commonly exhibit similar overall patterns of development. The major stages are as follows (after Mohr, 1982):

1. Development of a broad, shallow depression prior to doming or volcanism.
2. Diapiric asthenosphere is forcefully injected into the lithosphere. During injection, diapirs undergo adiabatic decompression leading to partial melting and the onset of magmatism.
3. Buoyant isostatic uplift of heated lithosphere leads to doming and it may occur episodically.
4. Doming and extensional forces in the lithosphere cause crustal attenuation and thinning of the lithosphere.
5. Episodic dike injection and volcanism alternate with faulting, and the duration of episodes and volumes of magma erupted decreases with time.
6. Rift valleys develop and may be associated with voluminous felsic magmatism, both extrusive and intrusive.
7. A rift may be aborted at a relatively early stage of its development, such as the late Proterozoic Keweenawan rift in the north-central United States (see Chapt. 9), or it may continue to open and evolve into an ocean basin, such as the Red Sea.

Toward a Classification of Rifts

Rifts are of different origins and occur in different regional tectonic settings. Although the immediate stress environment of rifts is extensional, the regional stress environment may be compressional, extensional, or nearly neutral.

As a starting point for understanding rift development and significance, rifts can be classified into two categories, depending on the mechanisms of rifting (Sengor and Burke, 1978; Condie, 1981b). One group of rifts, referred to as *active rifts*, are produced by doming and cracking of the lithosphere, where doming results from upwelling asthenosphere or rising mantle plumes (fig. 6.17). Another group of rifts, *passive rifts*, are produced by stresses in moving lithospheric plates. Passive rifts may evolve into active rifts if a large volume of asthenosphere upwells and spreads beneath a rift. Active rifts are characterized by relatively large volumes of volcanic rock, while in passive rifts, immature clastic sediments generally exceed volcanics in abundance. Active rifts are also characterized by early uplift and basement stripping resulting from crustal expansion due to a deep heat source. In general, uplift in passive rifts is confined to the stretched and faulted near-surface region and to the shoulder of the rift zone (fig. 6.17). In active rifts, uplift commonly extends for hundreds of kilometers beyond the rift zone proper. Lithospheric thinning is laterally confined to the rift zone in passive rifts, while in active rifts the zone of thinning is several times wider than the rift width.

Active and passive rifts can be subdivided into three categories (table 6.2). Active

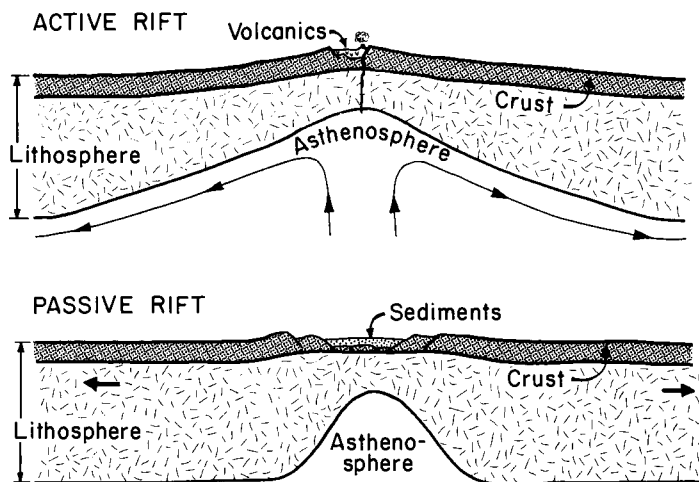


FIG 6.17 Diagrammatic cross-sections of active and passive continental rifts.

rifts are represented by ocean ridges, continental rifts and aulacogens, and back-arc basins. Both of the latter two types of rifts may evolve into the former. *Continental rifts* develop in orogenic belts or cratons, as exemplified by the Rio Grande rift and the East African rift system, respectively (Plate I). Aulacogens are failed or less-active arms of triple junctions, such as the Ethiopian and Benue rifts in Africa (Plate I). Spreading began in the Red Sea and Gulf of Aden 15–25 Ma ago, and may be beginning today in the Ethiopian rift. The Benue rift became a failed arm in the Late Cretaceous as the other two rift segments opened as part of the Atlantic basin.

A possible sequence in the break-up of a continent by continued spreading of a continental rift is shown in fig. 6.18. This history is based on the opening of the Red Sea beginning about 30 Ma ago. A mantle plume or upwelling convection current causes the lithosphere to thin and fracture into a series of grabens, which collect clastic sediments from intervening horsts (A,B), and basaltic magma is injected into the axial portion of the graben system. Eventually new oceanic crust (and lithosphere) is produced as the continent further separates and an ocean ridge system forms (C). Remnants of basaltic flows and sills, and clastic sediments reflecting early stages of rupturing may be preserved in grabens on retreating continental margins (D).

Passive rifts develop along faulted continental margins, in zones of continental collision, and perhaps in arc systems. Examples of rifts associated with faulted continental margins are those found adjacent to the San Andreas and related faults in California, and the multiple-rift system in western Turkey. Rifts can be generated along collision boundaries by irregularities in continental margins and by strike-slip and normal faulting caused by nonperpendicular collision. The Rhine graben in Germany has been suggested as an example of a rift that developed at a steep angle to a collisional boundary, and the term *impactogen* has been applied to this type of rift. Rifting may have occurred in the European continent as Africa collided with an irregularity along the European continental margin (fig. 6.19). An aborted attempt to subduce the irregularity resulted in depression and rifting. Molnar and Tapponnier (1975) have suggested that stresses associated with the Tibet–India collision (Chapt.

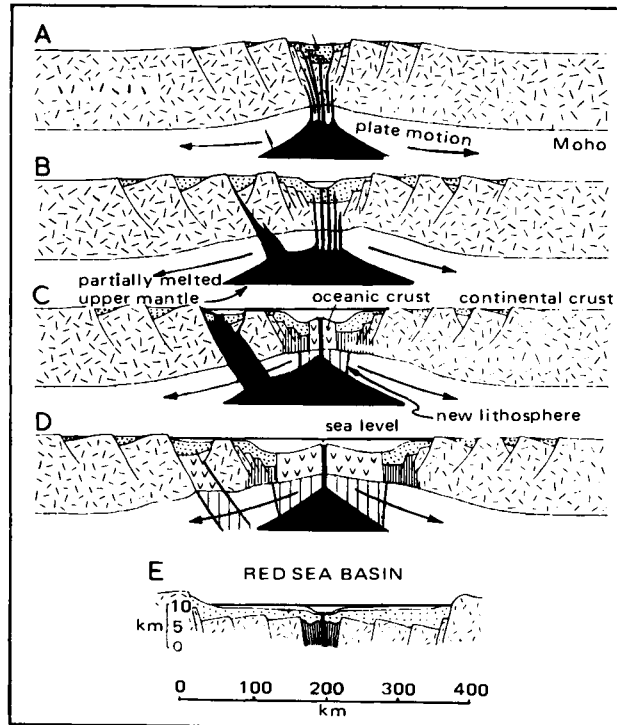


FIG 6.18 Stages in the rupture and fragmentation of a continent and development of a new ocean ridge. From Dewey and Bird (1970b).

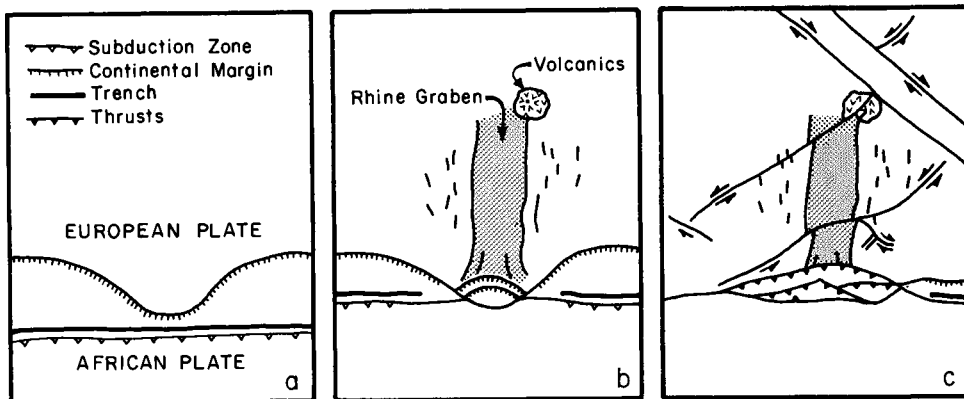
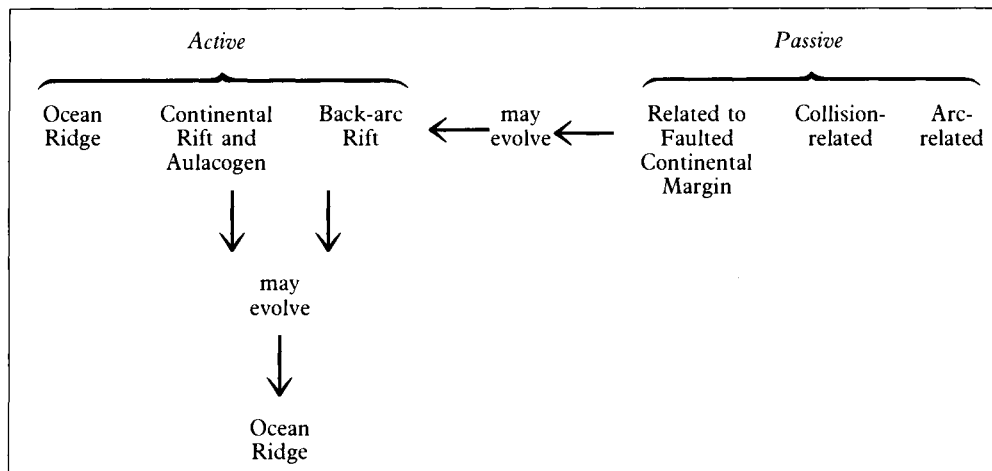


FIG 6.19 Idealized sequence of events leading to development of the Rhine graben. After Sengor (1976). (a) Convergence of Africa and Europe; (b) collision of irregularity on European margin with Africa and beginning of rift development; (c) suturing of continents complete, deformation characterized by strike-slip faulting.

TABLE 6.2 *Rift Classification*

8) were transmitted as strike-slip and normal faulting into Eurasia, forming such rifts as the Baikal rift in Siberia and the Shanshi graben system in China.

Various petrotectonic assemblages characterize rifts. Oceanic rifts are represented by ophiolites. Small arc-related rifts contain arc volcanics and graywackes, and moderate-to-large back-arc basins are characterized by mixed assemblages including some combination of ophiolites and deep-sea sediments, arc-derived graywackes, and cratonic sediments. Continental rifts and aulacogens contain arkoses, feldspathic sandstones, conglomerates, and bimodal volcanics. Passive rifts contain a variety of sediments, and in some instances minor volcanic rocks; immature terrigenous sediments generally dominate. As rifts evolve (table 6.2), relative proportions as well as kinds of rocks produced also change.

Mechanisms of Rifting

Two major mechanisms have been proposed for continental rifting as mentioned above: upwelling or diapirism of the mantle and stresses produced by moving plates. Both involve extension of the lithosphere and crust. Just how this extension occurs, and the relative roles of brittle and ductile deformation continue to be subjects of disagreement among geologists. In general, two models have been proposed for rifting in the crust. One model consists of multiple horsts and grabens which form during brittle extension of the upper crust. These blocks sink differentially into a plastically deforming middle and lower crust, where they become isostatically adjusted. The second, and currently more popular model, involves curved, downward-flattening *listric faults* which exhibit moderate to extreme rotation of hanging-wall blocks. Seismic reflection profiling in the Basin and Range Province strongly favors the existence of listric faults at mid-crustal levels. These faults appear to merge with, or be truncated by, a flat decollement or detachment fault that developed during extension. *Detachment faults* are major flat-lying shear zones separating disrupted, tilted blocks of an upper plate from ductilely deformed mylonitic rocks of a lower plate (fig. 6.20). Up to 100% crustal extension may occur with the combined offsets of listric and detachment faults. In addition to curving

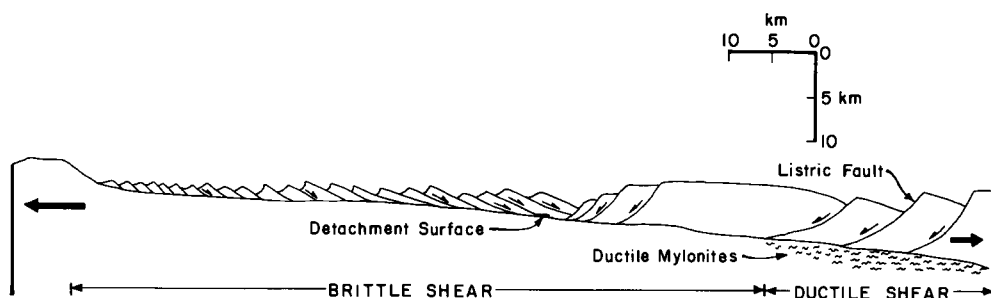


FIG 6.20 Idealized cross-section showing the major components related to a detachment fault.

listric faults, sets of near-planar faults may also accommodate extension as these faults rotate and flatten like a stack of tilted books in a bookcase (Rehrig, 1986).

In detail, detachment faults are not simple or single surfaces of displacement. They are generally multiple surfaces where exposed, with slivers of brecciated rock between surfaces. Detachment faults commonly bound upper-plate rocks (often Tertiary volcanics and sediments) that display chaotic tilting generally in one direction with consistently oriented rotational axes. Upper-plate extensional structures initiated at higher angles have been progressively flattened, whereas detachment faults have always been nearly horizontal. The observations are generally interpreted to indicate that a detachment fault is the lower decoupling surface upon which the upper plate breaks up and is transported (fig. 6.20).

Cratonic Basins

The rates of subsidence and uplift in cratons are a function of the time interval over which they are measured. Estimates of current rates are of the order of a few centimeters per year, while data from older Phanerozoic successions suggest rates 1–2 orders of magnitude slower. In general, Phanerozoic rates of uplift appear to have been 0.1–1 cm/a over periods of 10^4 – 10^5 years and over areas of 10^4 – 10^6 km². One of the most significant observations of cratonic basins is that they exhibit the same exponential subsidence as ocean basins (Sleep, 1971). Their subsidence can be considered in terms of two stages. In the first stage subsidence rate varies greatly, whereas the second stage is characterized by widespread, general subsidence. After about 50 Ma the depth of subsidence decreases exponentially to a constant value.

Several models have been suggested to explain cratonic subsidence (Bott, 1979; Sleep et al., 1980). Sediment loading, lithosphere stretching, and thermal doming followed by contraction are the most widely cited mechanisms. Although the accumulation of sediments in a depression loads the lithosphere and causes further subsidence, calculations indicate that the contribution of sediment loading to subsidence must be minor compared to other effects. Subsidence at passive continental margins may result from thinning of continental crust by progressive creep towards the suboceanic upper mantle of a relatively ductile lower crust. As the crust thins, sediments accumulate in overlying basins. Alternately, the lithosphere may be domed by upwelling asthenosphere or a mantle plume, and uplifted crust is eroded. Thermal contraction following doming results in platform basins, or a series

of marginal basins around an opening ocean, which fills with sediments. Most investigators now agree that subsidence along passive continental margins is due to the combined effects of thermal contraction of continental and adjacent oceanic lithosphere together with sediment loading.

Collisional Orogens

Introduction

Until plate tectonics, the classical geotectonic cycle encompassed a progressive sequence of events related to the geosynclinal theory of mountain building. In this model, elongate, fixed belts of subsidence and sedimentation known as geosynclines are necessary precursors of orogeny (Coney, 1970). A typical orogenic cycle involved accumulation of sediments ($\pm$ volcanic rocks) and intense deformation and plutonism followed by differential uplift and volcanism. Such a simple sequence of events is rarely, if ever, documented in an orogenic belt.

Orogeny occurs along convergent plate boundaries. Unlike the geosynclinal theory, plate tectonics does not require a regular sequence of events to occur in the same order in space or time in all orogenic belts. The sequence of events can vary significantly, depending upon the various possible interactions that may occur at convergent plate boundaries. Two general models have been suggested to relate plate tectonics to orogeny (Dickinson, 1971a). The *activation model* involves initiation of plate convergence at a passive continental margin, and the *collision model* involves an arc-continent, continent–continent or arc–arc collision (Ellis, 1988). The latter is characterized by cessation of subduction or relocation of a subduction zone along a nearby oceanic interface. Plate collision is well documented as the major process by which mountain chains are produced. Some investigators propose that all major mountain building involves collision. Whether any mountain belts are produced by activation orogeny involving only the subduction of normal oceanic lithosphere is a question of widespread interest and disagreement among geologists.

A schematic sequence of events in an orogeny caused by a continent–continent collision is illustrated in fig. 6.21a. As closure of an ocean basin occurs (A), slices of oceanic crust may be thrust towards the converging continent (B). Upon collision, the buoyancy of the converging continent prevents its consumption by the mantle, and large nappes and thrusts are produced as it resists subduction (C and D). Remnants of the intervening ocean basin (ophiolites) are emplaced along a suture zone together with melange and deep-sea sediments. The *suture* marks the boundary where the two plates are welded together, and is interpreted as evidence for a closed ocean or back-arc basin. Such suture zones characterize the northern flank of the Himalayas (the Indus suture) and the central Urals. A continent–continent collision results in the termination of a convergent plate boundary, and worldwide spreading rates must adjust to compensate for the loss of a subduction zone or a new subduction zone must develop. There is some evidence that a new convergent plate boundary, defined by a shallow seismicity zone extending southeast from India, may be in the early stages of development to compensate for the India–Tibet collision.

A possible sequence of events for a continent–arc collision is illustrated in fig. 6.21b. When a plate carrying a continent reaches a subduction zone (B and C), the subduction zone becomes choked with thrusts and nappes directed toward the

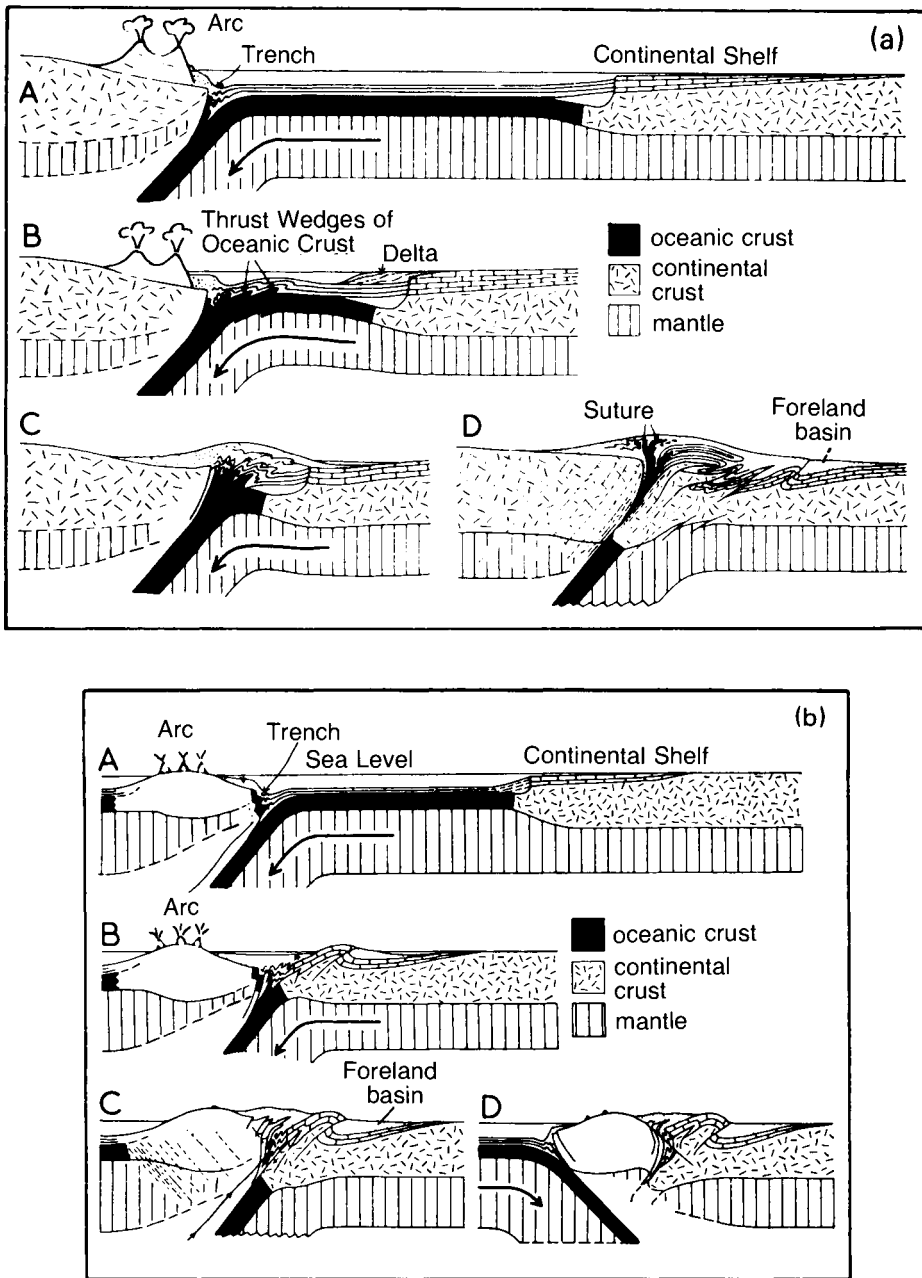


FIG 6.21 Schematic sequence of events in a continent-continent (a) and continent-arc collision (b). After Dewey and Bird (1970a).

continent. Spreading rates may adjust to compensate for the loss of the subduction zone, or the subduction zone may move to the ocean side of the arc, dipping in the opposite direction (D). The latter change, known as a *subduction-zone flip*, results in the accretion of an arc to a continent.

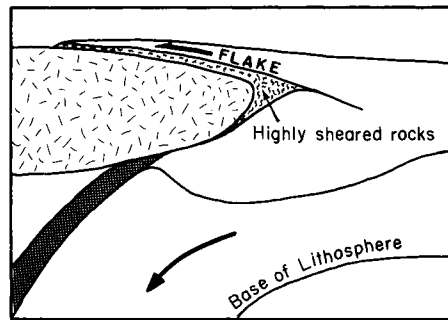


FIG 6.22 Schematic cross-section showing the emplacement of a flake or allochthon during a continental collision.

During plate collisions involving continental crust, major thrusts and nappes are directed toward the converging plate. In some instances, sheet-like slabs, commonly referred to as *flakes* or *allochthons*, may be sheared from the top of the converging plate and thrust over the overriding plate (fig. 6.22). In the Eastern Alps, for instance, Paleozoic metamorphic rocks have been thrust more than 100 km north over the Bohemian Massif and a zone of highly sheared Mesozoic metasediments lies within the thrust zone. The overriding flake is less than 12 km thick and appears to represent part of the Carnic plate (see fig. 8.17b) that was detached during the Mid-Tertiary. Continued convergence at collisional boundaries leads to thickening of the crust by ductile creep or thrusting at shallow levels. Seismicity along continent–continent collision boundaries suggests partial subduction of continental crust. Thickening of crust above collisional zones may result in partial melting producing felsic magmas that are erupted as ash-flow tuffs or intruded as plutons. Fractional crystallization of mantle-derived basalts may produce anorthosites in the lower crust, and losses of granite and volatiles from the thickened lower crust may leave behind granulite-facies mineral assemblages. Isostatic recovery is marked by the development of continental rifting with fluvial sedimentation and bimodal volcanism. Calc-alkaline volcanism may also continue for a short time in the overriding plate.

To some extent each collisional belt has its own character. In some instances, plates lock together with relatively little deformation, such as along the Caledonian suture in Scotland or the Kohistan suture in Pakistan. In other orogens, such as the Alps and Himalayas, flakes are thrust for considerable distances and stacked one upon another. Several hundred kilometers of shortening may occur during such a collision. In a few collisional belts, such as the late Proterozoic Damara belt in Namibia, a considerable amount of deformation and thickening may occur in the overthrust plate. In the Caledonides and parts of the Himalayan belt, ophiolite obduction precedes continent–continent collision and does not seem to be an integral part of the collision process. In some cases the collisional zone is oblique with movement occurring along one or more strike-slip faults. In other examples one continent will indent another, thrusting thickened crust over the indented block. Thickened crust tends to spread by gravitational forces, and spreading directions need not parallel the regional plate movement. Which block has the greatest effects of deformation and metamorphism depends on such factors as the age of the crust,

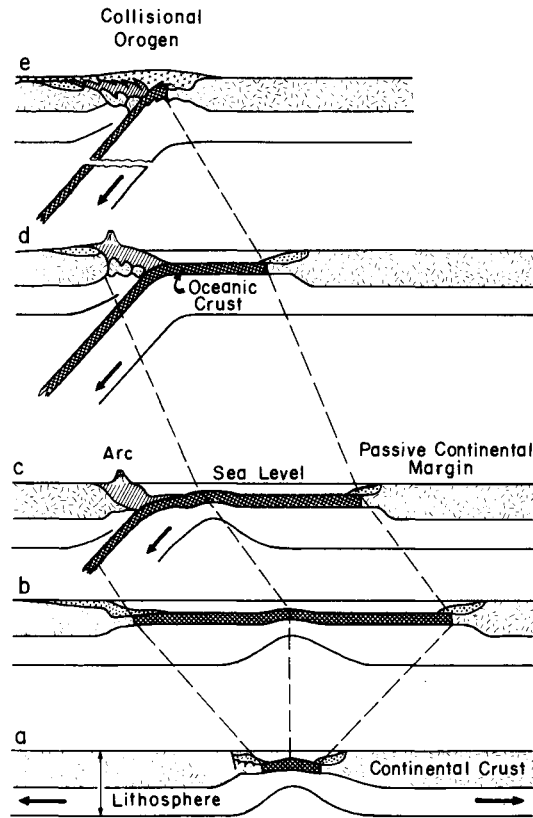


FIG 6.23 Idealized sequence of events in a Wilson cycle.

its thermal regime, anisotropy of the crust, and the nature of the subcrustal lithosphere. Old lithosphere generally has greater strength than younger lithosphere.

The Wilson Cycle

The opening and closing of an oceanic basin is known as *Wilson cycle* (Burke et al., 1977). As discussed in Chapter 5, Wilson cycles have distinct apparent polar wandering paths. Such a cycle begins with the rupture of a continent along a cratonic rift system, followed by opening of an ocean basin (fig. 6.23). Thus, the oldest petrotextonic assemblages are continental rift assemblages. As the basin continues to open, cratonic sediments are deposited along both passive continental margins, and abyssal sediments accumulate on the seafloor adjacent to these margins. At some point, subduction begins on one (fig. 6.23c) or both margins and the ocean basin begins to close. The net result of closure is a continent–continent collision. During the collision, arc rocks and ophiolites are thrust over passive-margin assemblages (e). The geologic record indicates that the Wilson cycle has occurred many times during the Phanerozoic. Because lithosphere is weakened along collisional zones, rifting may open new ocean basins approximately along older sutures, as evidenced by the opening of the Atlantic basin approximately along the Iapetus suture (see Chapt. 8).

Activation Orogeny

It is not yet agreed whether major orogenies can occur without collision. Earthquake first-motion studies in arc systems suggest that extensional or strike-slip forces dominate in most arcs. However, in arcs with thick accretionary prisms it is generally thought that compressive forces are important, at least in the forearc region. These same arcs usually have foreland fold-thrust belts, indicative of significant compression in back-arc regions.

The best candidate for a modern arc system that may be undergoing compressive, activation orogeny is the Andes in South America (fig. 6.16a). Although some accretionary terranes appear to have collided with the Andean Cordillera in the Mesozoic, three lines of evidence from the Andes favor activation orogeny for most of the deformation (Dalziel, 1986).

1. Most of the accretionary terranes in the Andes appear to have been sutured to South America before the onset of orogeny in the Middle Triassic.
2. The most intense deformation, plutonism, and uplift in the Andes occurred in response to a Mid-Cretaceous collision of a volcanic arc rifted from the continent in the Late Jurassic. Development of foreland basins followed this collision.
3. The great extent of the Mid-Cretaceous compressive event along the Pacific margin of South America (≥ 7500 km) seems to rule out collision of local accretionary terranes as a cause of orogeny. An increase in compressive stress during the Mid-Cretaceous may have been caused by a change in plate motions and an increase in seafloor spreading rates (100–120 Ma ago), both of which contributed to increasing the horizontal component of compression at a steep angle to the coast of South America.

Tectonic Elements of a Collisional Orogen

Continental collision involves progressive compression of buoyant terranes within subduction zones. These terranes may vary in scale from seamounts or island arcs to large continents. The scale of colliding terranes dictates the style, duration, intensity, and sequence of strain systems (Dewey et al., 1986). If colliding continental margins are irregular, the strain sequences are variable along great strike lengths. Prior to terminal collision, one or both continental margins may have had a long complex history of accretionary terrane assembly. Continental collisional boundaries are wide, complicated structural zones (fig. 6.24), where plate displacements are converted into complex and variable strains. Just how this conversion is accomplished remains a fundamental problem of plate tectonics. Horizontal strain alone cannot explain thickened crust in the Tibetan Plateau, where vertical stretching may play an important role.

Collisional orogens can be considered in terms of five tectonic components (fig. 6.25); thrust belts, foreland flexures, plateaus, widespread foreland/hinterland deformational zones, and zones of orogenic collapse. *Foreland* and *hinterland* refer to regions beyond major overthrust belts in the direction and away from the direction, respectively, of principal orogenic vergence. *Thrust belts* develop where thinned continental crust is progressively restacked and thickened toward the foreland. If

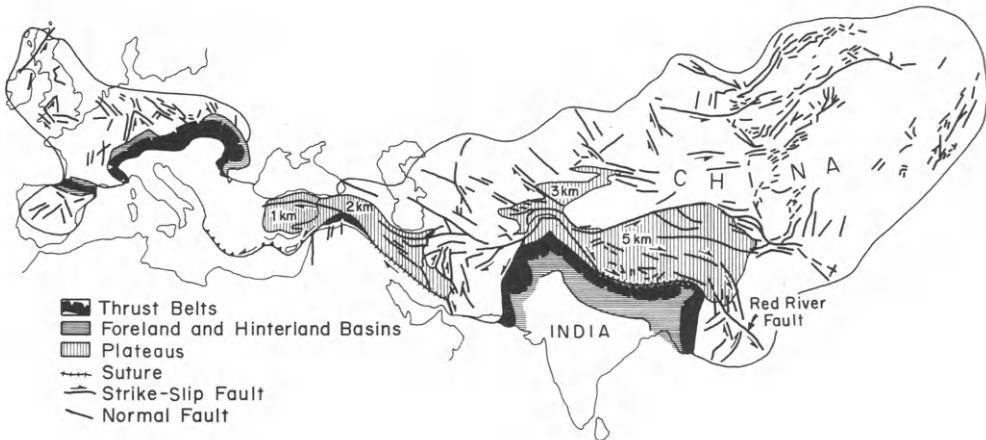


FIG 6.24 Simplified tectonic map of the Alpine-Himalayan orogen. After Dewey et al. (1986). Collisional plateau elevations given in kilometers. 1 km, Anatolian Plateau; 5 km, Tibetan Plateau.

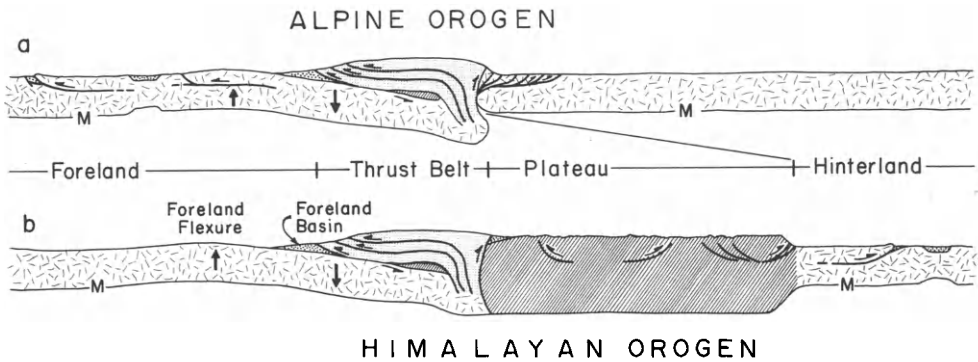


FIG 6.25 Schematic cross-sections of Alpine (a) and Himalayan (b) collisional orogens. After Dewey et al. (1986). M, Moho.

detachment occurs along a foreland thrust, rocks in the allochthon can shorten significantly, independent of shortening in the basement. The innermost nappes and the suture zone are usually steepened and overturned during advanced stages of collision. The upper crust (5–15 km thick) in collisional orogens is a high-stretch layer that may be thrust hundreds of kilometers as relatively thin, stacked flakes that merge along a decollement surface. For instance, in the southern Appalachians a decollement has moved westward over the foreland for at least 300 km (see Chapt. 8). Structural and paleomagnetic evidence indicate that most decollements have undergone rigid-body rotation. *Foreland flexures* are upwarps in the foreland lithosphere caused by progressive down bending of the lithosphere by advancing thrust sheets. The wavelength and amplitude of foreland flexures depends on rigidity and temperature distribution of the lithosphere. *Collisional plateaus*, such as the Tibetan and Anatolian plateaus (fig. 6.24), form in the near-hinterland area

adjacent to the suture and may rise to 1–5 km above sea level. These plateaus form in response to shortening and thickening of the lithosphere in the overriding plate or/and by buoyant subduction of continental crust. In the first case, plateaus rise in response to compressive forces that thicken the lithosphere and crust, whereas in the second case plateaus rise in response to isostatic adjustment caused by underthrusting of continental crust. The widespread *foreland/hinterland deformational zones* are evidence that stresses generated by collision can affect continental areas hundreds to thousands of kilometers from the thrust belt (fig. 6.24). The relative amounts of deformation taken up by the foreland and hinterland vary considerably, and depend on thermal age and anisotropy of the lithosphere in converging and overriding plates. The preferential hinterland deformation in Asia associated with the India–Tibet collision probably reflects warm and thin Tibetan lithosphere caused by pre-collisional subduction and accretion (Chapt. 8). Extensional *collapse zones*, such as the Aegean basin, are generally of local occurrence in foreland areas and may develop in response to rollback of a partially subducted slab.

Sutures

Sutures are melange zones ranging from a few hundred meters to over 3 km wide in collisional orogens (Coward et al., 1986). Rocks on the hinterland plate adjacent to sutures are chiefly arc-related volcanics and sediments from the former arc system on the overriding plate, whereas those on the foreland plate are passive-margin sediments. Fragments of rocks from both plates as well as ophiolite fragments are commonly mixed in suture melanges. These fragments are typically tens to hundreds of meters in size and are randomly mixed in a sheared matrix. Sutures can be considered as ductile shear zones produced by thrusting along the converging plate boundaries. Matrices generally exhibit strong planar fabrics and although composed chiefly of chlorite and micas, they vary considerably in mineral content.

Peripheral Foreland Basins

Peripheral foreland basins are like retroarc foreland basins in terms of sediment provenance and tectonic evolution. Major peripheral foreland basins developed in the Tertiary in response to the Alpine–Himalayan collisions (Allen and Homewood, 1986). Many peripheral basins exhibit similar stages of development that have been explained by thermal–mechanical models (Stockwell et al., 1986). During the first stage a passive-margin sedimentary assemblage is deposited on a stretched and rifted continental margin. Collision begins as an accretionary terrane is thrust against and over continental crust on the descending slab. Continental convergence results in thickening of this terrane and partial melting of the root zones to form syntectonic leucogranite magmas. Topographic relief develops at this stage as the thrust belt rises above sea level and sediments derived from erosion of the highland begin to fill a foreland basin. The first sediments come from distal low-relief terranes and are largely fine-grained, giving rise to deepwater marine shales and siltstones exposed in the lowest stratigraphic levels of foreland basin successions. Continued convergence causes thin-skinned thrust sheets to propagate into foreland basins and relief increases rapidly. Erosion rates increase, as do grain size and feldspar content of

derivative sediments in response to increased rates of tectonic uplift in the thrust belt. During this stage of development, thick alluvial fans may be deposited. Estimated rates of landward migration of foreland basins in response to converging thrusts generally range from 10 to 30 mm/a.

Accretionary Terranes

Accretionary terranes are fault-bounded continental segments of regional extent that have geologic histories different from neighboring terranes (Schermer et al., 1984). The general absence of evidence for ophiolite obduction or suture melange suggests that the boundaries of most terranes are faults. Collision and eventual accretion of these terranes to continents appears to have occurred chiefly along transform faults or subduction zones with a large strike-slip component. Many accretionary terranes contain faunal populations indicating they have been displaced great distances from their sources, and paleomagnetic studies support this conclusion. Results suggest that as much as 30% of the North American continent formed by terrane accretion in the last 300 Ma, and that terrane accretion has been an important process in the growth of continents.

Some accretionary terranes represent either arc systems or pieces of continental crust rifted away from a craton. Others may represent active and aseismic volcanic ridges and oceanic plateaus. Numerous potential accretionary terranes exist in the oceans today and are particularly abundant in the Pacific basin (fig. 6.26). Continental crust may be fragmented and dispersed by rifting or strike-slip faulting. In western North America, dispersion is occurring along transform faults such as the San Andreas and Fairweather faults, and in New Zealand movement along the Alpine transform fault is fragmenting the Campbell Plateau from the Lord Howe Rise. Dispersive processes may fragment individual terranes after their collision with a continent, such as occurred in Wrangellia in northwestern North America (Chapt. 8). Terrane displacements can be estimated from several approaches including estimates of fault displacement from geologic mapping, matching of ancient faunal provinces, and paleomagnetic studies. In some instances it has been possible, using paleomagnetic data, to reconstruct terrane trajectories from their place of origin to their final accretion.

Mineral and Energy Deposits

Introduction

Plate tectonics provides a basis for understanding the distribution and origin of mineral and energy resources in space and time (Rona, 1977). The occurrence of energy and mineral deposits can be related to plate tectonics in three ways: (1) geological processes driven by energy liberated at plate boundaries control the formation of energy and mineral deposits; (2) deposits form in specific tectonic settings, which are in turn controlled by plate tectonics; (3) reconstruction of fragmented continents can be used in exploration for new mineral and energy deposits. In attempting to relate mineral and energy deposits to plate tectonics it is important to know the relationship between the deposits and the host rocks. If deposits are syngenetic with host rocks, then they formed in the same tectonic

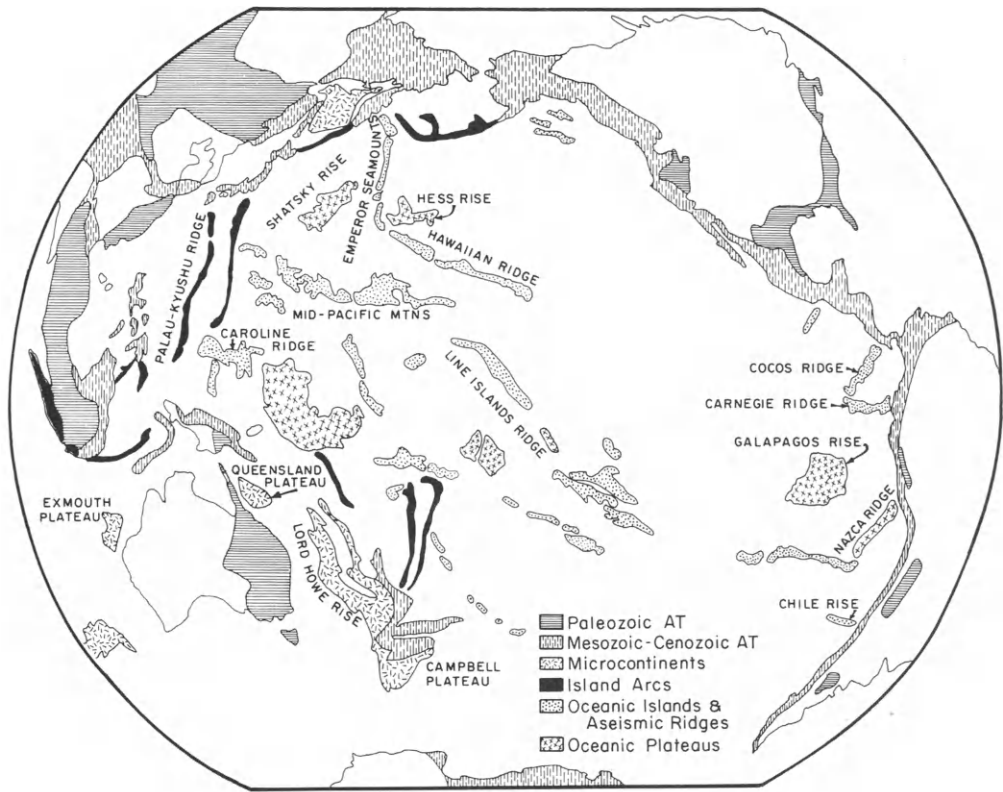


FIG 6.26 Generalized map of the circum-Pacific region showing distribution of accretionary and potential accretionary terranes. Modified after Schermer et al. (1984).
AT, accretionary terranes.

setting. If, however, mineral and energy deposits are secondary in origin and entered the host rocks later (such as by hydrothermal fluids or migration of oil), they may have formed in a tectonic setting different from the host rocks. Major *in situ* mineral and energy deposits in various tectonic settings are summarized in table 6.3.

Mineral Deposits

Ocean Ridges

Employing deep-sea photography and using small submersibles, numerous hydrothermal fields have been identified on ocean ridges (Corliss et al., 1979; Rona et al., 1986). The submersibles allow direct observations and measurements on the seafloor. On the Galapagos and Juan De Fuca ridges, hydrothermal veins ranging in

TABLE 6.3 *Summary of Major Mineral and Energy Deposits in Relation to Tectonic Setting*

Tectonic Setting	Mineral Deposits	Energy Deposits
1. Ocean ridge, back-arc basin (ophiolites)	Cyprus-type Cu-Fe massive sulfides Podiform chromite Metalliferous sediments	Geothermal, Hydrocarbons (back-arc basins)
2. Subduction Zone (a) Arc	Hydrothermal veins: base metals (Au, Cu, Mo, Pb) precious metals (Au, Ag) other metals (Sb, Hg, Sn, W) Porphyries (Cu, Mo, Sn) Massive sulfides (Cu, Pb, Zn) Red bed U, V, Cu	Geothermal
(b) Retroarc foreland basin		Hydrocarbons, coal
(c) Forearc basin		Hydrocarbons
3. Collisional Zones (a) Highlands	Ophiolite and arc deposits Sn-W granites Fe-Ti deposits (anorthosites) Gemstones	
(b) Peripheral foreland basin	Red bed U, V, Cu Stratiform Pb-Zu-Cu Stratiform Pb-Zn-Ag	Hydrocarbons, oil shale, Coal
4. Continental Rifts (Aborted rifts, aulacogens, hotspots)	REE, Nb, U, Th, P, halogens Ba, Sr in Alkaline complexes Sn granites Stratiform Cu Evaporites	Geothermal
5. Cratons, passive continental margins	Diamonds (kimberlites) Bauxite Ni-laterite Evaporites	Hydrocarbons, coal

size from 500 to 1600 m² are spread along fissure systems for 500–2500 m. The TAG hydrothermal field on the Mid-Atlantic ridge (26°N) occurs along a fault zone on the east wall of the medial rift valley. Chemical equilibria data and direct measurement indicate that water given off in these fields has temperatures up to 300°C. Animal communities living near the vents appear to be totally dependent upon energy derived from seawater-rock reactions and sulfur-oxidizing bacteria. Seafloor observations also indicate that hydrothermal vents are locally developed in areas of the youngest seafloor, and occur along both fast- and slow-spreading ridge segments. Each hydrothermal field has multiple discharge sites, with sulfide chimneys rising up to 10 m above the seafloor (fig. 6.27). Minerals recovered from chimneys are chiefly Fe, Cu, and Au sulfides with minor amounts of anhydrite and amorphous silica. Hydrothermal waters are acid, rich in H₂S, and are major sources of Mn, Li, Ca, Ba, Si, Fe, Cu, and Au. These metals are either deposited in the chimneys or with sediments near the vents.

The hydrothermal waters erupted along ocean ridges appear to represent seawater that has been circulated through hot oceanic crust with thermal gradients of ±150°C/km. The seawater probably circulates by convection through the upper part of the permeable crust, and is heated at depth and discharges along active fissures in

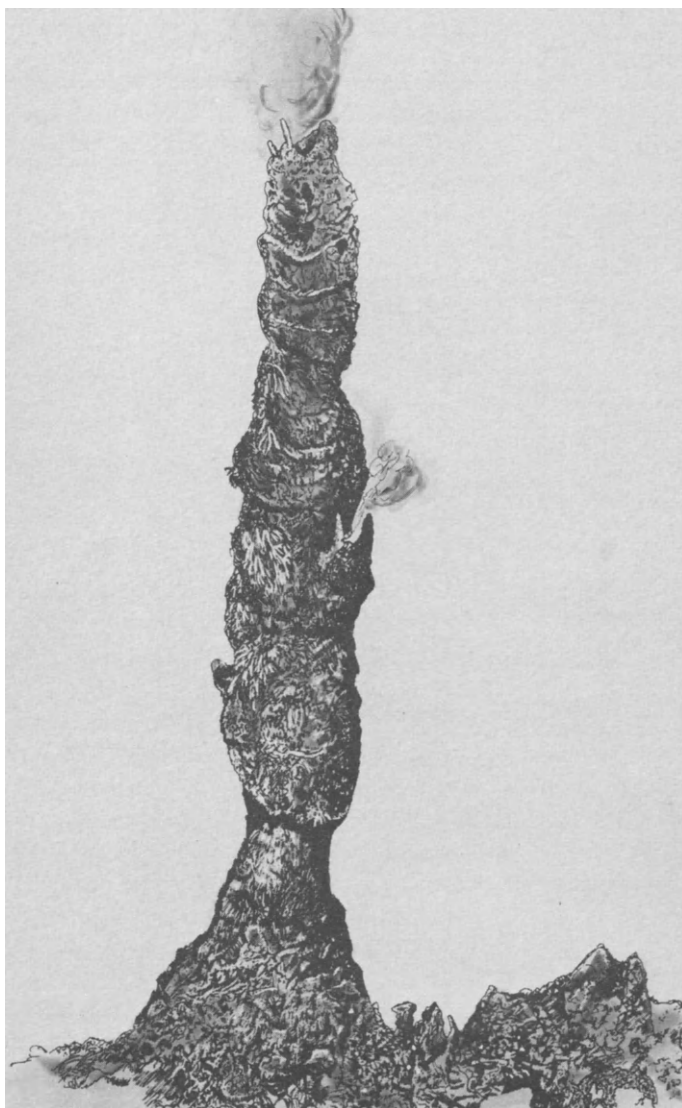


FIG 6.27 Interpretative line drawing of a 9-m high chimney from an active hydrothermal field on the Juan De Fuca ridge. Courtesy of U.S. Geological Survey, W. R. Normark and W. C. Shanks.

the medial rift. The discharging water leaches metals from the oceanic crust and alters the crust by addition of H_2O and such elements as Mg. Upon erupting on the seafloor, the hydrothermal waters are rapidly cooled and increase in pH, resulting in deposition of sulfides and sulfates.

In considering ancient oceanic crust, two major mineral deposits are formed in ophiolites: Cyprus-type Cu-Fe massive sulfides and podiform chromite. Cyprus-type ores occur as stratiform deposits in pillowed basalt layers. These ores appear to be exhalative deposits formed by hydrothermal vents along ocean ridges as described

above. Podiform chromite deposits are formed in ultramafic portions of ophiolites, and relict textures suggest they are the products of fractional crystallization.

Arc Systems

Metalliferous mineral deposits are important at both continental and arc convergent plate boundaries. A major part of known base metals (Zn, Cu, Mo, Pb), precious metals (Ag, Au) and other metals (Sn, W, Sb, Hg) are found in hydrothermal veins and lodes formed in arc systems. Following erosion, placer deposits of these metals are important in some geographic areas. Veins and lodes are commonly associated with volcanics or granitic plutons, where they probably represent late-stage fluids derived from differentiated arc magmas. Cu, Mo, and Sn porphyry deposits are formed also in arc systems. These large-volume, low-grade disseminated deposits occur in altered porphyritic granites and are important sources of Cu and Mo in the southwestern United States and in the Andes. Kuroko-type massive sulfides (Cu, Pb, Zn) are important in intra-arc and back-arc basin successions where they appear to have formed as exhalatives on the seafloor. Redbed U, V, and Cu deposits occur in some retroarc foreland basins such as in the Mesozoic redbeds of the Colorado Plateau.

Zonation of metallic mineral deposits has been reported in Late Tertiary rocks from the Andes (Sillitoe, 1976). Going east from the coast toward greater depths to subduction zone, the major zones encountered are: contact metasomatic Fe deposits; Cu, Au, and Ag veins; Cu–Mo deposits; Pb–Zn–Ag vein and contact metasomatic deposits; and Sn and Mo vein and porphyry deposits. Belts are believed to be caused by progressive liberation of metals from the descending slab, with Sn coming from an extreme depth of about 300 km. Geochemical and isotopic data support the general concept that metal deposits associated with subduction are derived from some combination of the descending slab and the overlying mantle wedge. Metals move upward in magmas or in fluids, and are concentrated in late hydrothermal or magmatic phases.

Collisional Orogens

Metalliferous deposits are abundant at collisional boundaries where a variety of tectonic settings exist, depending on distance from the thrust belt and stage of evolution. In addition, older mineral deposits associated with ophiolites, arc, craton, and continental rift assemblages occur in collisional zones. Sn and W deposits are associated with collisional leucogranites in the Himalayas and in the Hercynian orogen and, at deeper crustal levels, Fe and Ti deposits occur associated with anorthositic rocks. Many gemstones (ruby, sapphire) are also found in high-grade metamorphic rocks or in syntectonic nepheline syenites from collisional orogens. In peripheral foreland basins, stratiform Pb–Zn–Cu sulfides and redbed U–V–Cu deposits may be of economic importance.

Continental Rifts

One of the major occurrences of Pb–Zn–Cu stratiform deposits is in continental rifts (Sawkins, 1982). These deposits, which are not associated with igneous rocks, occur in marine carbonates and are probably deposited from brines which migrate to

TABLE 6.4 *Oil Reserves in Devonian and Younger Reservoirs (in percentages)*

	Cratons		Arcs and Collisional Orogens
	Intracratonic	Passive	Foreland Basins
Cenozoic	5	8	18*
Mesozoic	43	5	6
Paleozoic	13	1	1
Total	61	14	25

Modified after Bois et al. (1982).

*Fourteen percent in the Persian Gulf and 4% elsewhere.

the edge of rift basins. REE, Nb, U, Th, Ba, P, Sr, and halogens are concentrated in carbonatites and other alkaline igneous rocks that occur in some continental rifts. Granites intruded during the late stages of rifting are often associated with Sr and fluorite. Stratiform Cu deposits occur in rift-related shales and sandstones as exemplified by deposits in the Zambian Copper Belt. The Cu appears to be derived from associated basalts. Evaporites are important non-metallic deposits found in some rifts.

Cratons and Passive Continental Margins

Few if any metallic deposits are known to form on modern cratons or passive continental margins. Placer Au–U and banded iron formation, however, are examples of deposits formed in Precambrian cratonic basins. Among the non-metallic deposits formed in cratonic areas are diamonds from kimberlite pipes and associated placer deposits, bauxite, Ni-laterite deposits, and evaporites.

Energy Deposits

Several requirements must be met in any tectonic setting for the production and accumulation of hydrocarbons such as oil or natural gas. First of all, the preservation of organic matter requires restricted seawater circulation to inhibit oxidation and decomposition. High geothermal gradients are needed to convert organic matter into oil and gas and finally, tectonic conditions must be such as to create traps for the hydrocarbons to accumulate. Several tectonic settings can potentially meet these requirements.

Both oil and gas are formed in forearc and back-arc basins, which can trap and preserve organic matter, and geothermal heat facilitates conversion of organic matter into hydrocarbons. Later deformation, generally accompanying collisions, creates a variety of structural and stratigraphic traps in which the hydrocarbons accumulate. An important source of hydrocarbon formation is in foreland basins. The immense accumulations in the Persian Gulf area formed in a peripheral foreland basin associated with the Arabia–Iran collision in the Tertiary.

The majority of the oil and gas reserves in the world appear to have formed either in intra-cratonic basins or passive continental margins (table 6.4). The opening of a continental rift system into an ocean basin involves a stage when seawater moves into the rift valley and evaporation exceeds inflow and evaporites are deposited. This environment is also characterized by restricted water circulation in which organic

matter is preserved. As the rift continues to open, water circulation becomes unrestricted and accumulation of organic matter and evaporite deposition cease. High geothermal gradients beneath the opening rift, and increasing pressure due to burial of sediments, facilitates the conversion of organic matter into oil and gas. At a later stage of opening, salt beds in the evaporite succession, because of their gravitational instability, may rise as salt domes and trap oil and gas. Oil and gas may also be trapped in structural or stratigraphic traps as they move upward in response to increasing pressures and temperatures at depth. Supporting this model are data from wells in the Red Sea, which represents an early stage of ocean development. These wells encounter hydrocarbons associated with high geotherms and with rock salt up to 5 km thick. Also, around the Atlantic basin there is a close geographic and stratigraphic relationship between hydrocarbons and evaporite accumulation.

Hydrocarbon production in intracratonic basins may also be related to plate tectonic processes (Rona, 1977). Increases in seafloor spreading rates and ocean-ridge lengths may cause marine transgression, which results in deposition in intracratonic basins. Decreasing spreading rates causes regression, which results in basins with limited circulation, and hence organic matter and evaporites accumulate. Unconformities also develop during this stage. Burial and heating of organic matter in intracratonic basins facilitates hydrocarbon production, and salt domes and unconformities may provide major traps for accumulation.

Coal is formed in two major tectonic settings: cratonic basins and foreland basins. For coal to form, plant remains must be rapidly buried before they decay, and such rapid burial occurs in swamps with high plant productivity. Widespread transgression in the Cretaceous was particularly suitable for coal swamps in cratonic areas. Swamps in foreland basins are generally part of large lake basins, and the widespread Late Paleozoic coals of central Europe appear to have formed in such environments. Oil shales can accumulate under similar conditions, as exemplified by the Tertiary oil shales in the early Tertiary foreland basins of Eastern Utah.

Geothermal energy sources occur along ocean ridges with Iceland representing the only subaerial expression of a ridge. Other geothermal sources occur associated with hotspots (such as Yellowstone Park), arc systems (such as the Taupo area in New Zealand), and continental rifts (such as the Jemez site in New Mexico).

Summary Statements

1. Ophiolites are tectonically emplaced successions of mafic and ultramafic rocks that appear to represent fragments of oceanic or back-arc crust and upper mantle. In ascending stratigraphic order, ophiolites are composed of ultramafic tectonite, cumulate gabbros and ultramafic rocks, noncumulate igneous rocks, sheeted diabase dikes, and pillow basalts. Ophiolites are emplaced by obduction (thrusting) at collisional boundaries or as fault slices in accretionary prisms of arcs. Decoupling may occur near spreading centers as ocean ridges are subducted.
2. Seven environments are recognized in oceanic arcs. These are, with corresponding petrotectonic assemblages: (1) the trench (graywacke turbidites); (2) the accretionary prism (melange, fragments of ophiolite and sediment); (3) forearc basins (turbidites, hemipelagic sediments); (4) the volcanic arc (flows, fragmen-

- tal deposits); (5) active back-arc basins (arc or/and pelagic sediments); (6) the remnant arc (volcanics); and (7) inactive back-arc basins (arc or/and pelagic sediments). Retroarc foreland basins form behind continental margin arcs and are characterized by coarsening upwards, immature clastic sediments.
3. Continental rifts are characterized by immature, terrigenous sediments and generally bimodal igneous rocks. Craton assemblages include quartzites, shales, and shallow marine carbonates. Collisional assemblages are highly variable, involving some combination of immature terrigenous sediments (in peripheral foreland basins), graywackes in remnant ocean basins, and at deeper crustal levels, syntectonic leucogranites, granulites and anorthosites.
 4. Subduction begins along zones of lithosphere weakness such as extinct fracture zones or passive continental margins. Arcs grow by subduction-zone accretion and remain constant or decrease in size by subduction erosion and sediment subduction.
 5. Back-arc basins probably open in response to passive spreading in which either the descending or overriding plate retreats, resulting in an arc system being subjected to tensional forces.
 6. Subduction zones appear to range from high-stress to low-stress types. High-stress types have a large accretionary prism, relatively large earthquakes, and a shallow dip to the subduction zone, whereas low-stress types have little if any accretionary prism, few large earthquakes, and a very steep subduction zone.
 7. Active continental rifts, characterized by early and widespread uplift and abundant volcanics, develop in response to rising mantle plumes or convective upcurrents. Passive rifts, characterized a narrow zone of uplift and few volcanics, develop in response to stresses in moving lithospheric plates. Rifts may evolve from one type to another and rifting may cause fragmentation of a continent and formation of a new ocean basin.
 8. Extension during continental rifting is accommodated by successive listric faults that merge with a large subhorizontal detachment fault at mid-crustal levels.
 9. Cratonic basins subside at approximately the same exponential rate as ocean basins. Subsidence is probably related to thermal doming and eroding the top of the lithosphere, followed by contraction or thinning and stretching of the lithosphere.
 10. Collisional orogens vary greatly in character, ranging from those with large, overthrust flakes to those in which compression is taken up partly by shortening of the overriding plate. In some orogens, collision is oblique, producing a major strike-slip component. The opening and closing of ocean basins (a Wilson cycle) has occurred many times in the geologic past. Although activation orogeny may be locally important, most orogenies appear to have resulted from plate collisions.
 11. A collisional orogen is composed of, (1) a thrust belt where continental crust is restacked and propagates towards the foreland; (2) a foreland flexure caused by down-bending of the lithosphere by advancing thrust sheets; (3) plateaus produced by shortening and thickening of the lithosphere in the overriding plate or/and by buoyant subduction; (4) widespread foreland/hinterland deformation zones; and in some cases (5) extensional collapse zones.
 12. Sutures are melange zones up to 3 km thick separating collisional plates. They

contain a wide variety of rock fragments (up to hundreds of meters in size) in a sheared matrix.

13. Accretionary terranes are fault-bounded continental crustal blocks that have geologic histories different from neighboring terranes. Many of these terranes represent arcs or microcontinents. Matching of faunal provinces and paleomagnetic studies indicate that some provinces have traveled many thousands of kilometers from their sources. Terrane accretion is an important process in the growth of continents.
14. Mineral and energy deposits are closely related to tectonic setting. Sulfides are being deposited along modern ocean ridges by hydrothermal waters. Arc systems are major sites of formation of Zn, Cu, Pb, Mo, Ag, Au, Sn, W, Sb, and Hg deposits, and continental rifts are important sites of formation of REE, Th, U, P, Sr, Ba, and Nb. Sn, W, Fe, Ti, Pb, Cu, Zn, U, and V occur in collisional orogens. Most hydrocarbons and coal are formed in cratonic or foreland basins.

Suggestions for Further Reading

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CHAPTER 7

Magma Associations and Mantle Sources

Introduction

The distribution of igneous rocks on the Earth varies with tectonic setting. Basalts dominate in oceanic areas, andesites or basalts characterize oceanic arcs with felsic igneous rocks becoming more important in continental-margin arcs. Continental rifts often contain basalts and rhyolites with few intermediate compositions. Such distributions suggest a cause-and-effect relationship between plate tectonics and magma generation and source. A great deal of geochemical and isotopic data indicate that magma production mechanisms and sources vary with tectonic setting, and hence by studying igneous rocks we can learn more about magma sources in the mantle and crust.

Three principal *magma series* are recognized with each series composed of a group of closely related igneous rocks. These are the *tholeiite*, *calc-alkaline* and *alkaline series*. If felsic and mafic end members predominate, the association is referred to as *bimodal*. Some major igneous-rock types are given in table 7.1 with their SiO_2 and K_2O contents and principal minerals. Typical chemical compositions of major mafic and intermediate volcanic rocks are given in table 7.2. Various rock types in each magma series may be produced by varying degrees of melting of the source and/or fractional crystallization, and bimodal igneous suites may have end members that have different sources. Although magma series generally exhibit a continuous range of compositions, some rock types may greatly dominate, such as tholeiites in the tholeiite series and andesites in the calc-alkaline series. Komatiites, which are abundant only in the Archean, may comprise a third magma series. Although not limited to a given tectonic setting, each magma series is characteristic of one or more tectonic settings (table 7.3). Subduction zones are characterized by both the tholeiite and calc-alkaline series, with the tholeiite series showing iron enrichment when plotted on an AFM diagram (fig. 7.1). Basalts or basaltic andesites of the tholeiite series generally dominate in immature oceanic arcs, and andesites of the calc-alkaline series often dominate in immature arcs. Continental-margin arcs and collisional orogens are commonly characterized by felsic calc-alkaline igneous rocks. The magmas erupted at ocean ridges or continental rifts are chiefly basalts of the tholeiite series and show strong iron enrichment (fig. 7.1). The alkaline series is volumetrically very minor and occurs almost exclusively on oceanic islands and continental rifts. Igneous rocks in cratons are rare, and when found they are generally members of the alkaline series or kimberlites.

TABLE 7.1 *Some Chemical and Mineralogical Features of Igneous Rocks*

	SiO ₂ %	K ₂ O%	Principal Minerals ¹	Magma Series ²
Tholeiite (gabbro)	48-50	0.2-1	plg, pxs, ± ol	TH, CA
Calc-alkaline basalt (gabbro)	50-53	0.2-0.8	plg, pxs, ± ol	CA
Picrite	43-46	<0.4	ol, px, plg	TH
Andesite (diorite)	56-62	0.7-2.5	plg, pxs, mgt ± amp	CA, TH
Boninite	54-60	0.2-0.6	ol, pxs	TH
Dacite (tonalite)	62-68	1.2-2.5	plg, pxs, qtz, biot, mgt ± amp	CA, TH
Rhyodacite (granodiorite)	65-71	2.5-4	plg, qtz, amp, biot, mgt	CA
Rhyolite (granite)	71-75	4-5.5	K-feld, qtz, plg, biot, mgt	CA
Shoshonite	54-56	2.8-4	plg, K-feld, pxs	AL
Alkali basalt (alkali gabbro)	44-46	0.7-1	plg, ol, pxs, ± neph, lc	AL
Nephelinite (ijolite)	38-40	1-2	neph, pxs, mgt	AL
Latite (monzonite)	52-54	3.5-5	plg, K-feld, ± biot, ± amp	AL
Trachyte (syenite)	58-61	5-7.5	K-feld, plg, ± biot, ± amp	AL
Phonolite (nepheline syenite)	55-57	4-5.5	K-feld, neph, plg, ± amp, ± pxs	AL
Leucite (fergusite)	40-47	3-8	lc, pxs	AL
Kimberlite	30-40	0.5-2	ol, pxs, phl, gar, mgt, serp	
Komatiite	40-45	<0.2	ol, pxs	

Note: Extrusive rock names given with intrusive equivalents in parentheses.

¹In order of decreasing abundance. plg = plagioclase, pxs = pyroxenes, amp = amphibole, neph = nepheline, biot = biotite, K-feld = K-feldspar, qtz = quartz, ol = olivine, mgt = magnetite-ilmenite, gar = garnet, phl = phlogopite; lc, leucite; serp, serpentine; cpx = clinopyroxene; cord = cordierite; hb = hornblende; ilm = ilmenite; mon = monazite; musc = muscovite; sph = sphene.

²TH, tholeiite; CA, calc-alkaline; AL, alkaline.

The distribution of igneous rocks in part reflects the stress distribution in the lithosphere (table 7.3). It seems that, for large volumes of magma to reach the surface, extensional or strike slip faults are necessary. Where large volumes of volcanics occur, such as at ocean ridges and in arc systems, major extensional stress regimes are found. Very little magma reaches the surface at collisional plate boundaries where compressive stresses dominate. At these localities, magmas generally crystallize at depth, producing plutons. Stress regime also affects magma composition. Where extensional faults are important, basaltic magmas erupt in great quantities. In compressional regimes, magmas may pond and undergo extensive fractional crystallization, and thus felsic compositions may dominate.

Trace-element and isotope distributions in igneous rocks are particularly revealing in understanding the origin of magmas and in accounting for their distribution in terms of plate tectonics. Trace-element concentrations vary significantly between rocks with approximately the same overall major-element composition. As an example, elements like Rb, Ba, and Sr vary by two orders of magnitude between ocean-ridge tholeiites and alkali basalts (table 7.2). Elements like these, that are strongly concentrated in the liquid phase during partial melting or fractional

TABLE 7.2 Average Composition of Basalts and Andesites

	Ocean Ridge Basalt (NMORB)	Island Arc Basalt (IAB)	Continental Rift Tholeiite (CRB)	Island Tholeiite (OIB)	High-Al Tholeiite (CAB)	Oceanic Alkali Basalt	Continental Rift Alkali Basalt	Arc Andesite	Low-K Andesite	High-K Andesite	Shoshonite
SiO ₂	49.8	51.1	50.3	49.4	5.17	47.4	47.8	57.3	59.5	60.8	52.9
TiO ₂	1.5	0.83	2.2	2.5	1.0	2.9	2.2	0.58	0.70	0.77	0.85
Al ₂ O ₃	16.0	16.1	14.3	13.9	16.9	18.0	15.3	17.4	17.2	16.8	17.2
Fe ₂ O ₃ *	10.0	11.8	13.5	12.4	11.6	10.6	12.4	8.1	6.8	5.7	8.4
MgO	7.5	5.1	5.9	8.4	6.5	4.8	7.0	3.5	3.4	2.2	3.6
CaO	11.2	10.8	9.7	10.3	11.0	8.7	9.0	8.7	7.0	5.6	6.4
Na ₂ O	2.75	1.96	2.50	2.13	3.10	3.99	2.85	2.63	3.68	4.10	3.50
K ₂ O	0.14	0.40	0.66	0.38	0.40	1.66	1.31	0.70	1.6	3.25	3.69
Cr	300	50	160	250	40	67	400	44	56	3	30
Ni	100	25	85	150	25	50	100	15	18	3	20
Rb	1	5	31	5	10	33	200	10	30	90	100
Sr	135	225	350	350	330	800	1500	215	385	620	850
Ba	11	50	170	100	115	500	700	100	270	400	850
Zr	85	60	200	125	100	330	800	90	110	170	150
La	3.9	3.3	33	7.2	10	17	54	3.0	12	13	15
Yb	4.0	1.9	4.4	1.7	2.7	1.5	1.7	2.7	1.9	3.2	1.7
Th	0.18	0.5	1.5	0.67	1.1	4.5	4.0	1.3	2.2	5.5	2.8
Nb	3	1	25	23	5	30	35	1	3	6	4

*Total Fe as Fe₂O₃. Major elements in percent, trace elements in ppm.

TABLE 7.3 Plate Tectonic Classification of Magmas

	Plate Margin		Intraplate			
	Subduction zone	Ocean ridge	Oceanic	Continental		
			Oceanic island and seamounts	Rift systems (including plateau basalts)	Craton	Collisional zone
	Tholeiite	Tholeiite	Tholeiite	Bimodal	Alkali	Calc-alkaline
	Calc-alkaline		Alkali (Bimodal)	Tholeiite Alkali	Bimodal	Bimodal
	Extension (Compressive, strike-slip)	Extension	Passive	Extension	Passive	Compressive (Extension, strike-slip)

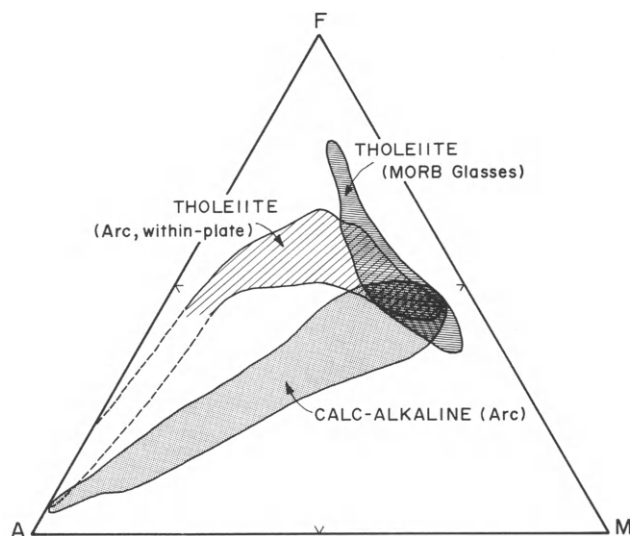


FIG 7.1 AFM diagram showing the tholeiite and calc-alkaline series. A, $K_2O + Na_2O$; F, FeO; M, MgO.

crystallization, are known as *incompatible elements* and are important in the study of magma production, source composition, and evolutionary histories of magmas. Unfortunately, many incompatible elements are also *mobile elements* in that they move about during secondary processes such as alteration and metamorphism. Among the most mobile elements are K, Rb, Cs, U, Sr, Ba, and Pb, and because of their mobility one must use caution in evaluating magma origin with these elements. Relatively immobile, incompatible elements include the *rare earth elements (REE)* and *high field strength elements (HFSE)* such as Ti, Ta, Nb, Zr, Hf, and Y.

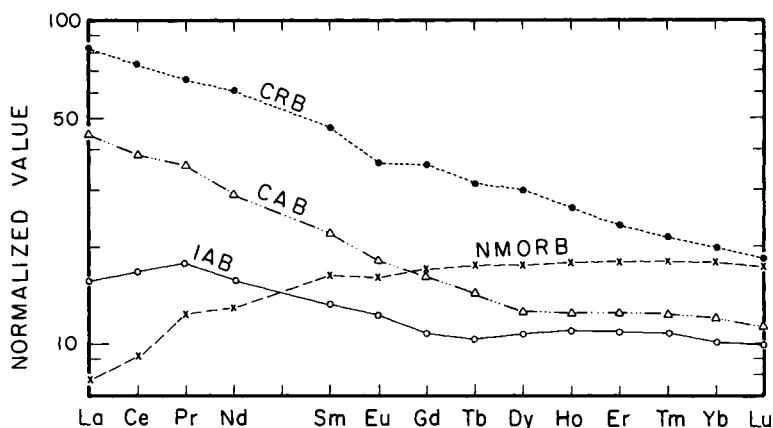


FIG 7.2 Chondrite-normalized REE distributions in typical basalts. NMORB normal ocean-ridge basalt; IAB, island-arc basalt; CAB, calc-alkaline basalt; CRB, continental-rift basalt. Data from BVTP (1981).

Incompatible element distributions are commonly portrayed by normalizing element concentrations to chondrites or ocean-ridge basalts and arranging elements in order of increasing incompatibility in a presumed mantle source (generally garnet lherzolite) from right to left on the horizontal axis. Such graphs are often referred to as *spidergrams* because of the complex patterns. Lines tend to radiate from the least incompatible end of the diagram. It is possible on spidergrams to show (1) differences in absolute concentrations, (2) differences in relative distribution patterns, and (3) anomalies in particular elements. A chondrite-normalized graph is commonly used to portray REE distributions. Basalts from different tectonic settings show a wide range of REE distributions (fig. 7.2). Normal ocean-ridge basalts (NMORB) show depletion in light REE, arc-related basalts range from flat to light REE enriched, and within-plate basalts show both relative light REE and total REE enrichment. Depletions or enrichments in Eu, known as *Eu anomalies*, are minor or absent in most basalts but may be significant in felsic igneous rocks. Another important spidergram is the NMORB-normalized graph. In this graph, various igneous rocks can be compared readily to NMORB. Arc-related basalts, for instance, arc enriched in highly incompatible elements including the *large-ion lithophile elements (LILE)* such as Rb, Th, K, Ba, and light REE (fig 7.3). They also show relative depletions in Ta and N and, in some cases, also in Ti. The depletion in Ta–Nb relative to neighboring LILE is known as a *subduction zone component (SZC)*. Within-plate basalts exhibit an overall enrichment in all incompatible elements, and in some cases show a positive Ta–Nb anomaly (figs 7.3 and 7.6). This element pattern is known as a *within-plate component (WPC)*.

In addition to spidergrams, a large number of geochemical graphs have been described to distinguish igneous rocks from various tectonic settings (Pearce and Cann), 1973; Wood et al., 1979). These are often known as *tectonic discriminant diagrams* (see for instance figs. 7.8 and 7.11). None of these diagrams is completely successful in identifying tectonic setting, and they should not be used alone to characterize tectonic settings of ancient igneous rocks. Only when used in conjunction with other sources of information such as petrotectonic assemblages,

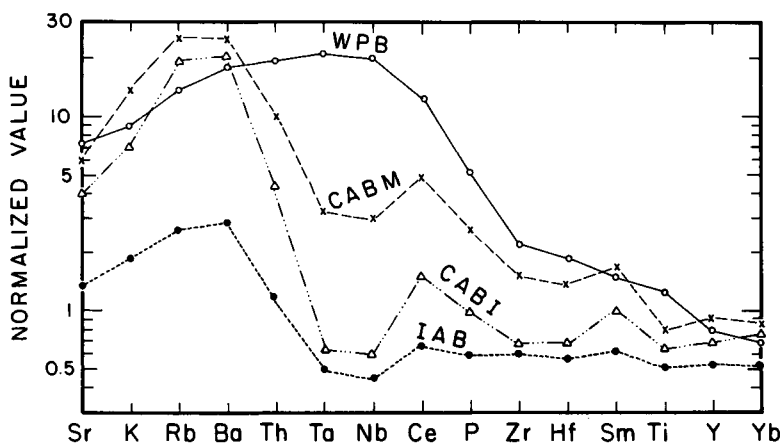


FIG 7.3 NMORB-normalized incompatible element distributions in basalts. Island-arc tholeiite (IAB); calc-alkaline basalts from island arcs (CABI) and continental-margin arcs (CABM); within-plate basalt (WPB). After BVTP (1981) and Pearce (1983).

metamorphic styles, sedimentary environments, and deformational characteristics, can tectonic discriminant diagrams be useful in constraining ancient tectonic settings.

Radiogenic and stable isotopes may be useful in characterizing magma sources. The growth of daughter isotopic ratios such as $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, and $^{143}\text{Nd}/^{144}\text{Nd}$ in source rocks varies with the parent-daughter element ratios (Rb/Sr, U/Pb, Sm/Nd, respectively) and with time. Thus, by measuring radiogenic isotopes in igneous rocks and correcting for the age of the rocks to obtain the initial ratios (see Chapt. 4), it is possible to characterize parent-daughter ratios of the sources. By using igneous rocks of different ages it is also possible to monitor changes in source composition with time. Stable isotopes such as oxygen, sulfur, and carbon may also be used in characterizing magma sources in that these isotopes (like radiogenic isotopes) are not significantly affected by magmatic processes such as fractional crystallization or batch melting.

Ocean Ridge Magmas

General Features

The most voluminous igneous rocks on Earth are the basalts erupted at oceanic ridges, known as *mid-oceanic ridge basalts* or *MORB*. MORB are erupted in the axial rift zones where they form new oceanic crust. Most are tholeiites showing varying degrees of depletion in LILE relative to chondrites. Relatively few samples of intermediate to felsic igneous rocks have been retrieved from ocean ridges. Dredged samples of MORB are vesicular, and exhibit glassy margins produced by rapid chilling during extrusion on the ocean bottom. They typically show pahoehoe surfaces that attest to low viscosity during eruption. They also commonly occur as pillows on the ocean floor. *Pillows* are rounded solidified masses of basalt (10–100 cm in diameter) that are produced by bulbous protrusions on submarine flows. They may detach themselves from parent flows and come to rest (while still hot and

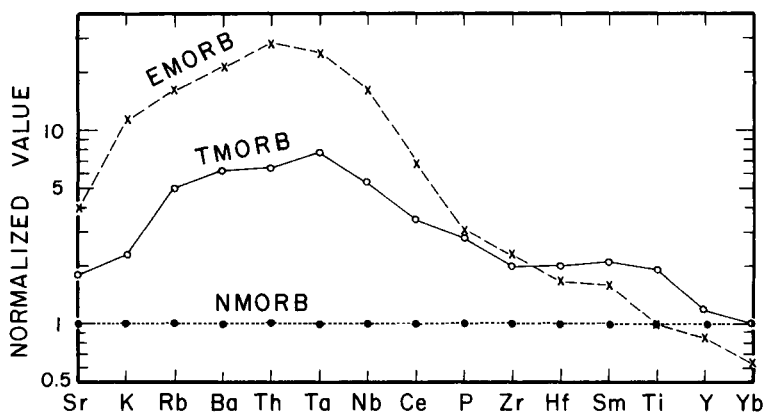


FIG 7.4 NMORB-normalized incompatible element distributions in ocean ridge basalts. NMORB, normal; TMORB, transitional; EMORB, enriched basalt. Data from Tarney et al. (1979).

plastic) on the seafloor. Texturally, ridge tholeiites are fine-grained and often contain large, sparsely distributed, partially resorbed plagioclase crystals. Smaller phenocrysts of olivine and plagioclase are common.

Composition

Except near the intersection of ocean ridges and islands, MORB are remarkably similar in major element composition throughout the ocean basins. They are characterized by low K_2O contents ($<0.3\%$). Other distinctive major elements include SiO_2 (49–51%), Al_2O_3 (14–17%) and MgO (6–12%). Most are olivine tholeiites and Mg numbers fall chiefly in the range of 55–65, indicating moderate amounts of fractional crystallization. It is well established that most MORB are depleted in LILE and especially in K, Rb, Cs, Sr, Ba, U, and light REE, and that they have low $^{87}Sr/^{86}Sr$ ratios (0.702–0.704) (BVTR, 1981). These basalts are known as *normal ocean ridge basalts* or NMORB. Some MORB are undepleted (TMORB) or show slight LILE enrichment (EMORB) as illustrated by an NMORB-normalized diagram (fig. 7.4). EMORB generally occur near seamounts or oceanic islands.

Geochemical studies of older MORB from various locations and depths in the oceanic crust (retrieved in part by the Deep Sea Drilling Project) indicate that deep-sea, low-temperature alteration is common. In general the degree of alteration seems to increase with distance from active ridge systems. Such alteration is characterized by varying enrichments in Fe^{+3} , K, H_2O , B, Li, Rb, U, and Cs, and sometimes in Sr, Ba, and light REE, and by increases in the amounts of zeolites and iron oxides (Christensen et al., 1973). Altered MORB typically exhibit $^{87}Sr/^{86}Sr$ ratios in the range 0.703–0.708. Measurements of density and seismic-wave velocities in ridge tholeiite samples exhibiting varying degrees of alteration also suggest that observed decreases in these quantities in layer 2 of the oceanic crust as a function of increasing distance from ridges may result from increasing degrees of alteration. In addition to low-temperature alteration ($<100^\circ C$), many MORB are affected by hydrothermal alteration ($>150^\circ C$) caused by seawater circulation at ocean ridges. Chemical

changes include enrichment in MgO and losses of SiO₂, CaO, K₂O, Rb, Cs, Ba, and Sr (Rona et al., 1983).

Within-plate Magmas

General Features

Within-plate (WP) igneous rocks occur in two major tectonic settings: oceanic islands and continental rifts. In addition, minor WP igneous rocks are formed in cratons. Volcanism within ocean basins occurs chiefly at volcanic islands and seamounts, and although individual volcanoes may have high magma production rates, total production rates of oceanic plus continental intraplate sites is $\leq 10\%$ of ocean ridge production rates (BVTP, 1981). Dredging of samplings from seamounts and from beneath sea level on islands suggests that volcanics from oceanic volcanoes differ from ocean-ridge volcanics in being composed dominantly of hyaloclastites, i.e., deposits of volcanic debris composed chiefly of broken volcanic glass and pillow fragments (Bonatti, 1967). Such differences indicate that island-seamount magmas are considerably more viscous than MORB magmas and readily fragment upon eruption into seawater. As discussed in Chapter 5, volcanic chains on the seafloor may result from the lithosphere moving over rising mantle plumes or hotspots. Although both the tholeiite and alkaline magma series occur in oceanic islands, the tholeiite series greatly dominates. In both series, basalts greatly exceed intermediate or felsic compositions. As exemplified by the Hawaiian–Emperor island-seamount chain (see fig. 5.22), volcanic chains may show a systematic change in magma composition with time. In Hawaii the oldest rocks from a given volcano are olivine tholeiites, followed by an increasing abundance of Fe-rich quartz tholeiites and terminating with minor volumes of alkali basalt and its differentiates. Such a sequence of events can be interpreted in terms of a hotspot model, in which early phases of magmatic activity reflect extensive melting in or above a mantle plume. As a volcano moves over the hotspot (still tapping the limbs of its magma chamber), less magma reaches the surface and shallow fractional crystallization becomes more prevalent, producing relatively more quartz tholeiites. As the volcano approaches the edge of the plume, isotherms drop and the depth and degree of melting decrease, resulting in the production of small volumes of alkali basalt. The fact that the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios do not appreciably change over this evolutionary history in Hawaii suggests that the same mantle source has been tapped throughout.

Igneous rocks in continental rifts range from extensive plateau or flood basalts (such as the Columbia River, Deccan Traps, and Karoo volcanics) to sporadically distributed small volcanic centers in passive rifts. At deeper crustal levels, such as exposed in the Oslo graben, granites are the dominant igneous rock. All of the volcanic rocks found in these areas share a similar tectonic environment characterized by extensional stresses. Magmas are commonly erupted from small to intermediate-sized cinder cones or, in the case of plateau basalts, from extensive fissure systems. Only rarely are large stratovolcanoes developed (such as Kilimanjaro in East Africa). Felsic magmas erupt as domes, plugs, or minor flows, the exception being the voluminous flows of phonolite and trachyte in the East African rift system. Most felsic magmas produced in rift systems, however, are probably intruded as high-K granites, syenites, and related rocks. These granitic rocks

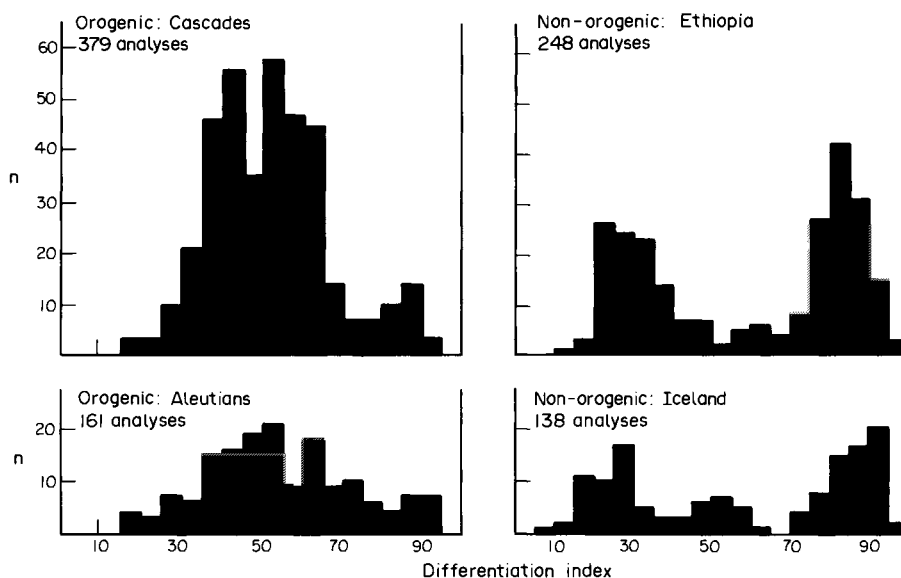


FIG 7.5 Frequency distribution of differentiation index in two calc-alkaline volcanic suites (Cascades, Aleutians) and two bimodal suites (Ethiopia, Iceland). After Martin and Piwinskii (1972). Copyright by American Geophysical Union 1972.

typically occur as small plutons in deeply eroded rifts such as the Oslo graben. Most plutons appear to be post-tectonic, in that they are not significantly deformed and have uniform textures and compositions. The so-called “anorogenic” granites (see Chapt. 9) belong to this group.

When large volumes of magma are erupted, large volcanic fields are formed such as the Columbia River Plateau in the northwestern United States, the Ethiopia field in eastern Africa, and the Deccan Traps in India (Plate I). In general, bimodal tholeiitic or alkaline suites characterize continental rifts, continental hotspots, and aulacogens. These suites include mafic and felsic magmas with a sparsity of intermediate members. This contrasts to the calc-alkaline series, in which intermediate members commonly dominate, as is illustrated in figure 7.5. In this figure, low values of the differentiation index are basalts, and high values are felsic volcanic rocks. The volume of bimodal volcanic rocks can vary from minor, as shown by the Basin and Range Province in the last 20 Ma, to quite large, such as found in the East African rift in Ethiopia. Commonly, felsic members are less abundant than mafic ones, a feature probably related to the greater viscosity of felsic magmas.

Igneous rocks injected into or erupted on continental cratons are rare. When found they are small intrusive bodies, dikes, sills, or volcanic pipes. The rocks are generally alkaline and may be bimodal. Kimberlites are of special significance in that they contain fragments of ultramafic rocks and eclogite from the upper mantle (Chapt. 3) as well as diamonds and other high-pressure minerals that indicate depth of origin for kimberlitic magmas of up to 300 km. Kimberlites are potassic ultramafic breccias containing larger grains of olivine, phlogopite, ilmenite, and pyroxenes in a fine-grained matrix of olivine, phlogopite, serpentine, carbonate, and other minor phases (Pasteris, 1984). They occur in all major Precambrian cratons

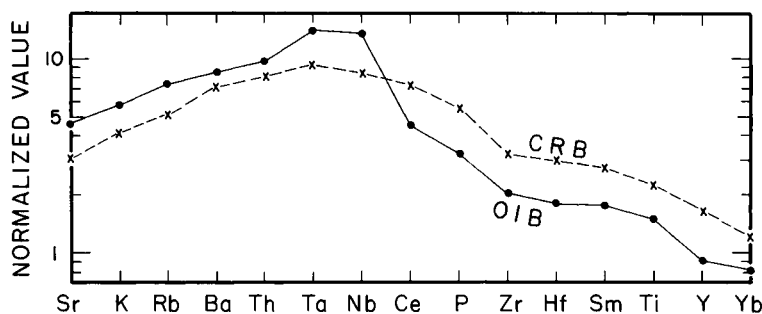


FIG 7.6 NMORB-normalized incompatible element distributions in within-plate basalts. CRB, continental rift basalt; OIB, oceanic island basalt. Data from BVTP (1981).

and range in age from about 1750 Ma to 20 Ma. Most kimberlites occur as pipes or dikes $<1 \text{ km}^2$ in cross-section and are commonly emplaced at fracture intersections. They may also be localized over hotspots and along postulated extensions of transform faults. Kimberlite magmas appear to have been produced by partial melting of an enriched upper mantle source (perhaps in the subcontinental lithosphere). The presence of CO_2 -rich fluid inclusions in mantle xenoliths from kimberlites is evidence for a CO_2 -rich fluid in their mantle sources.

Composition

Within-plate basalts (WPB) have variable major element contents but most have relatively low Mg numbers (40–60) suggesting degrees of fractional crystallization greater than most MORB. Variable and high SiO_2 and K_2O in some plateau basalts may reflect, in part, contamination with continental crust. There is a tendency in some rift systems for a zonation of volcanic types from the most SiO_2 -saturated types in rift valleys to the least saturated on the limbs and flanks of the rift. Such a zonation, for example, is reported in the northern Rio Grande rift, where tholeiites dominate in the floor of the rift and alkali basalts around the edges (Lipman, 1969). Both *oceanic basalts (OIB)* and *continental rift basalts (CRB)* are enriched in LILE, REE and HFSE compared to MORB (fig. 7.6). As mentioned above, this is known as a within-plate component. OIB, in addition, may show positive Ta–Nb anomalies on NMORB spidergrams. The within-plate component shown by WPB persists into more felsic within-plate igneous rocks. Strontium isotope ratios range from 0.702 to 0.706 in OIB and from 0.703 to 0.710 in CRB, with the high extremes of these ratios reflecting enriched sources of the basalts. Kimberlites and some of the ultramafic xenoliths they contain are highly enriched in LILE, indicating very enriched mantle sources.

Within-plate granites are typically *A-type granites* (table 7.4). Compared to ocean ridge and arc-type granites (M- and I-types), within-plate granites are enriched in REE and HFSE and do not show a subduction zone geochemical component (see figs 7.9 and 7.11). They also typically have high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and striking depletions in Sr, P, and Ti compared to most other granites.

TABLE 7.4 *Characteristic Features of Four Types of Granitic Rocks*

	M-Type	I-Type	S-Type	A-Type
Composition	Plagiogranite	Variable (tonalite-granodiorite dominate)	Leucogranites with narrow compositional range	High-K granite ($\pm$ syenite)
Characteristic Minerals ¹	hb, biot, cpx	hb, biot, mgt, sph	biot, cord, ilm, musc, gar, mon	biot, mgt, alkaline hb
Xenoliths	Mafic	Mixed	Chiefly metasediments	Mixed
Mol $\text{Al}_2\text{O}_3/\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO}$	≤ 0.6	0.5–1.1	> 1.1	0.9–1.1
Tectonic character	Pre-tectonic	Pre- to syntectonic	Syn- to post-tectonic	Post-tectonic
Ba, Ti, P depletion	Minor	Moderate	Moderate	Strong
SZC ²	Variable	Strong	Variable	Absent
$\delta^{18}\text{O}$	5.5–6	8–10	> 10	8–10
Tectonic Setting	Ophiolite (back-arc basin)	Arc or collisional orogen	Collisional orogen	Continental rift (?) (anorogenic)
Source	FXL of MORB	FXL of CAB (or IAB)	Partial melting of sediments	Partial melting of lower crust

¹ hb, hornblende; biot, biotite; cpx, clinopyroxene; mgt, magnetite; sph, sphene; cord, cordierite; musc, muscovite; gar, garnet; mon, monazite.

² Subduction zone component (Nb–Ta depletion relative to Th and light REE). FXL, fractional crystallization; CAB, calc-alkaline basalt; IAB, island arc basalt.

Subduction-related Magmas

General Features

Igneous rocks of the calc-alkaline series are typical of both island arcs and continental-margin arcs. The close relationship between active volcanism and descending plates implies a genetic connection between the two. Subduction-zone-related volcanism commonly starts abruptly at a *volcanic front*, which roughly parallels oceanic trenches and begins 200–300 km inland from trench axes adjacent to the arc–trench gap (see fig. 7.15). The volume of magma erupted decreases in the direction of dip of the subduction zone. The abrupt onset of volcanism at the volcanic front probably reflects the onset of melting in or above the descending slab, and the decrease in volume of erupted magma behind the volcanic front may be caused by either a longer vertical distance for magmas to travel or a decrease in the water content of the slab as a function of depth.

The common volcanic rocks in most arcs are basaltic andesites (immature oceanic arcs) or andesites (evolved oceanic arcs and continental-magma arcs). Andesites and basalts are usually erupted as flows in subaerial or submarine environments. Associated more felsic magmas are commonly erupted as plinian eruptions in which much of the ejecta is in the form of ash and dust. These eruptions give rise to ash-flows and associated pyroclastic deposits. Volcanic rocks are generally porphyritic containing up to 50% phenocrysts in which plagioclase dominates.

Arc volcanoes are typically steep-sided stratovolcanoes composed of varying proportions of lavas and fragmental materials (fig. 7.7). Their eruptions range from mildly explosive to violently explosive, and contrast strikingly to the eruptions of



FIG 7.7 Socompa volcano, a large andesitic stratovolcano in the Andes in Chile. A young andesite flow can be seen at the left. Courtesy of W. Zeil.

oceanic-island and continental-rift volcanoes. Large amounts of water are given off during eruptions. Rapid removal of magmas may result in structural collapse of the walls of stratovolcanoes, producing calderas such as Crater Lake in Oregon and the Valles Caldera in New Mexico. The final stages of eruption in some volcanic centers are characterized by the eruption of felsic ash-flows that may travel great distances. Seismic shadow-zone studies indicate that modern magma reservoirs in subduction-zone areas are commonly 50–100 km deep. The migration of earthquake hypocenters from depths of greater than 200 km over periods of a few months prior to eruption reflects the ascent of magmas at rates of 1–2 km/day (Scheinmann, 1971). The height of stratovolcanoes varies with magma composition (McBirney, 1976). The higher a volcano, the more felsic its composition. This appears to be due to increased fractional crystallization in areas of thickened crust in which volcanoes have higher base elevations due to isostasy.

The cores of arc systems are composed of granitic batholiths, as evidenced in deeply eroded arcs. Such batholiths, which are composed of numerous plutons, range in composition from diorite to granite, with granodiorite often dominating. Two factors seem to be responsible for granodiorite being the dominant plutonic component in arcs while andesites or basaltic andesites are the dominant volcanic component: magma viscosity and differences in andesite and granodiorite solidi. Andesite magmas are quite fluid and readily reach the surface, whereas felsic

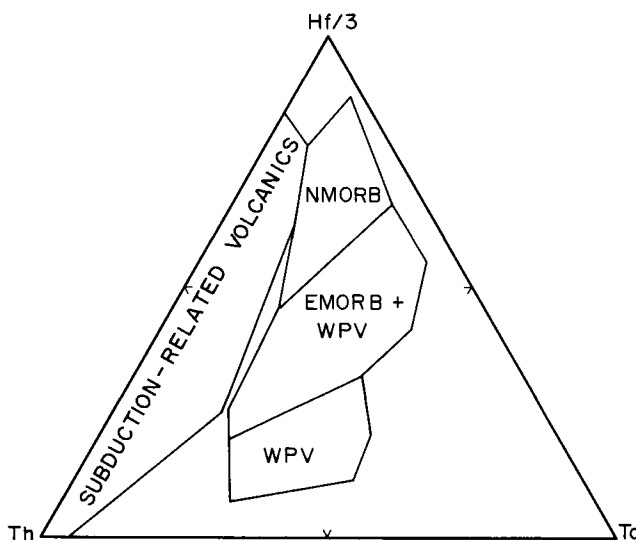


FIG 7.8 Th-Hf-Ta diagram showing distribution of major volcanic suites. Fields after Wood (1980). NMORB, EMORB, normal, and enriched ocean-ridge basalt, respectively; WPV, within-plate volcanics.

magmas are probably more viscous and tend to crystallize before reaching the surface. Perhaps more important is the fact that most felsic magmas are wet and intersect their solidus at depths of >5 km, while rising basalts and andesites do not intersect their solidus.

Composition

In contrast to WPB and MORB, arc basalts are commonly quartz normative. They generally contain 45–53% SiO_2 , high Al_2O_3 (16–20%) and low TiO_2 ($\leq 1\%$). $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in volcanics from island arcs are low (0.702–0.705) while those from continental-margin arcs are variable (0.703–0.710), reflecting variable contributions of continental crust to the magmas. Arc basalts, as previously mentioned, also exhibit a subduction zone component (SCZ), especially evident on MORB-normalized spidergrams. Two types of arc basalts are recognized: *island arc basalt* (IAB) and *calc-alkaline basalt* (CAB). IAB, which are common in immature island arcs such as the Kermadec and Mariana arcs, are characterized by moderate LILE enrichment, a small Ta–Nb anomaly, and relatively unfractionated REE and HFSE relative to NMORB (fig. 7.3). They also differ from MORB by their low Ni and Cr contents (table 7.2). CAB from island arcs (CABI) show greater LILE enrichment and a relatively large Ta–Nb anomaly, yet still relatively unfractionated heavy REE and HFSE, while CAB from continental-margin arcs (CABM) exhibit a gradual enrichment in incompatible elements from Yb to Ce (fig. 7.3). Andesites and felsic volcanics from arc systems exhibit similar NMORB-normalized element distributions to associated basalts, although elevated due to fractional crystallization. *Boninites*, which are high-Mg andesites that occur chiefly in immature oceanic arcs, have element distributions similar to IAB. Another graphical means of illustrating the SZC is by use of the Th–Hf–Ta diagram (fig. 7.8). This diagram clearly shows

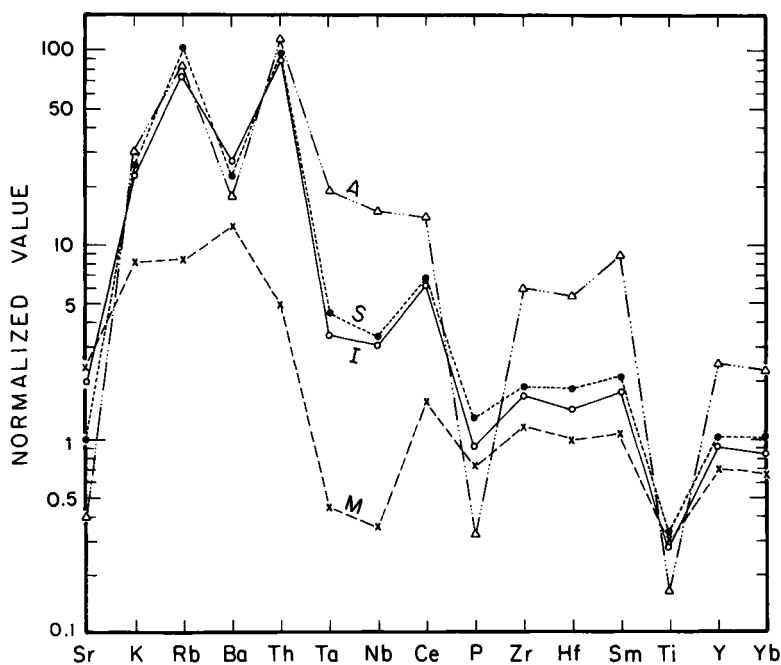


FIG 7.9 NMORB-normalized incompatible element distributions in granites. M, plagiogranite from ophiolite (immature island arc); I, arc-related granodiorite; S, S-type granite from SE Australia; A, anorogenic granite. Data are averages of many samples, in part from Whalen et al. (1987).

the depletion in Ta relative to Th and Hf in subduction-related volcanic rocks. It also has the advantage that the three elements are relatively immobile and igneous rocks of all compositions may be plotted on the figure because the distribution of points is rather intensive to fractional crystallization. The chief disadvantage of the Th–Hf–Ta diagram as a tectonic setting discriminator is that continental within-plate volcanics often plot in the Th-rich part of the subduction field.

Based on K_2O – SiO_2 relationships, the calc-alkaline series can be divided into several subseries which show increasing amounts of K_2O (and other LILE) at a given SiO_2 level (Peccerillo and Taylor, 1976). There are the low-K, medium-K, and high-K calc-alkaline subseries, the shoshonite subseries and the leucitite subseries. All are known from both island arc and continental-margin arcs. Volcanic rocks erupted in back-arc basins show variable geochemical characters with the SZC ranging from moderate to absent and magma compositions may change with time. For instance, as an oceanic back-arc basin opens, basalts may change from CAB to MORB-like in character.

Arc granites are chiefly I- or M-type granites in terms of the characteristics in table 7.4. *M-type granites* (plagiogranites) occur in many ophiolites and appear to have formed in immature oceanic arcs. *I-type granites* are typically meta-aluminous with tonalite or granodiorite dominating, and are the typical granitic rocks formed in most arc systems. Geochemically, arc-related granites are enriched in LILE and show prominent Ta–Nb anomalies (fig. 7.9). Most granites also show negative Sr, P, and Ti anomalies.

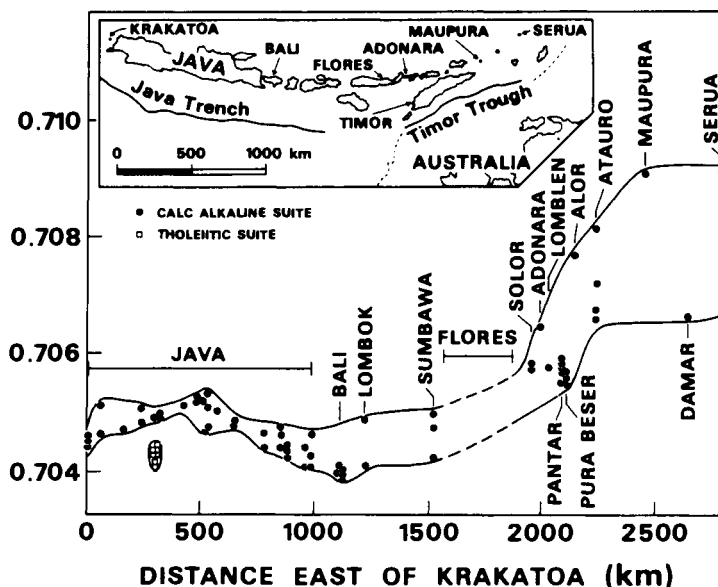


FIG 7.10 Sr isotope variations in young volcanic rocks along the Sunda-Banda arc, Indonesia (after BVTP, 1981).

Spacial and Secular Compositional Variation in Arcs

Geochemical studies indicate that igneous rocks from arcs vary in both space and time. It has long been recognized that some arc volcanics show increasing LILE and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios with increasing depth to subduction zone (Dickinson, 1970). This change in composition is known as *compositional polarity* and is well documented in Japan (Sakuyama and Nesbitt, 1986). Because many or perhaps most arcs do not show compositional polarity (Arculus and Johnson, 1978), the linear relationship between LILE content and depth to subduction zone proposed by Dickinson (1970) should not be used to determine polarity of ancient subduction zones unless supported by other data. Arc volcanics also exhibit lateral compositional changes, as evidenced for instance by the $^{87}\text{Sr}/^{86}\text{Sr}$ in the Sunda-Banda arc (fig. 7.10). In progressing eastward in this arc the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio increases, especially in the east of Flores. Geochemical studies suggest the arc can be divided into four geochemical segments with boundaries that correlate with changes in tectonic setting and geologic history (Wheeler et al., 1987).

Some oceanic arcs seem to show a geochemical evolution with time. The best-documented case is in Fiji where the earliest volcanics (25–30 Ma) are relatively depleted in LILE (Gill, 1987). Later volcanics are more typical of evolved arcs and show moderate LILE enrichment and a strong SZC. Most oceanic arcs do not show this simple evolution but reflect a complex and often unknown plate history, in some cases involving one or more arc collisions.

Collision-related Magmas

General Features

Igneous rocks produced in collisional mountain belts are chiefly calc-alkaline granitic plutons. Because of the strong compressional forces, most magmas do not

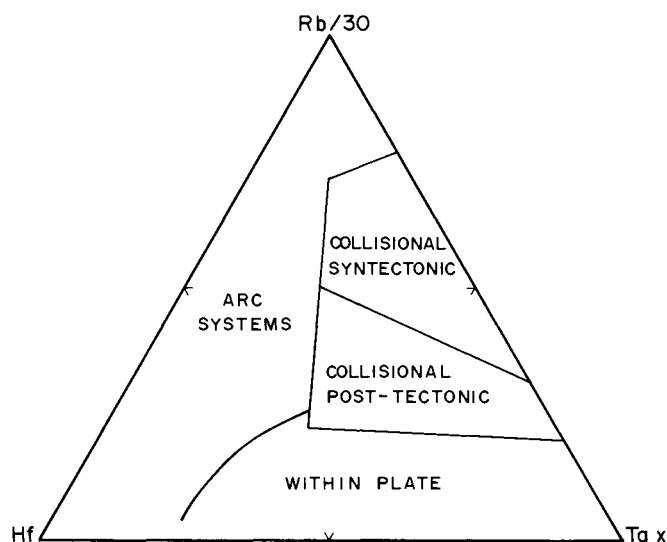


FIG 7.11 Rb–Hf–Ta diagram showing fields of granites from various tectonic settings. After Harris et al. (1986).

make it to the surface before crystallizing. Collisional granites include pre-, syn-, and post-collisional types (Pearce et al., 1984). Pre- and post-collisional types are chiefly I-type granites. Syn-collisional granites are generally leucogranites and many exhibit features of *S-type granites* (table 7.4). In particular, they are commonly peraluminous and often contain muscovite with or without biotite (Harris et al., 1986). Tourmaline may also be present. Silica contents of leucogranites generally exceed 70% and they are emplaced at crustal depths where migmatites form by anatexis. Post-collisional granites are sharply cross-cutting and are dominantly tonalite to granodiorite in composition. They generally post-date collision by 40–50 Ma.

Composition

Collisional granites generally exhibit a SZC when plotted on MORB-normalized plots (for instance S-type in fig. 7.9). The distributions of Rb, Ta, Hf, Y, Yb, and Nb are distinctive in most collisional granites (Pearce et al., 1984; Harris et al., 1986). As an example, the fields of Himalayan collisional granitic rocks are shown on figure 7.11 in terms of Rb, Hf, and Ta distributions. Fields of arc and within-plate granites from other areas are also shown for comparison. Arc-related granites are relatively depleted in Ta (the SZC previously described), collisional granites (both syn- and post-collisional types) are relatively depleted in Hf, and within-plate granites have relatively low Rb contents. In addition, collisional granites have very high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (up to 0.8) and high $\delta^{18}\text{O}$ values, indicating a crustal origin (Vidal et al., 1984).

Magma Production

Introduction

Seismic, heat flow, and experimental petrologic data indicate that basaltic and ultramafic magmas are generated in the upper mantle. Melting in magma source

areas requires thermal energy of about 100 cal/gm of magma produced. This energy may be supplied directly from local concentrations of radioactive elements in the mantle, energy given off during earthquakes or by upward convection currents or plumes from deeper in the mantle. In subduction zones, frictional heating along the seismic zone in the descending slab may contribute to melting. Since the melting point of silicates increases with load pressure, a decrease in pressure caused perhaps by faulting in a descending slab can reduce the effective melting temperature at a given depth. Water depresses the melting point of most silicates and hence an increase in water content (caused, for instance, by the breakdown of hydrous minerals carried into the mantle in subduction zones) can also result in melting. Although all of these mechanisms may operate in varying degrees in the generation of magmas, heat brought up by convection is probably chiefly responsible for melting beneath ocean ridges; melting in or above descending slabs appears to reflect chiefly dehydration of descending oceanic crust and frictional heating along the top of the slab; melting in the mantle beneath within-plate sites may be due to adiabatic melting of plumes as they rise in the mantle or to heat liberated from such plumes causing partial melting of the lithosphere (McBirney, 1980).

Data from the fields of experimental petrology and geochemistry provide important boundary conditions for the origin and evolution of magmas. It is possible in the laboratory to reconstruct the P-T regimes in the Earth where magmas are produced. It is also possible to vary water content and oxygen fugacity and to use starting compositions similar to those expected in the upper mantle and lower crust (Chapts 3 and 4). The upper and lower stabilities of minerals can be determined under a variety of conditions, and the composition of melts coexisting with various minerals can be determined by analyzing glasses produced by rapid chilling. The results of experimental petrologic studies that represent closed systems must be carefully evaluated in terms of the open or partially open systems that exist in the Earth. The composition of magmas produced in the mantle is dependent upon the composition of the source of rock and the degree of melting. The degree of melting is, in turn, dependent upon temperature, pressure, and water content of the source. Higher temperatures and water contents result in larger degrees of melting, whereas greater pressures may result in smaller degrees of melting. Large degrees of melting of upper mantle rocks (chiefly ultramafic rocks) produce tholeiitic or komatiitic magmas, whereas small degrees produce alkaline mafic magmas. Results of experimental data for the origin of basaltic magmas by partial melting of the upper mantle are summarized in figure 7.12. Important primary magmas are olivine tholeiite and picrite, which require 15–30% melting at depths of 30–80 km. Although tholeiite and quartz tholeiite can be produced by partial melting at shallower depths, it is unlikely that upper-mantle temperatures are high enough (except at ocean ridges) to cause melting at these depths; hence these magmas are generally thought to be the products of shallow fractional crystallization (Jacques and Green, 1980). Very small amounts of melting of garnet lherzolite at depths greater than 50 km can give rise to alkali basalts, nephelinites, and kimberlites. Komatiites are probably produced by $\geq 35\%$ melting at depths < 70 km and perhaps as deep as 200 km. With a few minor exceptions, only in the Archean were geotherms steep enough to facilitate such large degrees of melting.

Melting can be described in terms of two idealized processes that represent end-

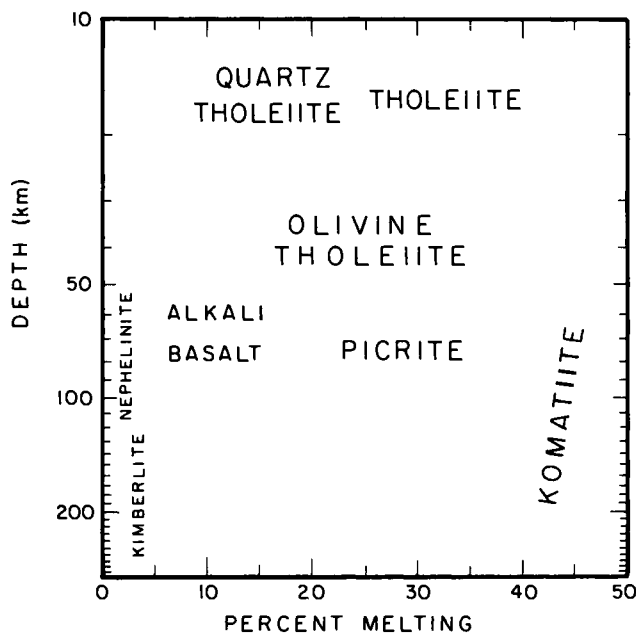


FIG 7.12 Pressure-temperature regimes of basaltic and komatiitic magmas (for $\leq 1\%$ H_2O). Magmas produced at depths >40 km depth are chiefly partial melts of ultramafic sources, whereas those produced at <40 km are products of fractional crystallization.

members of a continuum of melting conditions. *Equilibrium* (or *batch*) *melting* occurs when the melt and the residual solids remain in equilibrium until the melt is extracted from the source, and *fractional melting* occurs when magma is continually extracted from the source as melting proceeds. More complicated melting, such as *dynamic melting*, involves successive removal of melt increments as melting proceeds, and trapping of small amounts of liquid in the source. The composition of magmas may be changed by such processes as fractional crystallization, wall-rock reaction, and crustal contamination. *Fractional crystallization* is the process by which crystals are removed from a crystallizing magma by gravitational sinking or floating (or some other process). It can be considered in terms of closed- and open-system models. *Open-system models* allow injection of new magma into the fractionating reservoir whereas *closed-system models* do not. *Liquid fractionation* is caused by crystallization along the sides of large magma chambers resulting in decreases in density of adjacent magma. This lower density, more felsic magma rises to the top, producing a zoned magma chamber (McBirney, 1980). *Wall-rock reaction* is the process by which incompatible elements become preferentially enriched in a magma at or near the source by selective migration of these elements out of wall rocks. Wall-rock reaction can also occur at shallower depths provided the magmas remain in chambers at such depths for time periods long enough for significant wall-rock reaction to occur. *Crustal contamination* occurs when a magma passes through the continental crust and partially digests crustal rocks, or when crustal rocks are involved in the generation of magmas. It is often possible to quantitatively evaluate each of these processes using results from experimental petrology and the distribution of major and trace elements and isotopes in igneous rocks (Allegre and Minster, 1978; Zindler and Hart, 1986).

Following is a brief review of the major constraints on magma production in each major tectonic setting.

Ocean-ridge Magmas

Most experimental data indicate that MORB are produced by 15–30% partial melting of the upper mantle at depths of 50–85 km (Elthon and Scarfe, 1984). Melting at these depths produces olivine tholeiite magmas. Many MORB, however, are tholeiites or quartz tholeiites that define sub-linear arrays on normative phase equilibria diagrams around the low pressure (<10 kb) olivine-plagioclase cotectic. This is generally interpreted to indicate an origin by shallow fractional crystallization of olivine tholeiite. Trace element and isotopic data are also consistent with this model. It would appear that MORB are produced by partial melting of the asthenosphere (garnet lherzolite) in convective upcurrents beneath ocean ridges at depths ≥ 50 km. Much of this magma collects in shallow chambers (<35 km) where it undergoes fractional crystallization to produce tholeiites and quartz tholeiites, and minor amounts of more evolved liquids including plagiogranites. A residue of olivine and pyroxenes is left behind in the mantle and may be represented by the sheared harzburgites preserved in some ophiolites.

The depleted light-REE (fig. 7.2) and other incompatible elements in NMORB indicate a mantle source that has been depleted in these elements by earlier magmatic events. Low $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$, and high $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratios also demand a source that is depleted relative to chondrites. Incompatible element contents and isotopic ratios vary in MORB along ocean ridges and from ocean to ocean, as illustrated by the distribution of Nb and Zr in modern MORB from the North Atlantic basin (fig. 7.13). Although some of this variation can reflect differences in degree of melting of the source, fractional crystallization or magma mixing, most requires variation in the composition of the depleted MORB source (Tarney et al., 1979). The greater incompatible element contents of TMORB and EMORB also require source heterogeneity.

Within-plate magmas

When considered on normative phase diagrams, within-plate basalts (WPB) show much greater variability than MORB. Parental magmas appear to be picrites or olivine tholeiites produced either in rising mantle plumes or in the mantle lithosphere. Basalt compositions reflect variable depths of fractional crystallization. For instance, Karoo flood basalts appear to have fractionated at depths of 20–30 km in the Kaapvaal craton, while Columbia River basalts fractionated at depths ≤ 20 km (Thompson et al., 1983). The low Mg numbers and Ni contents in most WPB suggest extensive fractional crystallization of olivine and pyroxenes.

The WPC (i.e. enrichments in LILE, REE and HFSE) that characterizes all within-plate igneous rocks demands one or more sources that are enriched in incompatible elements relative to the NMORB source. Possible origins for these enriched sources are discussed in the next section. A deepening of melting on the flanks of ocean ridges is compatible with probable temperature distributions in oceanic areas (fig. 3.7), which indicate that geothermal gradients decrease away from

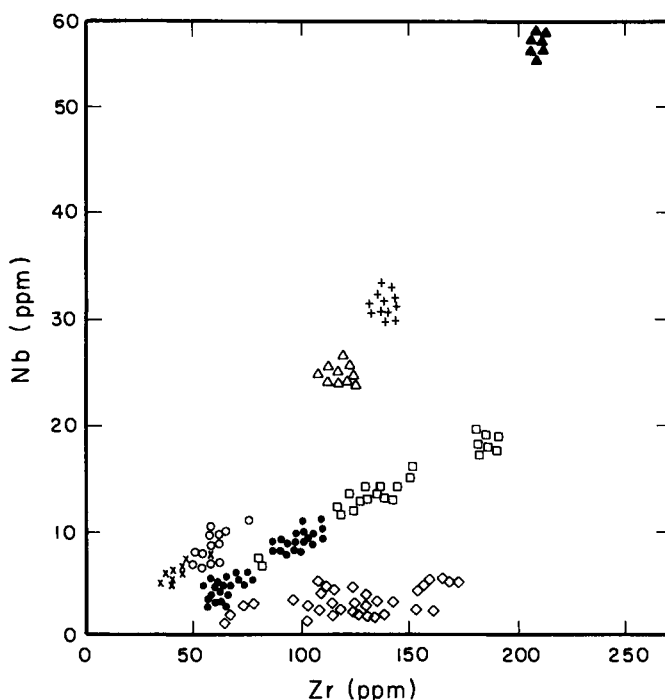


FIG 7.13 Distribution of Zr and Nb in modern ocean-ridge basalts from the North Atlantic. Each symbol represents a different geographic location on the Mid-Atlantic ridge. Modified after Tarney et al. (1979).

ridges. This suggests that melting occurs at greater depths and smaller volumes of magma should reach the surface. The degree of melting controls the composition of magma produced ($<15\%$ = alkaline basalts; $>25\%$ = tholeiites) and very small degrees of melting can explain the large enrichment of incompatible elements in alkali basalts. The absence of Eu anomalies in WPB indicates that plagioclase did not play a major role in their production. Differences in MORB and OIB sources are most pronounced in the transition zones between ocean ridges and islands (Schilling et al., 1983). As illustrated for the Reykjanes ridge south of Iceland (fig. 7.14), La/Sm , $^{206}\text{Pb}/^{204}\text{Pb}$, $^{208}\text{Pb}/^{204}\text{Pb}$, and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios increase significantly in MORB as the coast of Iceland is approached. Similar changes occur in other incompatible element contents and ratios. These changes are common in MORB of the Atlantic basin as islands, such as the Azores and Tristan da Cunha, are approached. Element and isotope anomalies also occur in MORB along strike from islands that have been displaced from the Mid-Atlantic ridge by seafloor spreading (Schilling et al., 1985). These transitions in isotope and incompatible element distributions cannot result from different degrees of source melting or fractional crystallization. Although they are generally interpreted to reflect mixing of the depleted MORB source with an enriched OIB source, simple binary mixing of two-end-members does not explain all geochemical data.

Trace element and isotopic compositions of A-type granites are consistent with an origin for these rocks by partial melting of a crustal source or, in some instances, fractional crystallization of WPB. In either case their WPC (fig. 7.9) indicates they

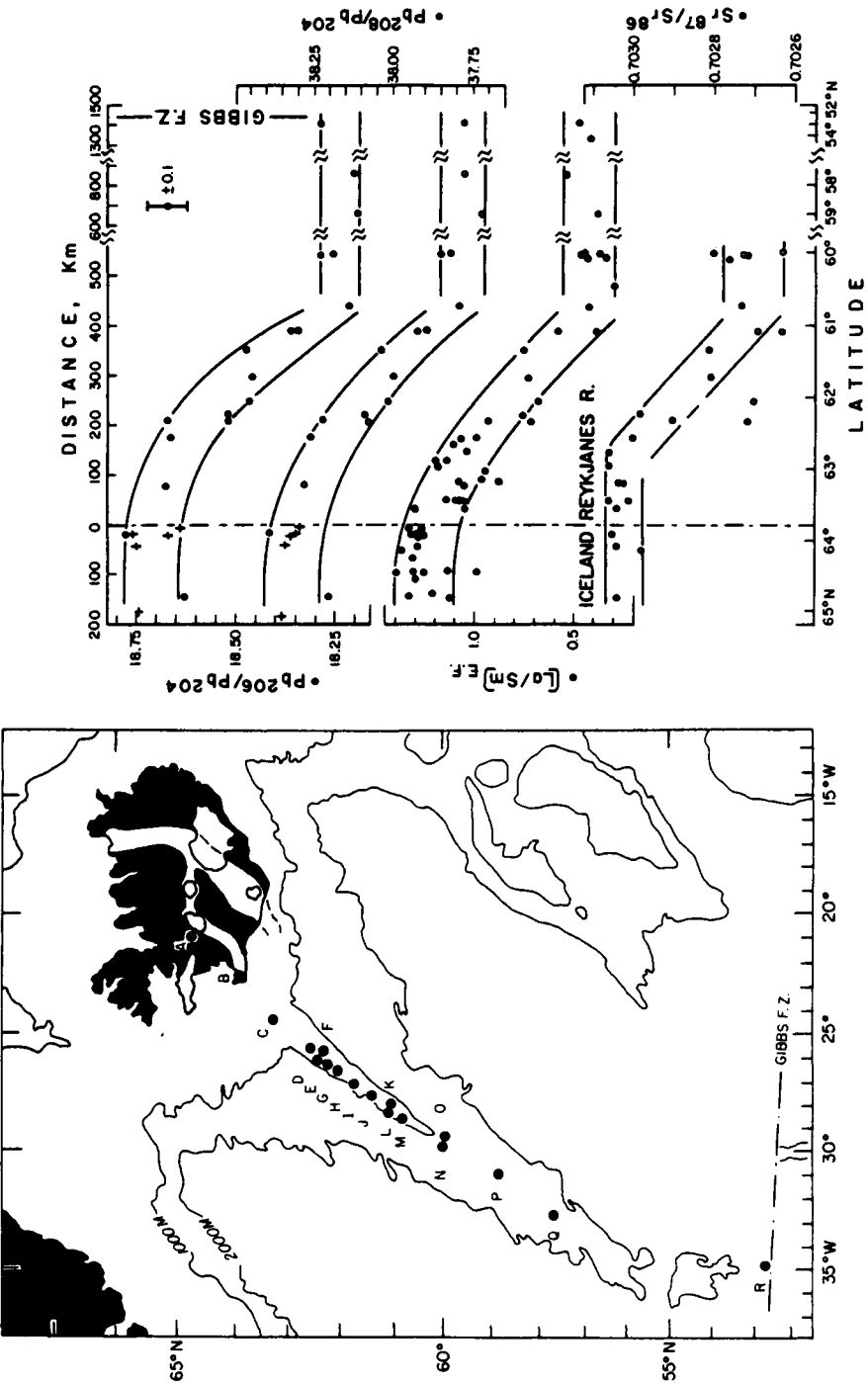


FIG 7.14 Variation in Pb and Sr isotope ratios and in La/Sm ratio in basalts from Iceland and the Mid-Atlantic Ridge south of Iceland (from Sun et al., 1975). $[La/Sm]_{E.F.}$ = chondrite-normalized La/Sm ratio.

TABLE 7.5 *Contrasting Features of Wedge and Slab Sources for Subduction-related Basalts¹*

	Mantle Wedge (Lherzolite) ²	Descending Slab (Eclogite)
1. Major elements	Low Mg numbers require significant low-pressure fractional crystallization	Models predict SiO ₂ contents higher than observed in IAB or CAB
2. Phase equilibria	Olivine may not be a liquidus phase, a condition which is necessary if IAB–CAB are derived from an ultramafic parent	Absence of liquidus olivine consistent with garnet-cpx source rock
3. Degree of melting required	20–30%	≥50%
4. Subduction zone component (SZC)	Acquired by devolatilization of descending MORB; Ta–Nb retained in residual phase(s)	Acquired in slab (altered MORB) or by wall-rock reaction in the mantle wedge; Ta–Nb retained in residual phase(s)
5. Depth of magma separation	≈100 km	85–100 km

¹ After Brophy and Marsh (1986).² Garnet or spinel lherzolite.

ultimately came from enriched mantle sources. Their striking depletions in Sr, P, and Ti appear to have developed by fractional crystallization involving feldspars (Sr), apatite (P), and magnetite-ilmenite (Ti).

Subduction-related Magmas

The following processes may play a role in the production of subduction-related magmas, and most magmas have the imprints of more than one process.

1. partial melting of a mantle source
2. fractional crystallization
3. a contribution from subducted sediments
4. contamination by continental crust
5. derivation from continental crust
6. magma mixing.

Geochemical and isotopic constraints on each process are reviewed in the following sections.

The Mantle Source

There is disagreement among petrologists as to whether arc basalts (IAB, CAB) are produced by partial melting of the mantle wedge or of the descending slab. A comparison of experimental and geochemical features of each source is summarized in table 7.5. Most data are consistent with either source. A major problem with the slab model, however, is the large amount of garnet left in the eclogite melting residue. This results in significant depletion in heavy REE and Y in derivative magmas, a feature which is not characteristic of IAB and most CAB. Melting in

excess of 50% is required to decrease residual garnet to acceptable values, and this amount of melting is unrealistic in terms of experimental data which indicate that melts separate from residues before 40% melting is reached. Not favoring an ultramafic source in the mantle wedge are results of an experimental study of an IAB sample (Mg No. = 60) which indicate that olivine is not a solidus phase as it should be if a magma is derived by partial melting of an ultramafic source (Brophy and Marsh, 1986). This problem may, however, be due to cumulus plagioclase in the IAB sample, which if removed would result in olivine being a solidus phase. Although the source problem is not resolved, most data seem to favor ultramafic rocks in the mantle wedge as the major source of arc basalts.

Fractional Crystallization

Normative compositions of arc basalts interpreted in terms of phase equilibria, low TiO_2 contents, and a rapid decrease in the $\text{CaO}/\text{Al}_2\text{O}_3$ ratio with decreasing Mg number, suggest that they have undergone shallow to intermediate-depth (<50 km) fractional crystallization with olivine and clinopyroxene $\pm$ magnetite as the most important crystallizing phases (Perfit et al., 1980; BVTP, 1981). This is a contrast to MORB where plagioclase is an important crystallizing phase. Also consistent with some magnetite removal is the fact that arc tholeiites do not show the large iron enrichment characteristic of MORB (fig. 7.1). Numerous geochemical studies of more evolved arc volcanics (i.e. andesites, dacites, rhyolites) indicate that these rocks are related to associated IAB or CAB by shallow fractional crystallization (Gill, 1981; Foden, 1983), which accounts for the continuous compositional series formed in most arcs. The similar incompatible element distributions and isotopic ratios in evolved rocks and associated basalts is also consistent with this origin. Modeling of boninites indicates they are derived from partial melting of a depleted mantle source that was later enriched in LILE (Hickey and Frey, 1982). Such a source may be the oceanic mantle lithosphere.

The close association of tholeiitic and calc-alkaline volcanic suites in the same arc system (sometimes at the same volcanic center) presents another problem. It is possible, because the mafic end-members of these two suites are geochemically indistinguishable, that both suites have the same mantle source and the suites deviate from each other in more evolved members due to different fractional crystallization schemes (Kay et al., 1982). The tholeiite series appears to have crystallized at shallow depths under relatively dry conditions and high temperatures. Minor magnetite crystallization produces moderate iron enrichment trends (fig. 7.1). In the Aleutian arc, tholeiitic volcanic suites occur between arc segments, probably above faults that may act as conduits for rapid transport of magma to the surface. This could account for the sparsity of intermediate and felsic members in the tholeiite series. Calc-alkaline magmas, on the other hand, appear to have fractionated in deeper chambers and with greater water contents, as evidenced by the presence of hydrous phenocrysts; magnetite crystallization suppressed any iron enrichment. Thus, differences in depth of crystallization and water content may be responsible for the two magma series found in arc systems.

Subducted Sediments

An important question related to the origin of arc basalts is whether subducted

sediments contribute to magma production. Incompatible element distributions, isotopic ratios, and ^{10}Be have been used to study this problem. Since both pelagic sediments and continental sediments can end up in trenches, subduction of both types of sediments needs to be considered. Because continental terrigenous sediments have significant negative Eu anomalies on chondrite-normalized plots, yet mafic (and intermediate) arc volcanics do not, a maximum of 10% subducted continental sediment can contribute to a MORB source (or to arc magmas) before a Eu anomaly should be seen in derivative melts (McLennan and Taylor, 1981). Furthermore, the absence of significant negative Ce anomalies in most arc volcanics limits the contribution of pelagic sediments to <1%, because pelagic sediments have major Ce anomalies. Sr, Pb, and Nd isotopic data, however, may allow a greater contribution of terrigenous sediments to arc magmas as attested to by high $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$, and low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in mafic volcanics from the Lesser Antilles arc (Davidson, 1983). The rapid increases in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in volcanics of the eastern part of the Sunda-Banda arc (fig. 7.10) appear to reflect subduction of terrigenous sediments derived from the Precambrian Australian craton. ^{10}Be is a radiogenic isotope produced in the atmosphere and in surface rocks by cosmic ray spallation of nitrogen and oxygen. Atmospheric ^{10}Be is rapidly removed into the oceans where it has a mean residence time of about 1600 years before being incorporated in pelagic sediments. If these sediments are subducted and contribute to arc magmas within a few half-lives of ^{10}Be ($t_{1/2} \approx 1.5$ Ma) and the magmas are erupted, it is possible to constrain the contribution of sediments. ^{10}Be activities in modern volcanics from several arc systems suggest variable but small contributions of pelagic sediments to arc magmas (Brown et al., 1982).

In conclusion, geochemical and isotopic data from arc volcanics indicate variable but, on the whole, small or negligible contributions of subducted sediments to arc magmas. Specific incompatible elements from subducted sediments (such as Sr, Pb, and Nd), however, may contribute to arc magmas in greater amounts.

Contamination by Continental Crust

Trace element and isotopic systems cannot distinguish between a subducted sediment contribution to arc magmas and continental contamination. If arc magmas reside in the continental crust during fractional crystallization for significant amounts of time, they may become contaminated with LILE by wall-rock reaction or assimilation. However, the same geochemical results discussed under sediment subduction would seem to limit the amount of continental contamination to a few percent, although selective LILE contamination could be much greater.

Continental Sources

The very high Sr and Pb, and low Nd isotopic ratios in some felsic volcanics and granitic batholiths from continental-margin arcs, such as those in the Andes, suggest that these magmas were produced either by partial melting or a large degree of contamination by significantly older continental crust. Incompatible elements, and particularly LILE, are also strongly enriched in these rocks, a feature which is consistent with a major crustal component in their source. An alternate explanation

for such felsic igneous rocks is that they have been produced by a two-stage process from an old, enriched mantle lithosphere. For instance, partial melting of an ancient enriched mantle lithosphere could produce basaltic magmas that may undergo fractional crystallization to form felsic magmas. A modern example of voluminous arc-related rhyolites is those in the Taupo area of New Zealand. Major and trace element data, as well as oxygen and Sr isotopic data, support an origin of these rocks by partial melting of Cretaceous graywackes (Cole, 1981).

The SZC (i.e. Ta–Nb depletion relative to LILE) in arc-related felsic igneous rocks must be inherited from their sources. Since continental crust exhibits a SZC, derivative magmas would also show a SZC. In the case of a mantle lithosphere source it may be necessary for the SZC to be superimposed on a WPC, perhaps during an earlier subduction regime.

Magma Mixing

Mixing of contemporary felsic and basaltic liquids has been called upon to produce andesites (Eichelberger, 1978). In some instances it is possible to explain major and some trace element distributions by mixing. However, rarely is it possible to explain all trace element and isotopic ratios by mixing. Also, experimental and theoretical considerations indicate that magmas with significantly contrasting densities (such as basalt and rhyolite) do not readily mix (McBirney, 1980). Observational evidence where felsic and mafic magmas have been brought into contact does not support significant mixing. Although local magma mixing, especially between magmas with similar densities, may be important, it appears unlikely that the large volumes of andesite and dacite formed in arc systems are the products of magma mixing.

Some Outstanding Problems

One unsolved problem related to the production of arc magmas is that of how the subduction zone geochemical component (SZC) is acquired by magma sources. Whatever process is involved, it requires decoupling of Ta–Nb, and in some cases Ti, from LILE and REE. If the mantle wedge is the major source of magmas, it may acquire a SZC by devolatilization of descending oceanic crust. Liberation of aqueous fluids from oceanic crust may carry REE and LILE, which are soluble in such fluids, into the overlying mantle wedge (Gill, 1980; Saunders et al., 1980). Partial melting of the descending slab may also aid in transmitting these elements into the wedge. Ta and Nb may be left behind in the slab in a residual phase such as rutile, sphene, or ilmenite, or some other minor phase whose stability field increases in the presence of water. This phase may also be residual if the descending slab (as eclogite) undergoes partial melting. The net result of such a process is liberation of LILE and REE into the mantle wedge and concentration of Ta–Nb in the slab, and magmas generated in the mantle wedge should inherit this SZC. Experimental data do not favor rutile as the residual phase, however, in that derivative magmas must contain $\geq 7\%$ TiO_2 before rutile is stabilized and arc basalts contain $< 2\%$ TiO_2 (Ryerson and Watson, 1987).

Experimental data indicate that hydrous secondary minerals in descending oceanic crust (such as chlorite, biotite, amphiboles, talc) break down and liberate water by about 100 km burial depth. Only phlogopite may persist to greater depths.

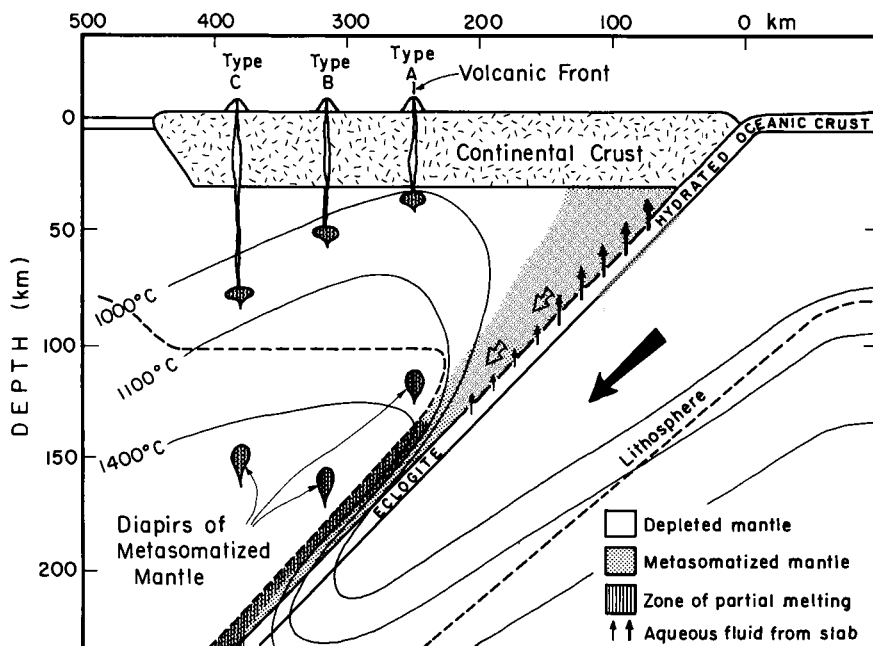


FIG 7.15 Hypothetical cross-section of a subduction zone showing shallow devolatilization of descending oceanic crust and magma production in ascending diapirs. Modified after Sakuyama and Nesbitt (1986).

Yet the volcanic front generally appears at subduction-zone depths of 125–150 km (fig. 7.15). Devolatilization of descending slabs by 100 km depth should add a SZC only to the mantle wedge above this segment of the slab, which is shallower than the depth of partial melting. How then, do magmas acquire a SZC in the wedge? Two possibilities have been suggested, although neither has been fully evaluated: (1) asthenosphere convection extends into the “corner” of the mantle wedge and carries ultramafic rocks with a SZC to greater depths; and (2) frictional drag of the descending plate drags mantle with a SZC to greater depths (fig. 7.15).

Also remaining unsolved is the question of compositional polarity: why does it occur in some places and not in others, and what causes it? Increasing degrees of fractional crystallization or decreasing degrees of melting as a function of increasing depth to subduction zone do not seem capable of accounting for the large magnitude of changes observed in LILE concentrations. If the mantle wedge is stirred by convection so that all of it has a SZC, increasing magma path lengths between the slab and the surface may result in more LILE enrichment of magmas formed at greater depths by wall-rock reaction. Alternately, partial melting just above the slab may give rise to diapirs (melt and solid) that melt adiabatically as they rise towards the surface (Sakuyama and Nesbitt, 1986). These diapirs may stop as the viscosity of surrounding mantle increases with the diapirs closest to the trench reaching the shallowest depths (Type A, fig. 7.15) having the greatest degree of melting, and hence the lowest LILE contents. Magmas farthest from the trench (Type C) require <10% partial melting to explain their high LILE contents. Neither of these models, however, satisfactorily explains why polarity occurs in some but not all arcs with dipping slabs.

Komatiites

Komatiites, which are only important in the Archean, have experimental liquidus temperatures of up to 1650°C (Arndt, 1986). As discussed more fully in Chapter 10, they may be produced beneath ocean ridges with geotherms steeper than those beneath most modern ocean ridges. Alternately, or in addition, they may be produced in mantle plumes that rise rapidly from the deep mantle. Although only the steeper geotherm model requires hotter mantle, both of these models are consistent with a mantle 100–300°C hotter than at present.

Collision-related Magmas

Pre-collisional granites are I-type granites probably formed by fractional crystallization of mafic magmas produced in the mantle prior to collision. The absence of pre-collisional granites in some orogens, such as the Alps and Hercynides, may reflect closure of fairly small ocean basins. Isotopic and trace element distributions in syncollisional granites indicate they are largely the products of partial melting of the crust and, in some instances, the source is metasediments (Vidal et al., 1984; Harris et al., 1986). Volatiles given off by underthrust sediments may penetrate the overlying crust causing partial melting. Resulting melts are enriched in Rb, F and B (and some in Ta) which are carried in the fluid phase, and depleted in Y, heavy REE and HFSE, which are left in the residues of melting.

Post-collisional granites comprise suites with isotopic and geochemical characteristics similar to pre-collisional granites. They appear to be derived from basalt fractional crystallization during continued convergence of the two colliding plates.

Bimodal Magmatism

Bimodal igneous suites, which contain an abundance of mafic and felsic end-members, are common in continental rifts and in Precambrian lithologic assemblages (see Chapt. 9). Numerous mechanisms have been proposed for the production of bimodal magmas, and it is probable that bimodal suites have more than one origin (Hildreth, 1981). The following mechanisms are most commonly cited for the production of bimodal magmas.

Immiscible Liquid Segregation

Volcanic rocks sometimes contain glassy globules (10–50 mm across) of coexisting mafic and felsic compositions. The felsic globules fall near the granite minima in composition, while the mafic ones are generally Fe-rich pyroxenite in composition (Philpotts, 1982). These globules, which occur in mafic and intermediate volcanics, are generally interpreted as immiscible liquid segregates. The fact that they are very minor in abundance, and are of very specific compositions, suggests that they are not parental to large volumes of bimodal volcanics (or granites). Also, there are few if any igneous rocks with the composition the mafic globules.

Fractional Melting of the Mantle

Phase equilibria studies indicate that felsic and mafic liquids can be derived

sequentially by fractional melting of ultramafic rock at water pressures of the order of 20 kb (Yoder, 1973). However, three problems render this mechanism of unlikely importance:

1. In terms of our present understanding of water contents in the mantle, water pressures of 20 kb are not realistic.
2. The felsic liquids produced are not similar in composition to known felsic igneous rocks.
3. Because of the low incompatible element contents in the mantle, exceedingly small degrees of melting ($<3\%$) are required to produce felsic end members.

Liquid Fractionation

Liquid fractionation can produce a zoned magma chamber in which a felsic magma accumulates at the top, and a mafic magma at depth, with a very narrow boundary zone of intermediate magma (McBirney et al., 1985). As crystallization proceeds from the walls of the magma chamber inwards, derivative felsic magmas, due to their low density, rise and collect at the top. The density contrast between the parental mafic magma and the overlying felsic magma prevents significant mixing, and the andesitic boundary zone is therefore narrow and not often tapped during eruptions.

This mechanism may account for bimodal magmas produced in some large stratovolcanoes, and especially in those from continental-margin arcs.

Fractional Crystallization

Bimodal magmas, and especially those in arc systems, may be produced by selective eruption or selective preservation of mafic and felsic end-members. This mechanism is appealing in that many bimodal arc volcanics (especially from oceanic arcs) have similar Sr, Nd, and Pb initial isotopic ratios in mafic and felsic end members. Also, mafic and felsic end-members commonly plot at the ends of calculated continuous fractionation lines. Three mechanisms have been proposed to selectively tap or preserve mafic and felsic end members.

(1) A minimum in density contrast between a magma and surrounding host rock may exist in the andesite compositional range. This would tend to minimize the buoyancy of andesitic magmas relative to neighboring basalt and rhyolite magmas. Hence, andesitic magmas may tend to pond in the crust and undergo fractional crystallization to produce felsic magmas. This mechanism is probably not important in that the country rocks for basalt and andesite are probably the same since most andesites appear to be the products of basalt fractional crystallization.

(2) Basalts, due to their high fluidity, escape to the surface, whereas many andesites pond in the crust and undergo fractional crystallization. When about 50% fractional crystallization is reached, the crystalline residue forms an interconnecting network and is subject to fracturing in response to stresses. Such fracturing allows residual felsic liquids to escape, the net result being an abundance of erupted basalt and felsic volcanics (Thompson, 1972).

(3) A preservational bias may be caused by basalts and rhyolites ending up in back-arc basins while andesites remain near arc volcanoes. If back-arc successions

are preferentially preserved compared to arcs, bimodal volcanics may characterize these successions (see Chapt. 9).

Two-source Models

Most bimodal suites from continental rifts have contrasting isotopic ratios and incompatible element distributions suggesting different sources. In these suites, basalts have mantle isotopic ratios and felsic volcanics (or granites) have isotopic ratios suggestive of crustal origins. These suites can be produced by partial melting of the upper mantle in a rising plume or incipient convective upcurrent that gives rise to the basalts. The basaltic magmas promote partial melting of the lower crust giving rise to felsic liquids. Because of the extensional environment, basalts rise rapidly to the surface before fractional crystallization produces significant quantities of intermediate liquids.

Another type of two-source bimodal magmatism is recorded in Archean greenstone successions and more fully discussed in Chapters 9 and 10. In these successions, dacitic liquids may be produced in garnet-bearing mafic sources possibly in descending oceanic crust, and basaltic liquids in the mantle wedge.

Mantle Sources

The differences in incompatible element contents and isotopic ratios in basalts demands a heterogeneous mantle. However, the nature, scale, preservation, and history of mantle heterogeneities remain problematic. Experimental studies on diffusion rates of cations in mantle minerals provide a basis to constrain the minimum scale of mantle heterogeneities. Results suggest that, even in the presence of a melt where diffusion rates are high, heterogeneities larger than a kilometer in size can persist for several billion years (Zindler and Hart, 1986). Geochemical heterogeneities in ultramafic portions of ophiolites occur on scales of meters to kilometers, suggesting that similar scales exist in the upper mantle. Geochemical and isotopic variations in basalts provide a means to evaluate larger scales of heterogeneity. For instance, isotopic variation within single volcanoes requires heterogeneity on the scale of several kilometers and variations within MORB along ocean ridge segments or along single volcanic arcs demand heterogeneities on scales of 10^2 – 10^3 km. At the large end of the scale is the *Dupal anomaly* between the equator and 60°S latitude (fig. 7.16). This anomaly, which can be defined with both Pb and Sr isotope data from all oceanic basalts, crosses ocean ridges, subduction zones, and continents and appears to extend around the globe. Although it is not clear how the anomaly formed, it is noteworthy that it roughly corresponds to the geoid high and to seismic tomographic anomalies in the mantle (see figs 3.6 and 3.12). Pb isotopic data indicate that the anomaly has existed for billions of years (Hart, 1984).

It has been suggested that linear arrays of data from modern volcanics on $^{87}\text{Sr}/^{86}\text{Sr}$ –Rb/Sr diagrams (and corresponding U/Pb and Sm/Nd diagrams) represent *mantle isochrons* (Brooks et al., 1976). Such isochrons may be interpreted as the times at which major mantle fractionation events occurred. Alternately, they may represent mixing lines and have no age significance. The fact that most arrays of Sr and Pb isotope data define isochron ages in the range of 1.6–1.8 Ga, however, seems to favor an isochron interpretation. Nd arrays give somewhat older ages of ~2 Ga.

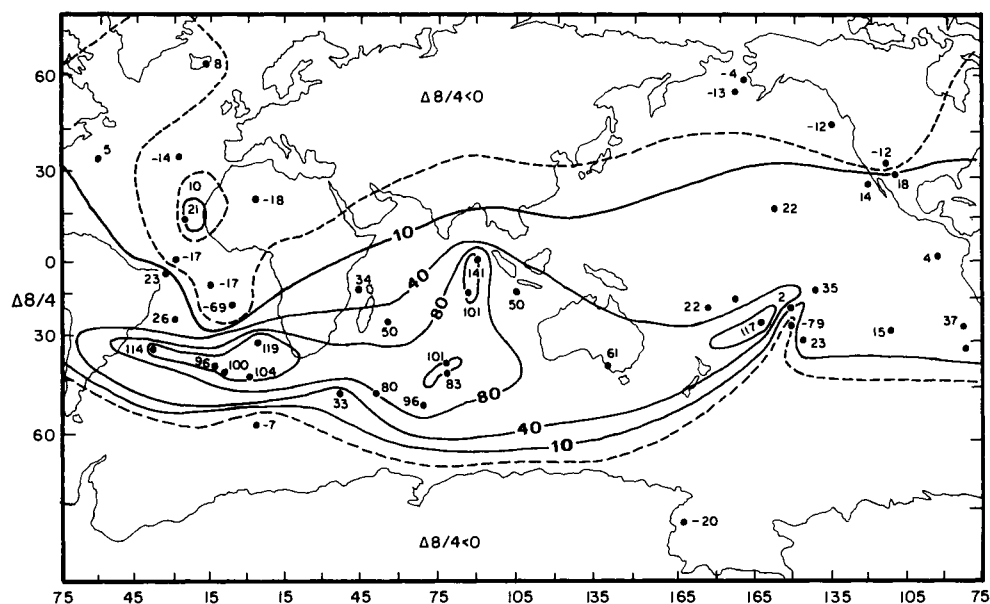


FIG 7.16 The Dupal anomaly. Contoured is the vertical deviation in $^{208}\text{Pb}/^{204}\text{Pb}$ ratio ($\Delta 8/4$) from a Northern Hemisphere reference line on a $^{208}\text{Pb}/^{204}\text{Pb}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ plot. From Hart (1984).

Although a strict isochron interpretation of these data may not be warranted, most geochemists agree they reflect geochemical heterogeneities in the mantle that are $>10^9$ a in age.

Geochemical Mantle Reservoirs

Geochemical mantle reservoirs are regions of the mantle with distinct chemical compositions that have existed for more than a billion years. There are seven radioactive decay schemes with which geochemical mantle reservoirs can be defined: ^{87}Rb - ^{87}Sr , ^{238}U - ^{206}Pb , ^{235}U - ^{207}Pb , ^{232}Th - ^{208}Pb , ^{147}Sm - ^{143}Nd , ^{176}Lu - ^{177}Hf , and $(^{238}\text{U} + ^{235}\text{U} + ^{232}\text{Th})$ - ^4He . Examples of data arrays from oceanic basalts on two binary isotopic diagrams are shown in figures 7.17 and 7.18. It is clear from these plots that models involving only two mantle end-members that mix to form sources for all oceanic basalts are not tenable. This is further documented by other isotopic data arrays (Hofmann and White, 1982; Lupton, 1983; Zindler and Hart, 1986; Allegre et al., 1986). It would appear that a minimum of four and perhaps five or six major mantle reservoirs are required to explain isotopic data arrays. Some possible mantle end-members are briefly considered in the following sections.

Primordial Mantle (PM)

Primordial mantle (PM) is mantle that has not undergone any chemical changes since accretion of the Earth. Whether such mantle has actually survived is a topic of heated debate among geochemists. Two lines of evidence are commonly cited to support the existence of primordial mantle: (1) some basalts have Nd isotopic ratios

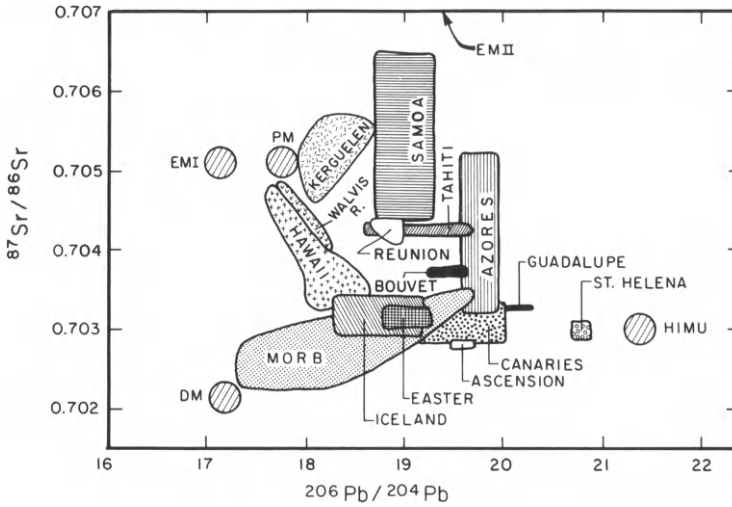


FIG 7.17 $^{87}\text{Sr}/^{86}\text{Sr}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ diagram for oceanic basalts. Modified after Zindler and Hart (1986). Hypothetical mantle reservoirs: DM, depleted mantle; PM, primordial mantle; HIMU, high U/Pb mantle; EMI and EMII, enriched mantle.

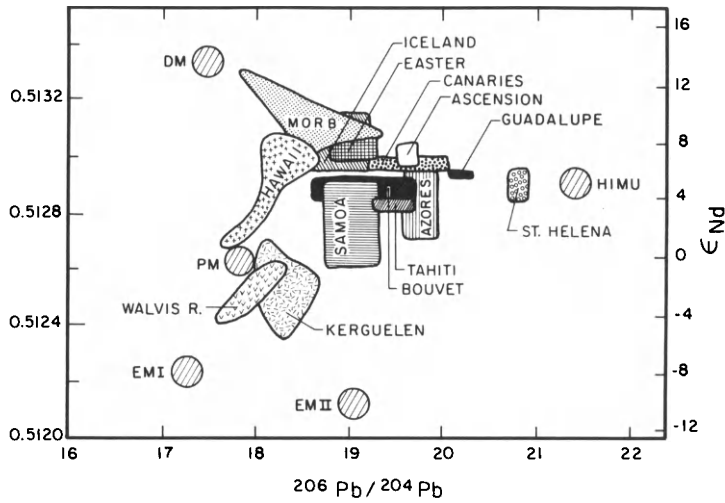


FIG 7.18 $^{143}\text{Nd}/^{144}\text{Nd}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ diagram for oceanic basalts. Modified after Zindler and Hart (1986). Abbreviations given in fig. 7.17.

similar to chondrites, and (2) some basalts have $^3\text{He}/^4\text{He}$ ratios higher than atmospheric values. Since ^3He is not formed by radioactive decay, high ratios of $^3\text{He}/^4\text{He}$ may record primitive mantle. Neither of these lines of evidence is definitive, however. Chondritic Nd isotopic ratios may result from mixing and high $^3\text{He}/^4\text{He}$ ratios may reflect relatively primitive mantle, but not primordial mantle. Also, the high $^3\text{He}/^4\text{He}$ rocks do not have Nd isotopic ratios similar to chondrites. It would seem that isotopic data allow, but do not demand, the survival of primordial mantle.

Depleted Mantle (DM)

Depleted mantle (DM) is mantle that has undergone one or more periods of fractionation involving extraction of basaltic magmas. The depleted isotopic character (low $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$ and high $^{143}\text{Nd}/^{144}\text{Nd}$) and low LILE contents of NMORB require the existence in the Earth of a widespread depleted mantle reservoir. TMORB and EMORB appear to have been derived from sources that are mixtures of depleted and enriched mantle reservoirs. Although most of the geochemical variation within NMORB can be explained by magmatic processes such as fractional crystallization, variations in isotopic ratios demand that the depleted mantle reservoir is heterogeneous, at least on scales of 10^2 – 10^3 km. This heterogeneity may be caused by small amounts of mixing with enriched mantle components. Depleted mantle is known to underlie ocean ridges and probably extends beneath ocean basins, although it is not the source of oceanic island magmas which are known to have feeder systems that extend to depths ≥ 100 km. The depleted isotopic character of mantle xenoliths brought to the surface in continents indicates that depleted mantle also extends beneath continents.

High U/Pb Mantle (HIMU)

The extreme enrichment in ^{206}Pb and ^{208}Pb in some oceanic island basalts (such as those from St Helena, figs 7.17 and 7.18) requires the existence of a mantle source enriched in U+Th relative to Pb on the order of 1.5–2.0 Ga ago, known as *HIMU*. Because this mantle component has $^{87}\text{Sr}/^{86}\text{Sr}$ ratios similar to MORB, however, it has been suggested that it may represent subducted oceanic crust in which the U+Th/Pb ratio was increased by preferential loss of Pb in volatiles escaping upward from descending slabs. Pb also may have been removed by metasomatic fluids from depleted mantle, giving rise to local HIMU reservoirs. Also supporting a mantle reservoir comprised of recycled altered MORB are Ta–Nb distributions in basalts. As previously mentioned, the relative depletion in Ta–Nb in arc basalts appears to reflect the presence of a residual minor phase(s) into which these elements are concentrated in descending slab. Thus, the residual mafic part of the slab should have enriched Ta–Nb contents. The fact that some OIB show enriched Ta–Nb (i.e. positive Ta–Nb anomalies on MORB-normalized graphs such as fig. 7.6) relative to LILE supports the idea that their sources may be residual oceanic crust that has been recycled into the mantle.

Enriched Mantle Components

Enriched mantle components are mantle reservoirs enriched in Rb, Sm, U (and Th) relative to Sr, Nd, and Pb, respectively compared to primordial mantle ratios of these elements. Possible origins for enriched mantle include: (1) metasomatism of the lithosphere, (2) subduction of altered oceanic crust, (3) subduction and mixing of continental sediments in the mantle, and (4) detachment and sinking of subcontinental lithosphere (Hofmann and White, 1982; Menzies, 1983; Hawkesworth et al., 1984). At least two enriched components seem to be required (Zindler and Hart, 1986): EMI with moderate $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and low $^{206}\text{Pb}/^{204}\text{Pb}$ ratios and EMII with high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and moderate $^{206}\text{Pb}/^{204}\text{Pb}$ ratios. Both have low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (figs 7.17 and 7.18). Metasomatized mantle and recycled

lower continental crust are the two chief candidates for EMI. EMII has isotopic ratios similar to average upper continental crust or subducted continental sediments (i.e. $^{87}\text{Sr}/^{86}\text{Sr} > 0.71$ and $^{143}\text{Nd}/^{144}\text{Nd} \approx 0.5121$). Subducted continental sediments are favored for this end-member because EMII commonly contributes to island arc volcanics such as those of the Lesser Antilles and the Banda arc (fig. 7.10). It is notable that there are no known EMI-type arcs. EMII signatures in oceanic island volcanics are chiefly limited to the Southern Hemisphere and serve to delineate the Dupal anomaly (fig. 7.16). Perhaps this anomaly reflects enhanced subduction of sediments from Pangaea during the Late Paleozoic and Early Mesozoic.

Mixing of Mantle Reservoirs

It would appear that a minimum of four mantle reservoirs (DM, HIMU, EMI, EMII) are required to mix to explain the isotopic compositions of oceanic basalts (figs 7.17, 7.18) (Zindler and Hart, 1986). PM is also a permissible end-member. Although consistent mixing relationships exist on Sr, Pb, and Nd isotopic diagrams, diagrams involving He isotopes are more complex. This is probably due to variable degassing of He from the mantle and decoupling of He from LILE during mantle metasomatism.

Summary of Isotopic and Trace Element Constraints

The following constraints for mantle composition seem to be well documented from isotopic and trace element distributions in oceanic basalts.

1. NMORB is derived from a depleted source which is located in the upper mantle and contains small-scale heterogeneities.
2. Depleted mantle may grade downward into mantle that is less depleted on the average, or it may be in contact with less depleted mantle perhaps at the 670-km discontinuity.
3. Primordial mantle, if it exists, cannot comprise more than half of the mantle.
4. The Dupal anomaly in the Southern Hemisphere documents the existence of large-scale heterogeneities in the upper mantle.
5. Enriched mantle sources and the HIMU source are produced by either magmatic or metasomatic processes in the mantle or/and by recycling of oceanic crust into the mantle at subduction zones.
6. Enriched and HIMU sources have existed in the mantle for billions of years.

Although numerous models for origin and survival mantle heterogeneities have been proposed, most results can be interpreted within the framework of one of two models: the layered mantle model and the plum pudding model. Each will be briefly reviewed.

The Layered Mantle Model

The *layered mantle model* involves separately convecting upper and lower mantle reservoirs. The upper mantle, above the 670-km discontinuity, is generally equated

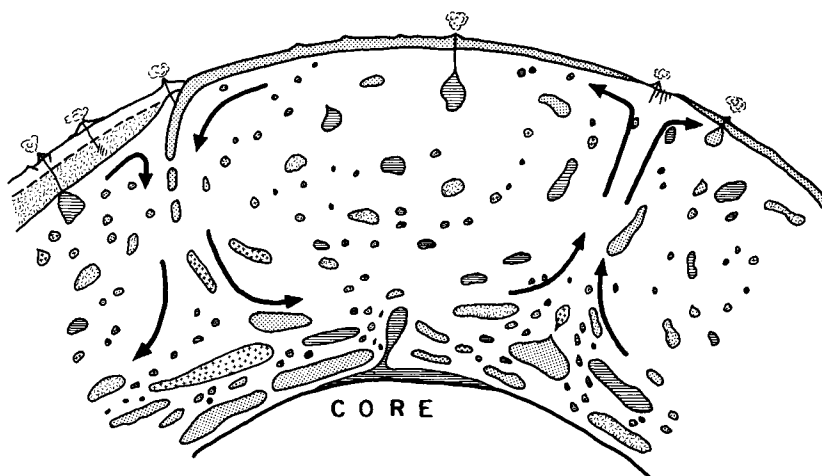


FIG 7.19 Hypothetical drawing of the plum pudding model for the mantle. The blobs (or plums) represent various enriched mantle components in a depleted mantle matrix (NMORB source). Modified after Davies (1984).

with the depleted mantle reservoir and the lower mantle with either primordial mantle or less depleted mantle (Allegre, 1982; O'Nions, 1987). The depleted upper mantle is assumed to have formed chiefly by the extraction of continental crust from primordial mantle. Model calculations employing Nd and Sr isotopes indicate that such a depleted mantle reservoir must comprise between 30 and 50% of the entire mantle.

One of the major lines of evidence used to support a layered mantle comes from the isotopic composition of rare gases in mantle-derived volcanics (Lupton, 1983; O'Nions, 1987). The very high $^3\text{He}/^4\text{He}$ ratio in MORB indicates that primordial ^3He escapes from the Earth today. ^3He is even more enriched in many OIB. Since many or most OIB appear to be hotspot-related and thus derived from mantle plumes, they probably come from a mantle source deeper than the depleted MORB source. Similar arguments can be made from Ar isotopes. The simplest model to satisfy these constraints is a two-layer convective mantle with the upper layer strongly depleted in rare gases and LILE and the lower layer, which is relatively undepleted, is the site of generation of mantle plumes (relatively rich in ^3He) that rise and serve as sources for oceanic island volcanics.

Some theoretical and laboratory studies of convection are also consistent with layered convection in the Earth. As reviewed in Chapter 3, however, there are significant geophysical problems with layered convection models and most geophysical data seem to favor whole-mantle convection.

The Plum Pudding Model

The *plum pudding model* for the mantle suggests that it is a mixture of variably depleted mantle (the pudding) with embedded enriched sources (the plums) (fig. 7.19). The plum pudding model is consistent with whole-mantle convection and accommodates the geophysical constraints discussed in Chapter. 3. Heterogeneities

are generally envisioned as having many sizes and degrees of enrichment and variable, often elongate shapes (Morris and Hart, 1983; Davies, 1984). A moderate increase in viscosity may occur with depth which slows the convective flow and mixing rates in the lower mantle. Heterogeneities enter at subduction zones as descending oceanic crust, subducted sediments, or detached subcontinental lithosphere (fig. 7.19). Other heterogeneities may be produced by metasomatism, perhaps at the base of the lithosphere in response to rising mantle plumes. Plumes may rise from the hot D'' region at the base of mantle (see Chapt. 5). Small degrees of melting of plumes give rise to OIB, whereas large degrees of melting beneath ocean ridges dilute the effect of enriched plumes such that MORB chiefly reflect the depleted mantle source (the pudding). CRB may be formed by partial melting of enriched mantle plumes or by partial melting of enriched subcontinental lithosphere.

Summary Statements

1. Basalts erupted at ocean ridges (MORB) are chiefly tholeiites exhibiting varying degrees of depletion in LILE and low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The degree of low-temperature alteration of MORB, which results in addition of LILE, increases with increasing distance from ocean ridges.
2. Within-plate magmas are produced at oceanic islands and continental rifts, and to a minor extent beneath cratons. They belong chiefly to the tholeiite or alkaline series and may be bimodal. Kimberlites, which occur in cratons, contain mantle xenoliths and may contain diamonds and appear to have been produced at depths of 200–300 km. Compared to MORB or chondrites, within-plate magmas are enriched in LILE, REE, and HFSE.
3. Igneous rocks produced in subduction zones belong to the tholeiite or calc-alkaline series and exhibit a subduction zone component (SZC) (i.e. depletion in Ta–Nb relative to LILE).
4. Some arc volcanics (and granites) may show increases in LILE content with depth to subduction zone (compositional polarity) and most arc volcanics vary in composition along the strike of the arc.
5. Igneous rocks produced in collisional orogens are chiefly calc-alkaline granites (leucogranites) with a SZC and are produced by partial melting of sediments or other crustal rocks.
6. Granites are classified into four major categories: (1) M-type, plagiogranites from ophiolites which are depleted in LILE and appear to be derived by fractional crystallization of MORB; (2) I-type, granitic batholiths from arcs or collisional orogens produced chiefly by fractional crystallization of CAB-IAB; (3) S-type, granites that are peraluminous and may contain cordierite and appear to be the products of partial melting of sediments in collisional orogens; and (4) A-type post-tectonic granites highly enriched in LILE, REE, and HFSE, that may be produced by partial melting of the lower crust.
7. MORB are produced by 15–30% partial melting of a variably depleted ultramafic source at depths of 50–85 km beneath ocean ridges.
8. Within-plate basalts are probably produced by partial melting of relatively enriched mantle in rising plumes or/and in enriched lithosphere.

9. Subduction-related basalts are probably produced by partial melting of enriched ultramafic rocks in the mantle wedge. Although fractional crystallization plays the dominant role in the evolution of subduction-related magmas, each of the following may also contribute to these magmas: subducted sediments, contamination by continental crust, partial melting of continent, and magma mixing.
10. A subduction zone component (SZC) may be produced by devolatilization of the descending slab in which LILE are enriched in the mantle wedge (which later becomes the source of CAB-IAB) while Ta-Nb are retained in residual minor phases in the slab. Because slab devolatilization is completed before the volcanic front, SZC-mantle must be dragged to greater depths by convection or frictional drag along the descending slab before melting can occur.
11. Bimodal magmatism may be caused by liquid fractionation producing a zoned magma chamber; fractional crystallization in which andesites are not commonly erupted or in which a preservational bias leads to an abundance of felsic-mafic components in arc-related basins; or by two different magma sources (mantle and crust).
12. Geochemical and isotopic data indicate the mantle is heterogeneous on scales ranging from a few meters to $>10^4$ km, and that many of these heterogeneities have ages of several billion years.
13. A minimum of four and perhaps six or more end-member mantle reservoirs are required to explain incompatible element and isotope ratios in oceanic basalts. Possible reservoirs include: (1) primordial mantle, mantle which has not undergone chemical change since the Earth's accretion; (2) depleted mantle which has undergone one or more periods of magma extraction and is the source of MORB; (3) HIMU mantle, with a high U-Th/Pb ratio and is a source of some oceanic island volcanics; (4) various enriched mantle components that contribute to sources of within-plate basalts.
14. Enriched mantle sources may include (1) metasomatized lithosphere, (2) subducted oceanic crust, which may be the source of HIMU mantle, (3) mantle with mixed subducted sediments, and (4) detached subcontinental lithosphere.
15. Two competing models are proposed for the origin and distribution of mantle heterogeneities: the layered mantle model in which the upper (depleted) mantle and lower mantle convect separately, and the plum pudding model in which enriched heterogeneities reside in a variably depleted matrix. The plum pudding model is consistent with geophysical data supporting whole-mantle convection.

Suggestions for Further Reading

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CHAPTER 8

Phanerozoic Orogenic Systems

Introduction

General Features

Major Phanerozoic orogenic systems include the Appalachian–Caledonian in eastern North America and western Europe, the Hercynian and Uralian in Europe, the Samfrau in Gondwana, and the Cordilleran extending from western Europe to southeast Asia (Plate I). These and other Phanerozoic orogenic belts have formed in response to complex plate interactions at convergent plate boundaries chiefly involving collisional orogenies. Before outlining the plate histories of portions of these orogenic systems, it is instructive to summarize continental fragmentation and assembly during the last 600 Ma.

Although the positioning of continents is based chiefly on paleomagnetic data, paleontologic and paleoclimatic results and the distribution of petrotectonic assemblages provide important constraints on reconstructions (Irving, 1979b; Ziegler et al., 1979). The largest supercontinent during the Phanerozoic is *Pangaea*, which includes most of the existing continents. *Gondwana* is comprised principally of South America, Africa, Arabia, Madagascar, India, Antarctica and Australia. Portions of southwestern Eurasia and southern Europe were also part of Gondwana during much of the Paleozoic. *Laurasia* consists of North America, Europe, and most of Asia; and *Laurentia* includes most of North America, Scotland, and Ireland north of the Caledonian suture, Greenland, Spitzbergen, and the Chukotsk Peninsula of eastern Siberia. Asia north of the Himalayas and east of the Urals consists of numerous microcontinents, the best-defined of which are Siberia, Kazakhstan, North China, South China and SE Asia. Siberia is bordered on the west by the Uralian suture and on the southwest by the Irtysh crush zone.

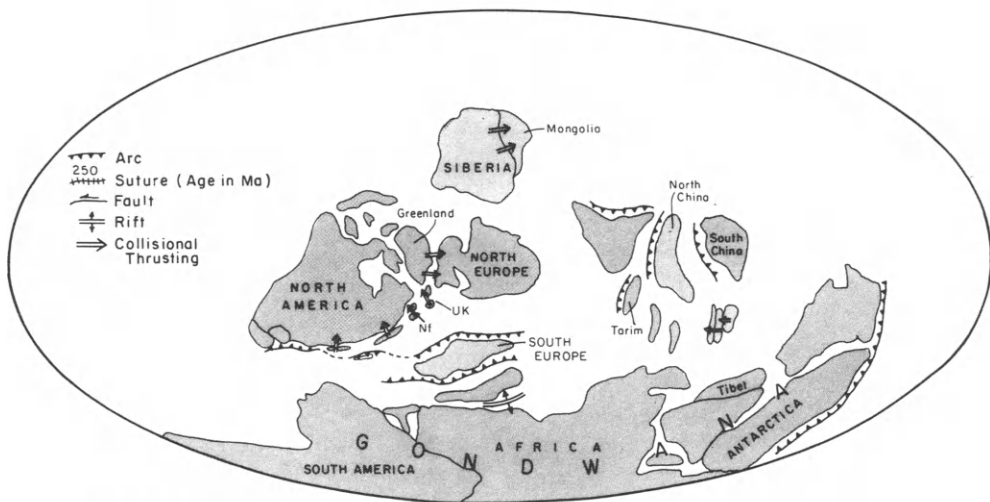
The Paleozoic

As discussed in Chapter 9, it is possible that a large supercontinent existed in the late Precambrian (fig. 9.28) and that, by no later than 750 Ma, this supercontinent began to fragment. New passive margins were created along the edges of North America, Siberia, and North and South Europe. In addition, an unknown number of microcontinents may have been rifted from Gondwana at this time. Paleomagnetic and paleoclimatic data indicate that during the Early Paleozoic most continents



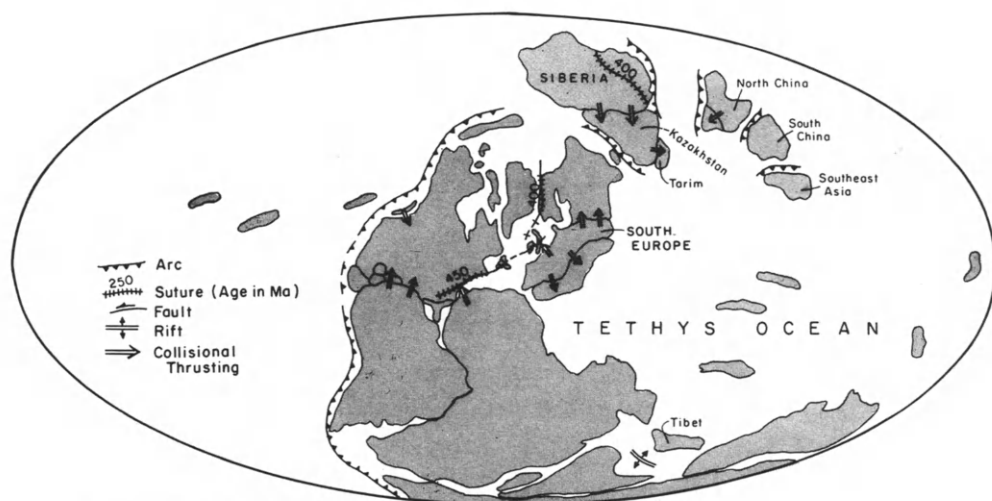
a. LATE CAMBRIAN

FIG 8.1 Phanerozoic continental reconstructions. In part after Scotese et al. (1979, 1981) and Scotese (1984). Nf, Newfoundland; UK, United Kingdom.

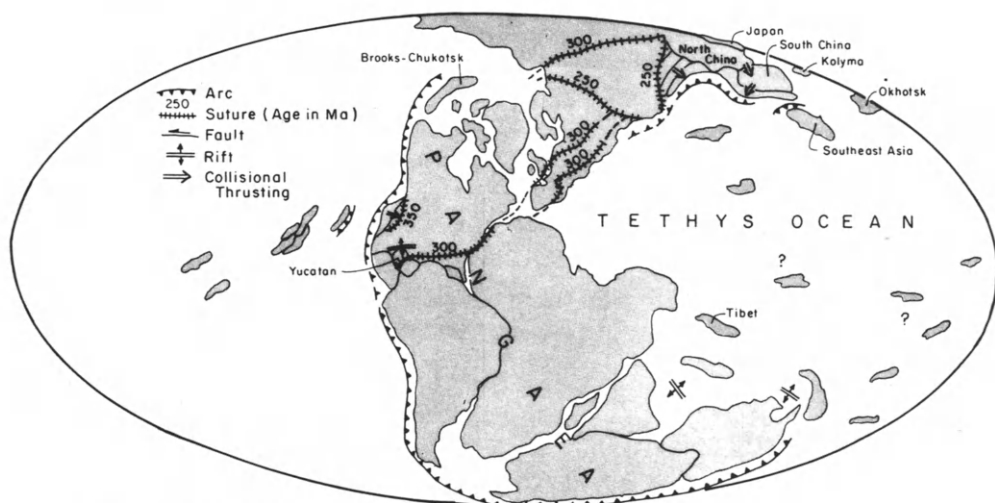


b. MIDDLE SILURIAN

remained at low, equatorial latitudes (Scotese et al., 1979; Scotese, 1984) (fig. 8.1a). Gondwana, however, also extended into southern polar latitudes and was glaciated during the Ordovician. Extensive carbonate platforms in North America, Siberia, and North and South China support paleomagnetic results that indicate near-equatorial positions in the Early Paleozoic. During the Cambrian and Ordovician, subduction zones existed along the south coast of Laurentia, the west coast of North Europe, and along the coast of several microcontinents (Avalonia, Cadomia) west of North Europe (fig. 8.1a). A major ocean, the *Iapetus Ocean*, existed between these subduction systems. Note that Florida was part of Gondwana at this time and that

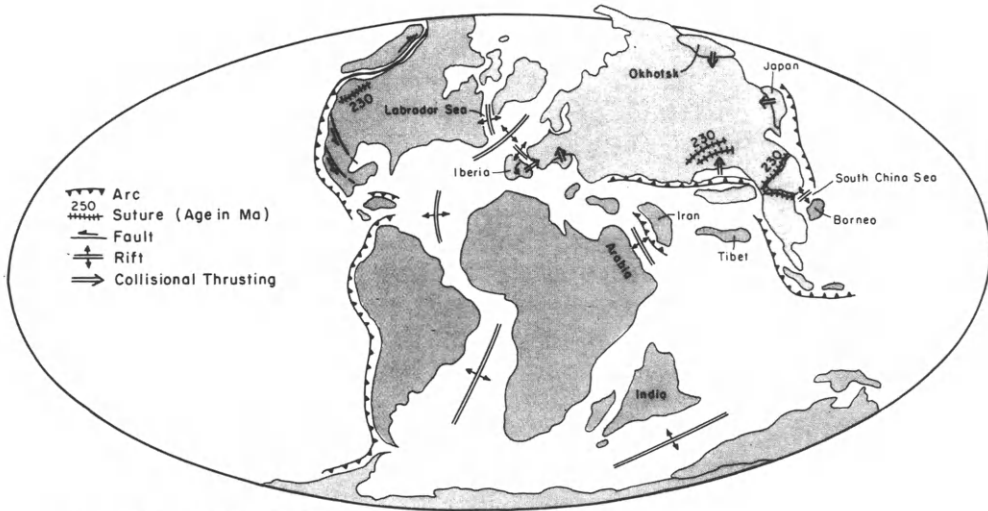


c. LATE CARBONIFEROUS

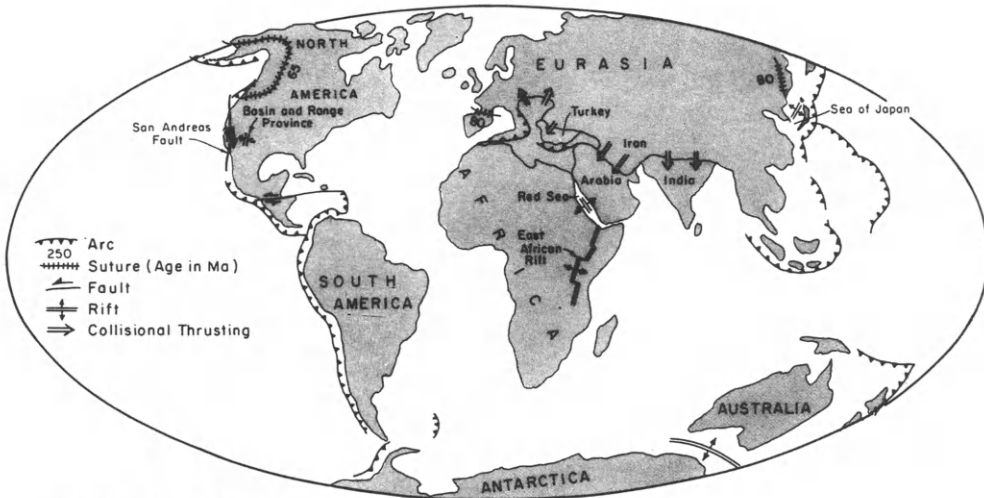


d. MID-TRIASSIC

the United Kingdom was not yet united: i.e. Cadomia (England, Wales) was on one side of the Iapetus Ocean and Scotland on the other. Newfoundland and parts of New England and NE Canada were also separated by this ocean. The Chukotsk Peninsula in extreme NE Siberia was connected to northern Alaska (Brooks-Chukotsk) and this landmass was part of Laurentia. Mongolia, which had been rifted from Siberia in the late Precambrian, had a marginal subduction zone and was closing on Siberia in the earliest Paleozoic. Other subduction systems existed along both coasts of South Europe, along western Kazakhstan and western Tarim, and a very long subduction system, the *Samfrau arc* extended from eastern Australia to southern America (fig. 8.1a). Paleomagnetic data suggest that North and South China were separated by Kazakhstan (Lin et al., 1985). As evidenced by the number



e. LATE CRETACEOUS



f. MID-TERTIARY

of Paleozoic accretionary terranes in central and southern Asia (see fig. 8.18), other arcs and microcontinents also probably existed during the early Early Paleozoic in the region between Kazakhstan and Gondwana.

During the Ordovician and Silurian the Iapetus Ocean began to close and Avalonia, Cadomia, and North Europe collided with Laurentia producing a succession of collisional orogenies recorded in the Appalachians in North America and the Caledonides in Scandinavia (fig. 8.1b). During the Ordovician and Silurian, Siberia moved toward Laurentia, Gondwana was centered approximately over the South Pole and shallow seas covered large portions of the continents. Also probably during the Silurian, Mongolia collided with Siberia and several microcontinents collided south of South China. Collision continued along the east coast of Laurentia

in the Devonian, Carboniferous, and into the Permian when Africa collided with southeastern North America. Beginning in the Early Carboniferous and ending in the Late Carboniferous, the *Hercynian orogen* developed by collision along both margins of South Europe (fig. 8.1c). In the Late Carboniferous and Permian, South America collided with North America causing the *Ouachita orogeny*, and the Roberts Mountains allochthon collided with western North America. Siberia and Tarim collided with Kazakhstan, and North China collided with a small microcontinent in the Late Carboniferous. In addition, Tibet may have been rifted from Gondwana. During the Permian, Kazakhstan collided with North Europe producing the *Uralian orogeny*, and North China and several other small plates collided with Siberia. By the Permian, Pangaea was nearly complete with the *Tethys Ocean* occupying a large re-entrant on the east (fig. 8.1c). Throughout the Paleozoic the Samfrau subduction system was active, propagating with time along the southwest coast of South America and joining with a subduction system in western North America. Thus, a major subduction system surrounded Pangaea along its western and southern margins in the Late Paleozoic. During the Paleozoic, Australia grew eastward forming the *Lachland* and *New England orogens*, largely by the addition of accretionary prisms and closure of back-arc basins. Antarctica and the southern tip of South America also grew by similar processes along the Samfrau arc system.

Plate motions during the Paleozoic were characterized by continual continental convergence and collision, the net result of which was the aggregation of Pangaea, which extended from pole to pole by the Permian.

The Mesozoic

In contrast to the Paleozoic the Mesozoic and Cenozoic are times of continental fragmentation leading to the present distribution of continents. During the Triassic, the amalgamation of much of China was largely completed, and the Golconda allochthon was accreted to the western United States (fig. 8.1d). In the Late Triassic the Gulf of Mexico began to open as Yucatan and other microplates were rifted away from Texas. An unknown number of microcontinents and arcs existed in the Tethys basin, many of which were rifted from Gondwana. These continued to collide with western North America and southern and eastern Asia during most of the Mesozoic. During the Jurassic and Cretaceous, several terranes were accreted to western, southeastern, and eastern China and several arc systems were accreted in the northwestern United States. Kolyma and Okhotsk collided with NE Siberia in the Early Jurassic in the Cretaceous, respectively, thus completing most of Siberia. Continued collisions in the western Tethys led to formation of a microcontinent that collided with northwestern North America in the Late Cretaceous and Early Tertiary. Brooks–Chukotsk rotated counterclockwise beginning in the Early Cretaceous, and collided with Siberia and Alaska in the Mid-Cretaceous.

In the Middle and Late Jurassic, North America and Africa were rifted apart and the North Atlantic began to open. Major opening of the North Atlantic, however, did not occur until the Cretaceous (fig. 8.1e). Also, during the Middle to Late Jurassic, Africa began to rift away from India–Antarctica. In the Mid-Cretaceous the South Atlantic began to open as Africa was rifted away from South America, the western Arctic basin began to open, and the Bay of Biscay opened as Iberia rotated

counterclockwise. During the Late Cretaceous the Labrador Sea began to open as Greenland moved away from North America, and Africa continued its counterclockwise motion closing the Tethys Ocean. Also during the Late Cretaceous, microplates rifted from North Africa and began to collide with Eurasia, and the South Atlantic became a wide ocean, Madagascar reached its present position relative to East Africa, and India began rapid motion northwards toward Asia. Deformation began in the Alps and Carpathians in response to microplate collisions and Iran was rifted from Arabia (fig. 8.1e). During the Jurassic and Cretaceous the Gulf of Mexico continued to open, the North American subduction zone propagated southward and a new subduction zone formed along the margin of Cuba. Major deformation occurred in response to collisions and subduction along the Cordilleran orogen in western North and South America and Antarctica. Between 60 and 80 Ma ago New Zealand was rifted from Australia as the Tasman Sea opened. About 100 Ma ago, India was rifted from Gondwana and Borneo was rifted from China as the South China Sea opened.

The Cenozoic

During the Early Tertiary, India collided with Tibet, promoting the *Himalayan orogeny* (fig 8.1f), and continued convergence of microplates in the Mediterranean area caused the *Alpine orogeny*. In addition, the North Atlantic between Greenland and Norway began to open and the Sea of Japan opened. During the Mid-Tertiary, Australia was rifted from Antarctica and moved northward, continued convergence in southern Europe led to the formation of large nappes in the Alps and Carpathians, the Rhine graben formed, and at about 30 Ma the Red Sea began to open as Arabia was rifted from Africa. Collision of the East Pacific rise with western North America led to development of the San Andreas transform fault, and Iceland formed on the Mid-Atlantic ridge.

In the Late Tertiary, and continuing to the present, oceanic arc systems formed in the South Pacific, subduction migrated into the eastern Caribbean and the Lesser Antilles arc formed, and southward propagation of the Middle America subduction zone resulted in the growth of Panama connecting North and South America. Arabia collided with Iran in the Late Tertiary and was sutured along the Zagros crush zone. The Basin and Range Province and the Rio Grande rift formed in North America, the East African rift system developed in Africa, and the Gulf of Aden completely opened. Finally, at about 3–4 Ma ago, Baja California was decoupled from the North American plate along the San Andreas transform fault and the Gulf of California opened.

The Appalachian–Caledonian System

Major Paleozoic orogenies are recorded along both coasts of the North Atlantic. The *Caledonian orogeny*, which occurred in the Upper Silurian and Lower Devonian, is widespread in the British Isles, Scandinavia, and eastern Greenland. The *Taconic orogeny* (Middle Ordovician), *Acadian orogeny* (Middle Devonian), and *Alleghanian orogeny* (Late Carboniferous and Permian) are prominent in the Appalachian Mountains in eastern North America. These orogenies can be related to a complex

history of the North Atlantic region involving opening, closing, and reopening of the Iapetus Ocean basin. This basin began to form in the Late Precambrian when North America and Africa were rifted apart, and during this time, thick sections of cratonic sediments were deposited on both retreating continental margins. Before the beginning of the Cambrian, a convergent plate margin developed along the African coast and during the Cambrian, along the American coast. The subduction zones may have been composite and, along at least portions of the Atlantic margin, probably dipped oceanward.

Orogeny began in North America and Scandinavia in the Ordovician. It is recorded in the Early and Middle Ordovician along the east coast of North America by westward-directed thrusts and folds and syntectonic metamorphism and plutonism. In Scandinavia, oceanic crust was obducted onto the Baltic Shield in the Late Ordovician. During the Late Silurian and Devonian, Africa and Europe collided with the northeastern coast of North America. Associated thrusting, metamorphism, and magmatism record the Acadian and Caledonian orogenies. The Caledonian orogeny in Scandinavia involved collision of Greenland and Norway in the Late Ordovician and Silurian, which resulted in eastward advancement of thrusts and nappes derived from Greenland over the Baltic Shield. Obducted ophiolites in Newfoundland are interpreted as remnants of the Iapetus seafloor. Compressive forces continued into the Late Pennsylvanian and Permian in the central and southern Appalachians (and in the Ouachita area), producing large continent-directed thrust sheets and nappes (the Alleghanian orogeny). This deformation records the final collision of Africa with the southeastern coast of North America. Africa is welded to North America along a suture zone which is now buried by younger sediments in the Atlantic coastal plain. A complex history of Paleozoic deformation and magmatism paralleling that recorded in eastern North America is also recorded in northwest Africa, although it is not as well understood. The opening of the Atlantic Ocean began in the Triassic and is continuing today with new sediment assemblages forming on the trailing edges of North America and Africa. Although this simplistic evolutionary scheme describes the overall history of the Appalachian-Caledonian belt, the detailed history at any locality is very complex, as the following examples illustrate.

The Southern Appalachians

The southern Appalachian Mountains in the southeastern United States can be divided into several tectonic zones as shown in fig. 8.2. The Valley and Ridge province includes folded and faulted Paleozoic cratonic sediments. The Blue Ridge province is comprised of Precambrian igneous and metamorphic rocks that were transported westward with the Inner Piedmont and Charlotte and Carolina belts along a decollement thrust. A COCORP seismic-reflection profile in approximately the same location as fig. 8.2 indicates that the decollement is 6–15 km thick (Cook et al., 1979), and geologic data indicate that it probably overlies Paleozoic cratonic sediment. The Brevard zone is interpreted as a subsidiary thrust emanating from the main thrust. The Inner Piedmont province consists of Precambrian metasediments and metavolcanics intruded with a variety of plutonic rocks, the King Mountain belt is a synclinal zone, and the Charlotte belt is a region of high-grade gneisses of

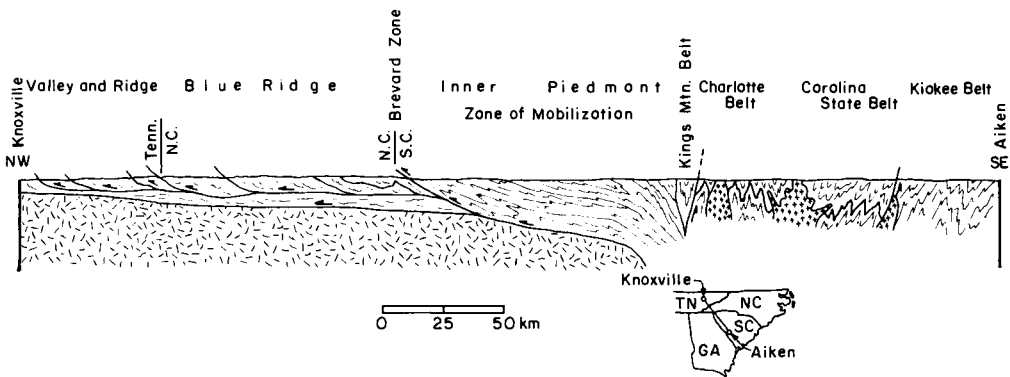


FIG 8.2 Diagrammatic cross-section of the southern Appalachians. Modified after Hatcher (1981).

Precambrian and Paleozoic age. The Carolina Slate and Kiokee belts are comprised of deformed and metamorphosed Early Paleozoic volcanics and volcanoclastic sediments. Precambrian basement is unconformably covered with Mesozoic rift and cratonic sediments in the Coastal Plain.

A model for the evolution of the southern Appalachians consistent with geological and geophysical data is presented in fig. 8.3. During the Late Precambrian, rifting of Laurasia produced at least three small ocean basins with two continental fragments, the eastern Blue Ridge–Piedmont (EBR–IP) and the Charlotte–Caroline–Kiokee belts (CCK) (a,b). The metavolcanics and metasediments of the western Blue Ridge province were deposited in the ocean basin between North America and the EBR–IP block. Subduction and related volcanism began beneath the CCK province by 650 Ma ago (c) and along the western margin of the EBR–IP province by 500 Ma ago (d). The small ocean basin between the EBR block and North America closed, and cratonic sediments derived from America began to accumulate by 500 Ma ago. This closure between 450 and 500 Ma produced the Taconic orogeny as the Blue Ridge and Piedmont provinces were thrust westward over the North American craton (e). After the closure, sediments were derived primarily from the east. The Acadian orogeny, between 350 and 400 Ma, is characterized by widespread deformation, metamorphism, and plutonism, and appears to have been triggered by the closing of the ocean basin between the IP and CCK blocks. The Kings Mountain belt may be a surface remnant of the suture between the Inner Piedmont and CCK belt. After the Acadian orogeny a new convergent boundary developed beneath the CCK belt dipping to the west (f). The Alleghanian orogeny (250–300 Ma) resulted from the final closing of the Iapetus Ocean and the collision of Africa with America (g). This collision produced extensive plutonism and large-scale overthrusting directed toward both the American and the African plates. The African plate was fragmented from the North American plate in the Triassic as the Atlantic basin began to open. The line of rifting is east of the Carolina Slate belt, buried today by Atlantic-Coast cratonic sediments.

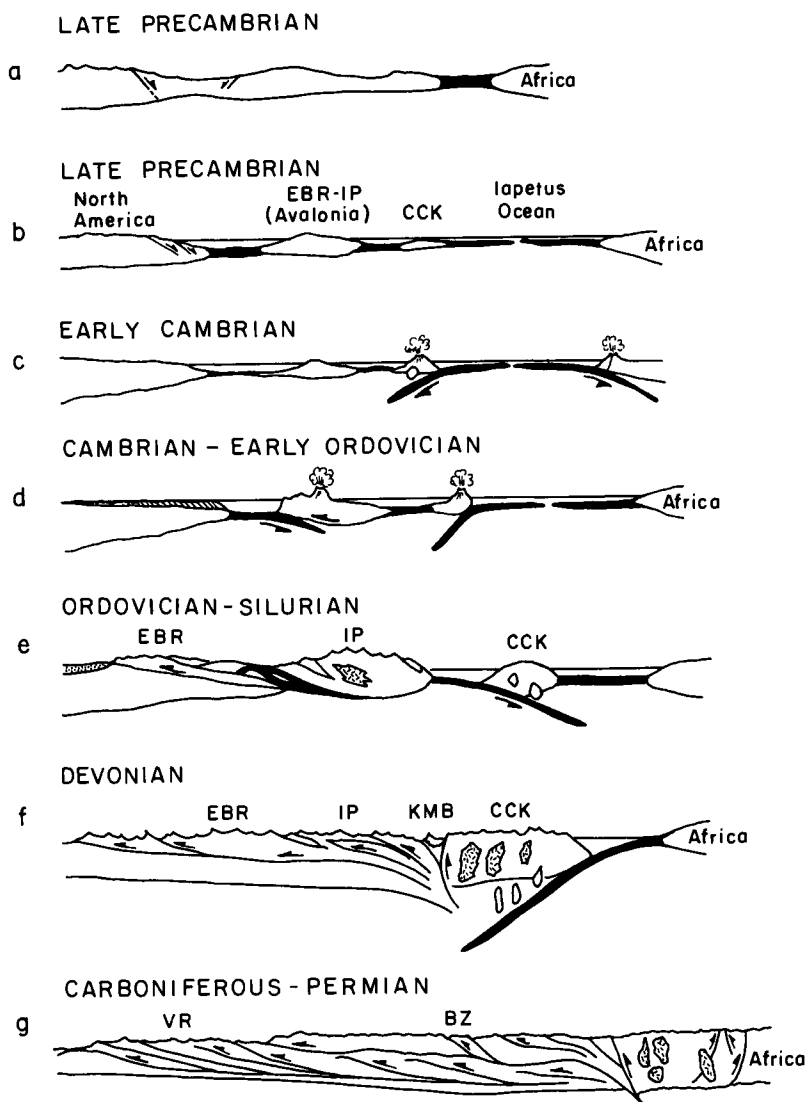


FIG 8.3 Plate tectonic history of the southern Appalachians. Modified after Hatcher and Odom (1980). EBR, eastern Blue Ridge; IP, Inner Piedmont; CCK, Charlotte-Carolina-Kiokee belts; KMB, Kings Mountain belt; VR, Valley and Ridge; BZ, Brevard zone.

Newfoundland

Fragments of both eastern and western margins of the Iapetus Ocean are preserved in Newfoundland. The succession of rocks from the edge of the Avalon platform northwest into the Central Paleozoic mobile belt is interpreted as the southeast margin of the Iapetus Ocean (Colman-Sadd, 1980). An evolutionary model for this area is summarized in fig. 8.4. In the Late Precambrian a subduction zone dipped eastwards beneath the Avalon platform, with the Iapetus Ocean on the west (a). The protoliths of the Little Passage gneisses appear to represent an accretionary prism.

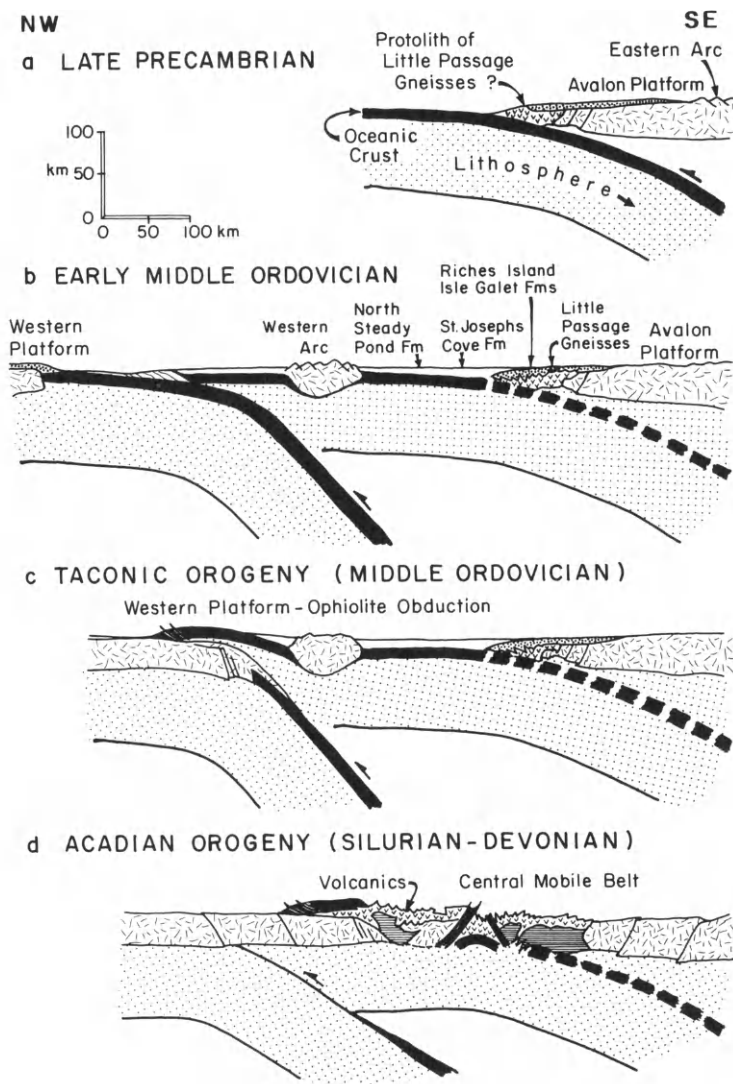


FIG 8.4 Evolutionary model for the eastern margin of the Iapetus Ocean as recorded in Newfoundland. After Colman-Sadd (1980).

By the Mid Ordovician a new subduction zone developed with an extensive arc and back-arc basin system (b). Sediments of the North Steady Pond, St Joseph's Cove, Riches Island, and Isle of Galet Formations were deposited in these back-arc basins, and include mixed arc and pelagic sediments. The eastern arc also must have been active, in that sediments are found in the Isle Galet Formation. The closing of the Iapetus Ocean began in the Mid-Ordovician with the obduction of ophiolites over the Western Platform (now in northwestern Newfoundland) (c). Ophiolite obduction was separated from major Taconic deformation and metamorphism by about 50 Ma. Collision of the Western and Avalon platforms did not occur until Late Silurian–Early Devonian, and is recorded by extensive thrusting, folding, and

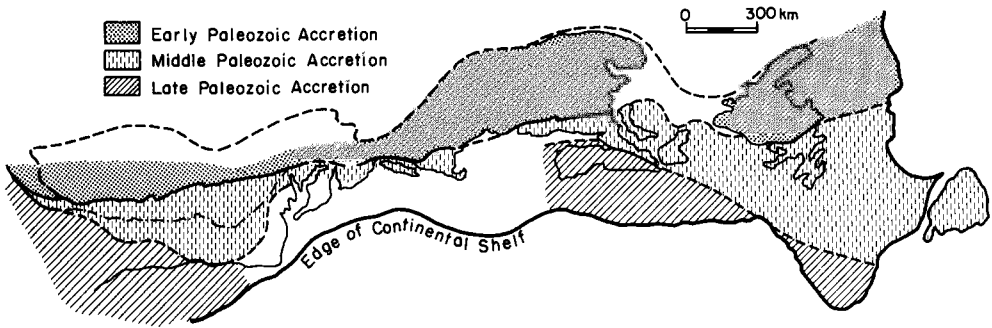


FIG 8.5 Paleozoic accretionary belts in the Appalachian orogen. After Williams and Hatcher (1983).

volcanism associated with the Acadian–Caledonian orogeny (d). During this time the crust was thickened and the lower portion partially melted, giving rise to granitic plutons in the Central mobile belt.

Accretion in the Appalachian Orogen

Accretion in the Appalachian orogen progressed from the North American craton outwards (Williams and Hatcher, 1983) (fig. 8.5). The boundaries of earlier accreted terranes are marked by melange zones and obducted ophiolites, implying head-on collisions. The more eastern terranes have as boundaries steep mylonite zones and faults, suggesting oblique collisions. Times of major deformation coincide with the main phases of accretion in the orogen. The effects of the Taconic orogeny are restricted to the first accreted western terranes, and the Acadian orogeny affected the central and some eastern terranes as well as the earlier Taconic deformed zone. The effects of the Alleghanian orogeny are widespread in the southern part of the orogen but its most intense effects are in the Avalon terrane in Nova Scotia. The major periods of plutonism and metamorphism correspond with the periods of deformation and, in general, also decrease in age from west to east.

The Hercynian System

The Hercynian orogenic system extends from eastern Europe to the southern coast of the United States (Ouachita belt) and developed in response to the collision of South Europe with North Europe and Africa and the collision of Africa–South America with North America. In Europe the Hercynian orogen is an irregular belt approximately 6000 km long with a central metamorphic core and two bordering fold belts (fig. 8.6). The central core is composed of variably metamorphosed Precambrian and Early Paleozoic rocks intruded with syn- to post-tectonic granitic plutons. In the marginal fold belts the age of folding and thrusting decreases with increasing distance from the core zone. These fold belts are generally thought to have developed in association with two marginal subduction zones that migrated outwards from the core (Lorenz and Nicholls, 1984). In this respect, South Europe would have been an elongated continent (500–1000 km wide) with two marginal arc systems and oceans on both sides. The fact that folds and thrusts face outwards on both margins of the microcontinent suggests that subduction zones dipped inward.

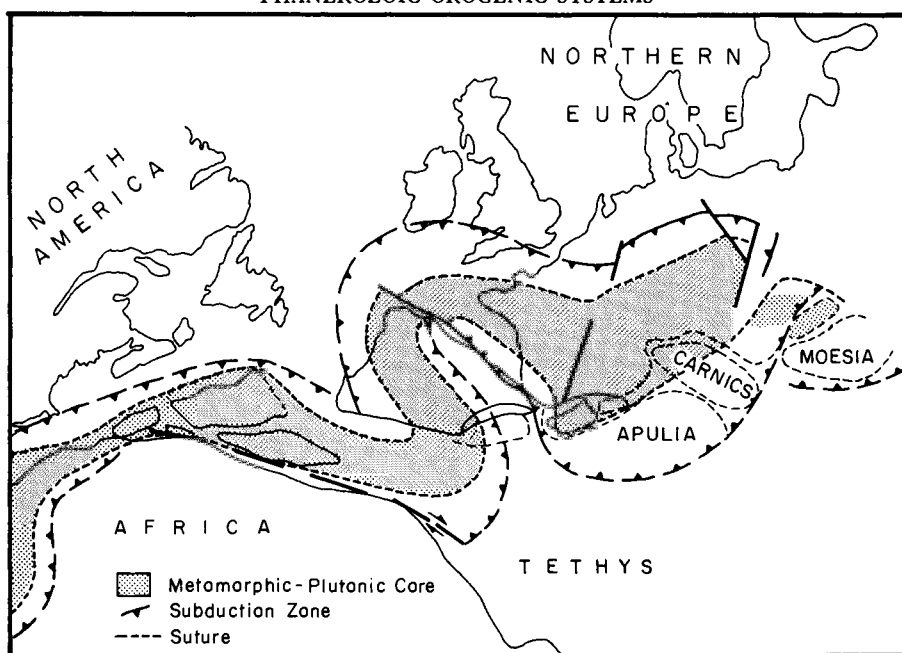


FIG 8.6 Schematic map of the Hercynian orogen. Modified after Lorenz and Nicholls (1984).

The marginal orogenic belts appear to have developed in response to collisions of South Europe with North–America–Europe on the north, and with African-derived microcontinents on the south. Collision began on both coastlines in the Early Carboniferous and suturing along both borders was largely completed by the Late Carboniferous for a total orogenic timespan of about 30 Ma. The very irregular outline of the suture zones probably reflects, in part, irregular continental coastlines prior to collision. Paleomagnetic studies indicate differential movements of various blocks within the Hercynian belt, and such movements are a necessary consequence of convergence of continents with irregular margins. Plutonism and volcanism were both active during the Hercynian collisions and also persisted for 10–20 Ma afterward.

The Urals

The Uralian orogen is a linear Paleozoic fold belt between western Europe and Kazakhstan (Plate I) that averages 400–450 km wide and extends for thousands of kilometers. It occurs between two largely undeformed plates and the central zone is composed chiefly of deformed oceanic and island arc assemblages. The marginal zones of the orogen are composed of deformed cratonic sediments upon which thrusts and nappes from the central zone have advanced.

The Uralian orogen developed in response to the collision of the North European and Kazakhstan plates in the Permian. A summary of the major features of this collision is given in figure 8.7. Beginning at about 500 Ma a fragment of the European plate (Mugodjaria) was rifted off and the Uralian Ocean opened (a). Between 400 and 500 Ma a subduction zone developed in the Uralian oceanic crust, and the Sakmarian arc formed as the Uralian ocean closed (b). At about 500 Ma

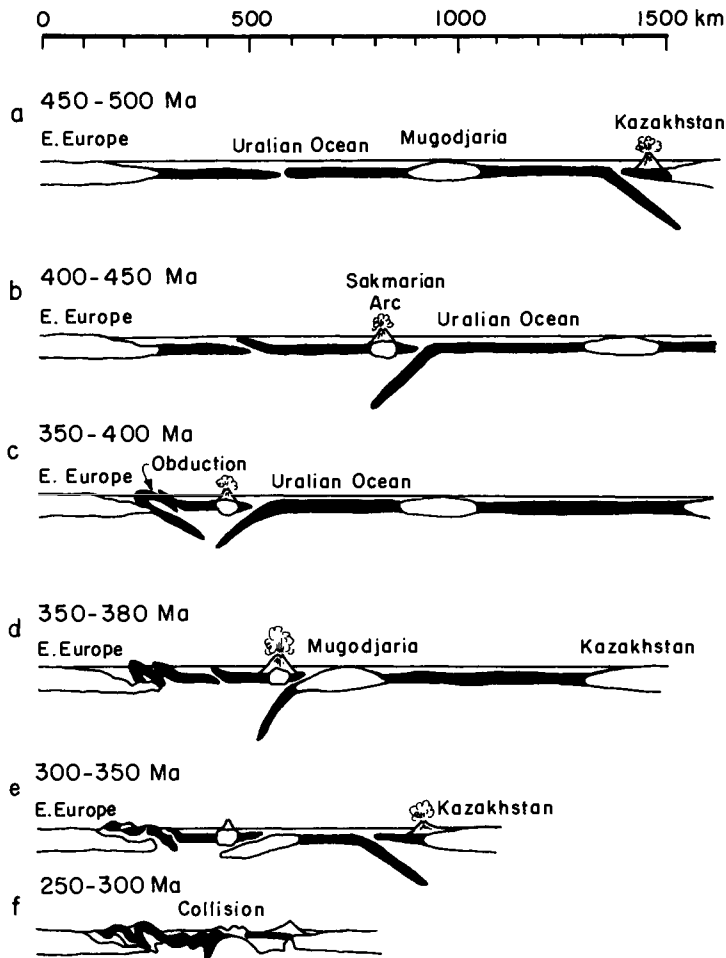


FIG 8.7 Schematic evolution of the southern Ural Mountains during the Paleozoic. Modified after Zonenshain et al. (1984).

back-arc crust was obducted onto the European plate as the Sakmarian back-arc basin closed (c). This was followed by a short-lived reopening of a back-arc basin and collision of Mugodjaria with the Sakmarian arc at about 350 Ma (d). This collision caused subduction to shift eastward with a new convergent margin dipping beneath the Kazakhstan plate (e). Continued closing of the Uralian ocean basin resulted in a continent–continent collision in the Permian (f). The history of the Urals covers about 400 Ma, with the first 250 Ma involving creation of new oceanic crust, followed by 100 Ma of ocean closure and subduction, and then 50 Ma of collisional deformation.

The Cordilleran System

Accretionary Terranes

The conclusion that much of the Cordillera in western North America is composed of a collage of accretionary terranes is now well documented by geological,

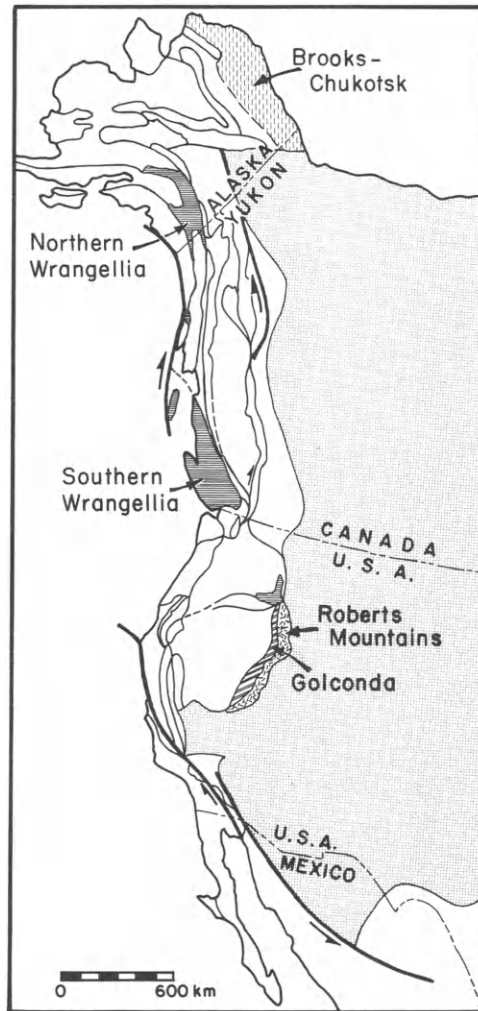


FIG 8.8 Distribution of major Cordilleran accretionary terranes in western North America. After Coney et al (1980). Terranes discussed in the text are labeled.

paleontological, and paleomagnetic evidence (Coney et al., 1980; Jones et al., 1986) (fig. 8.8). More than 200 terranes that lie west of the Precambrian craton have been recognized in the Cordillera. Most of these terranes have been added to North America during the Mesozoic and Cenozoic, during which time the continental margin was extended by as much as 800 km. Paleozoic accretion is confined chiefly to Nevada. Most Cordilleran accreted terranes appear to represent fragments of continents, oceanic plateaus, or portions of arc systems, some of which traveled >5000 km from their sources. Collision with North America occurred, in part, along transform faults with right lateral motion, and other terranes were accreted as flakes during subduction.

One of the best-known terranes is Wrangellia which docked with North America in the Late Mesozoic. Wrangellia contains upper Paleozoic arc-related assemblages

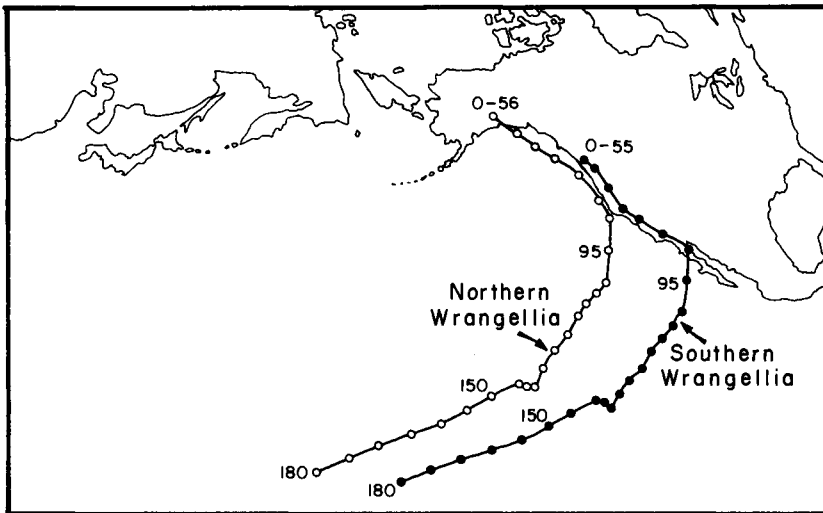


FIG 8.9 Paleomagnetic reconstruction of trajectories of Northern and Southern Wrangellia. After Debiche et al. (1987). Ages given in Ma.

overlain by Permian cratonic sediments. Triassic rocks include pelagic black shales and carbonates and associated basalts, and Jurassic and Cretaceous rocks are predominantly clastic sediments. Wrangellia has been fragmented by strike-slip faulting and major fragments occur in SE Alaska and southern British Columbia (fig. 8.8). Faunal populations, however, are essentially the same in all fragments, indicating a similar history. Both paleomagnetic data and faunal assemblages indicate that, during the Triassic, Wrangellia was situated near the equator. Paleomagnetic data suggest that Southern Wrangellia (SW British Columbia) travelled thousands of kilometres from the Southern Hemisphere on the Farallon plate colliding with North America near Baja California about 90 Ma ago (fig. 8.9). It was then driven northward along the coast by strike-slip faulting until it reached its present position at about 55 Ma. Northern Wrangellia followed a parallel trajectory, colliding with northern California at about 90 Ma and thence traveling northward to southern Alaska by about 56 Ma.

Western United States

Late Precambrian and Paleozoic

During the Late Precambrian and Early Paleozoic, the western coast of North America was a passive continental margin (fig. 8.10a) The presence of aulacogens suggests that a portion of the Precambrian craton was rifted off western North America between 650 Ma and 850 Ma (Dickinson, 1981). In Canada, rifting preceded deposition of the Belt Supergroup at about 1450 Ma. Possible candidates for the continental fragments(s) rifted away are the Precambrian basement blocks in Siberia. Cratonic sediments were continuously deposited along the new western continental margin until the Late Devonian with cumulative thickness reaching 10 km in some areas. The northern termination of the passive margin sediments in central Idaho (fig. 8.10a) may be caused by offsets along Precambrian faults that developed during the rifting.

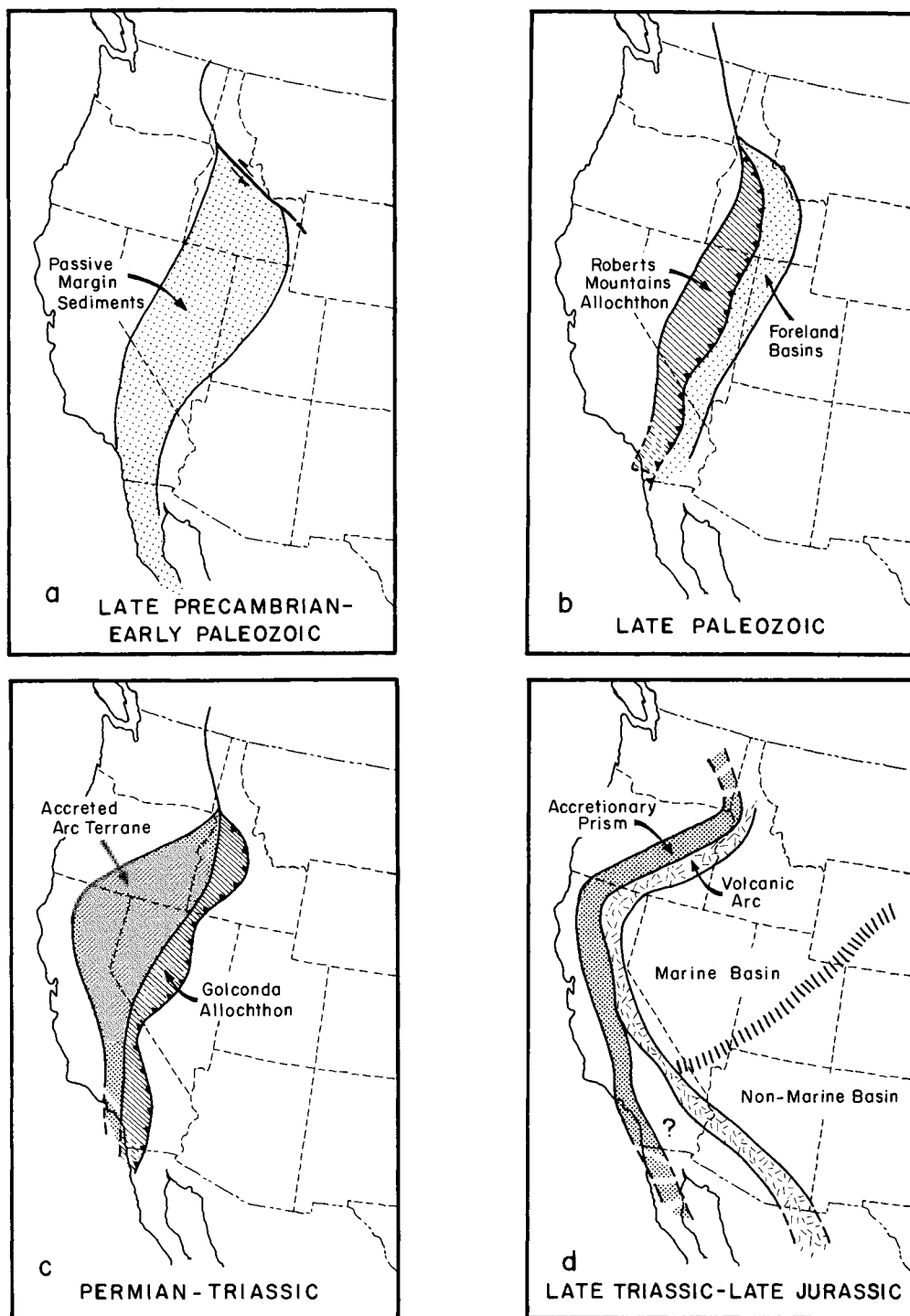
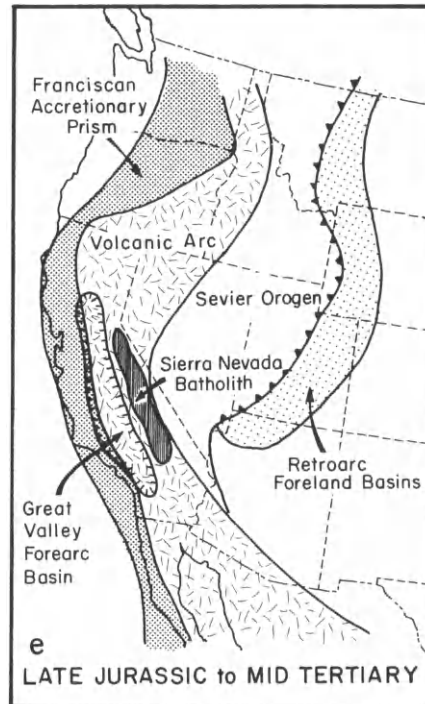


FIG 8.10 Major tectonic elements in the Phanerozoic evolution of the western United States. After Dickinson (1981).



By the Ordovician, one or more arcs had formed off the West Coast. The *Antler orogeny* (300–375 Ma) and the *Sonoma orogeny* (200–280 Ma) probably reflect collisions of these arcs with the passive continental margin (fig. 8.10b,c). During the Antler orogeny the *Roberts Mountains allochthon* was thrust over the passive continental margin (fig. 8.11). This allochthon contains turbidites, cherts, and volcanics of arc origin. During the collision, a long series of foreland basins formed in eastern Nevada (Speed and Sleep, 1982) and as the Roberts Mountains thrust moved eastward, so did the foreland basins.

Mesozoic to Mid-Tertiary

During the Permo-Triassic Sonoma orogeny, island arc terranes exposed in northern California and northwestern Nevada were accreted to the continental margin. During the Middle Triassic the *Golconda allochthon* was thrust part way over the Roberts Mountains allochthon (fig. 8.11). Stacking of slices of terrigenous sediments and volcanics with slices of hemipelagic sediments and submarine basalts in the Golconda allochthon suggest a long transport distance.

Beginning in the Late Triassic, a continental-margin arc system developed along the western coast of the United States. This probably developed as a flip or reversal in arc polarity following the Golconda collision. From Late Triassic to Late Jurassic the arc system circled from Oregon through eastern California into Arizona (fig. 8.10d). The accretionary prisms associated with the arc are exposed in the Klamath Mountains and Sierra Nevada in California. Volcanic detritus from the arc was partially shed into a large foreland basin now occupied by the Colorado Plateau, and bentonitic shales of the Chinle and Morrison Formations are representative of this

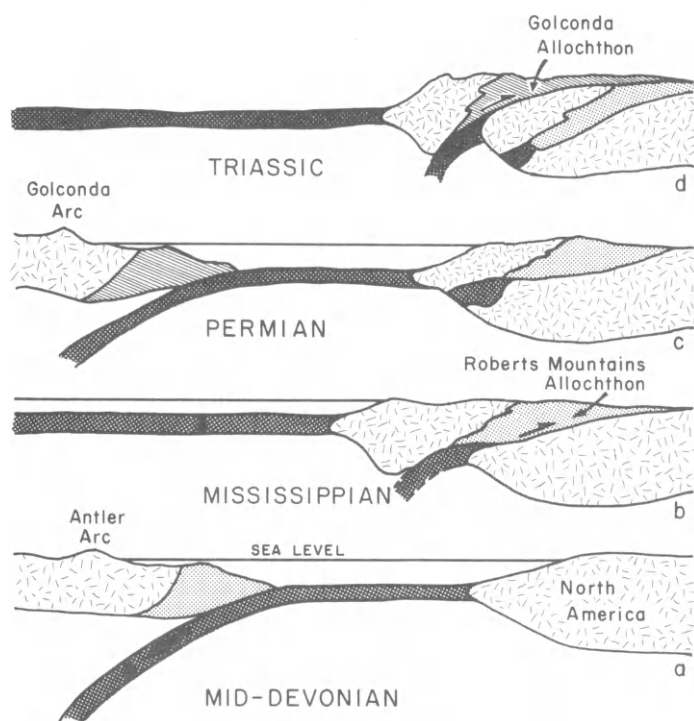


FIG 8.11 Diagrammatic cross-sections showing the tectonic development of the Roberts Mountains and Golconda allochthons. Constructed from data in Speed and Sleep (1982) and Babaie (1987).

input. During the Middle to Late Jurassic, several arcs collided with the coastal arc system causing the *Nevadan orogeny* (140–150 Ma). These arcs were accreted to the continent and the subduction zone stepped oceanward. In the Cretaceous the Franciscan Complex in California was the accretionary prism, the Great Valley Sequence accumulated in an elongated forearc basin, and the Sierra Nevada batholith represents the root zone of the arc (fig. 8.10e). Behind the arc system a wide fold-thrust belt developed in Utah and Nevada producing the *Sevier orogeny* (80–130 Ma), and prominent retroarc foreland basins formed east of the thrust belt. By the end of the Cretaceous, arc magmatism ceased in the Sierra Nevada and deformation shifted eastward, causing the *Laramide orogeny* (50–80 Ma). The different styles of deformation characterizing the Sevier (chiefly eastward-directed thrusts) and Laramide orogenies (chiefly westward-directed thrusts) continues to be a subject of interest among geologists. Any acceptable model must explain the rapid eastward displacement of deformation (between the Sevier and Laramide orogenies) and the localization of Laramide deformation in a Narrow belt around and in the Colorado Plateau. Also notable is a geographic gap in volcanic activity during the Laramide in this region (fig. 8.12).

One model that has been moderately successful in explaining these features calls upon buoyant subduction of the Farallon plate beginning about 80 Ma ago (Cross, 1986). Prior to 80 Ma, deformation was compressive, concentrated in the Sevier fold-thrust belt (fig. 8.13), and beginning at about 75 Ma, North America changed motion (with reference to Pacific hotspots) and moved perpendicular to the west

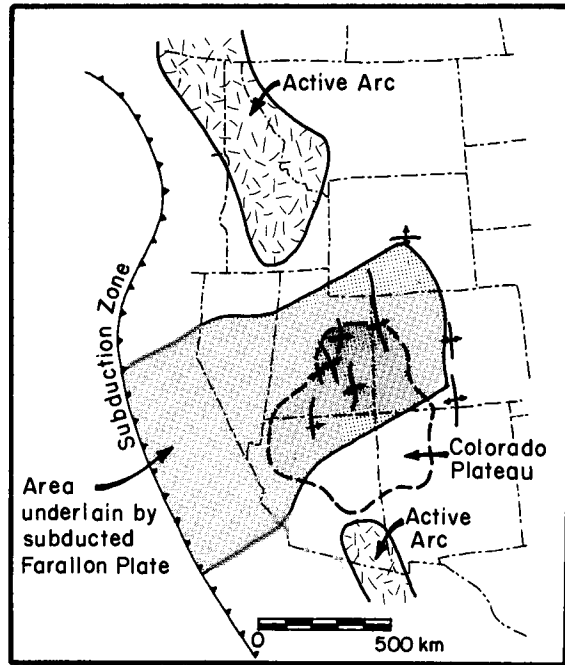


FIG 8.12 Diagrammatic map showing the major tectonic features in the central Cordillera at about 70 Ma. After Cross (1986). Fold symbols show area of Laramide deformation.

coast trench. This change corresponded to a period of rapid relative convergence of the North American and Farallon plates. These factors may have caused North America to override the trench and induce low-angle, buoyant subduction. Because the mantle wedge was largely eliminated, magmatism also stopped during shallow subduction. Laramide deformation may reflect a combination of: (1) a brittle character of crust east of the Sevier orogen and foreland basin, and (2) widespread transmission of stresses from the near-horizontal Farallon plate into the overlying plate. Laramide deformation is concentrated along the juncture between the double and single thicknesses of lithosphere (fig. 8.13, 60–65 Ma). This juncture should represent the position of greatest contrast in mechanical properties of the lithosphere and thus the loci of least strength. The model also explains the relative tectonic rigidity of the Colorado Plateau; i.e. the plateau is underpinned and supported by the Farallon plate. It has been suggested that late Laramide deformation (50–60 Ma) reflects subduction of an aseismic ridge beginning at about 65 Ma. The end of the Laramide orogeny at about 50 Ma corresponds to the time that subduction slowed or stopped, and the Farallon plate decoupled from the North American plate and sank into the mantle (fig. 8.13, 45 Ma).

Late Cenozoic

Tectonic elements peculiar to the Cordilleran Late Cenozoic include core complexes, the San Andreas transform fault system, and the Basin and Range Province. Beginning in the Mid-Tertiary arc magmatism began again along the west coast as

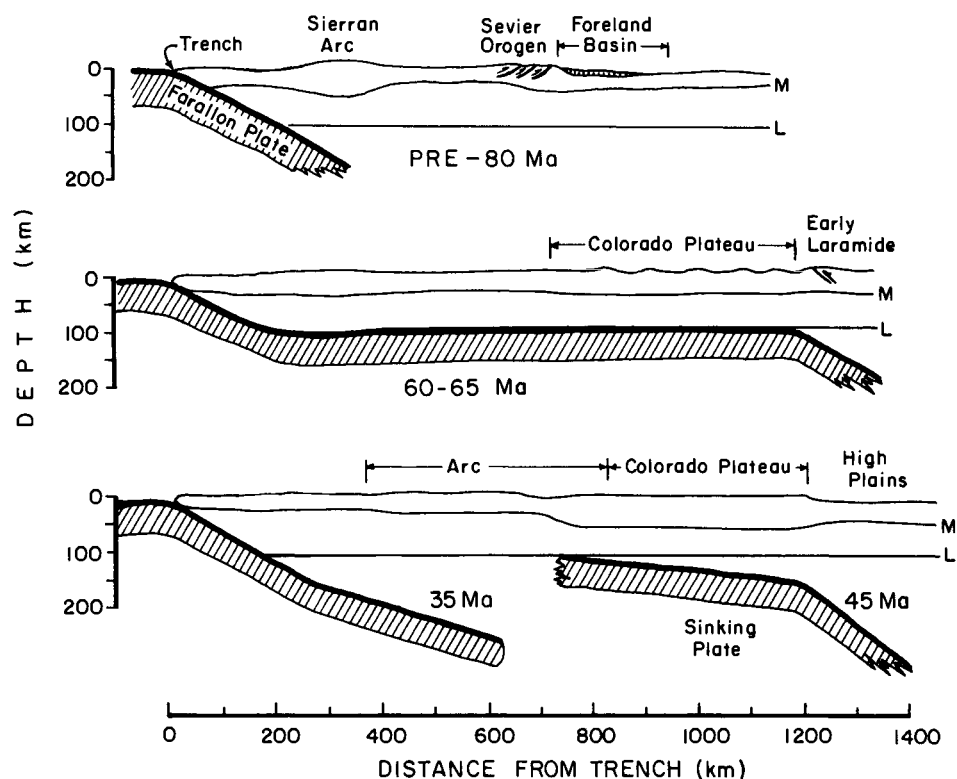


FIG 8.13 Cross-sections showing buoyant subduction of the Farallon plate in the Late Cretaceous and Early Tertiary. After Cross (1986).

the Farallon plate again began to subduct. By the Miocene, a continuous arc extended south along the coast from southern Canada. About 30 Ma ago the East Pacific rise approached the subduction zone and subduction of the rise began by about 25 Ma ago (fig. 8.14a). There is still disagreement as to whether the rise was subducted or annihilated (Atwater, 1970; Elston, 1984). The annihilation model has the advantage of also accounting for the origin of the San Andreas transform fault (fig. 8.14b,c). Such a fault must develop if the Pacific plate is not subducted and the motion on the fault (right lateral) is dictated by the relative motions of the Pacific and North American plates. Each end of the fault is a triple junction that migrates north and south, respectively, as subduction continues. The underplated, hot oceanic lithosphere heats the overriding plate, resulting in localized partial melting and magmatism. As ductile extension continues, partial melting of the lower crust produces granitic melts (Elston, 1984).

As ductile extension begins to subside around 20 Ma, core complexes rise isostatically. *Core complexes* are basement domes composed of highly deformed metamorphic-plutonic rocks overlain by relatively undeformed supracrustal rocks. Separating the two is a detachment fault (Coney, 1980) (fig. 6.20), and rocks beneath the detachment surface are mylonitic gneisses with a well-developed lineation formed by ductile shear. More than 30 core complexes have been described in the American Cordillera, extending from southern Canada into Mexico. Most of these

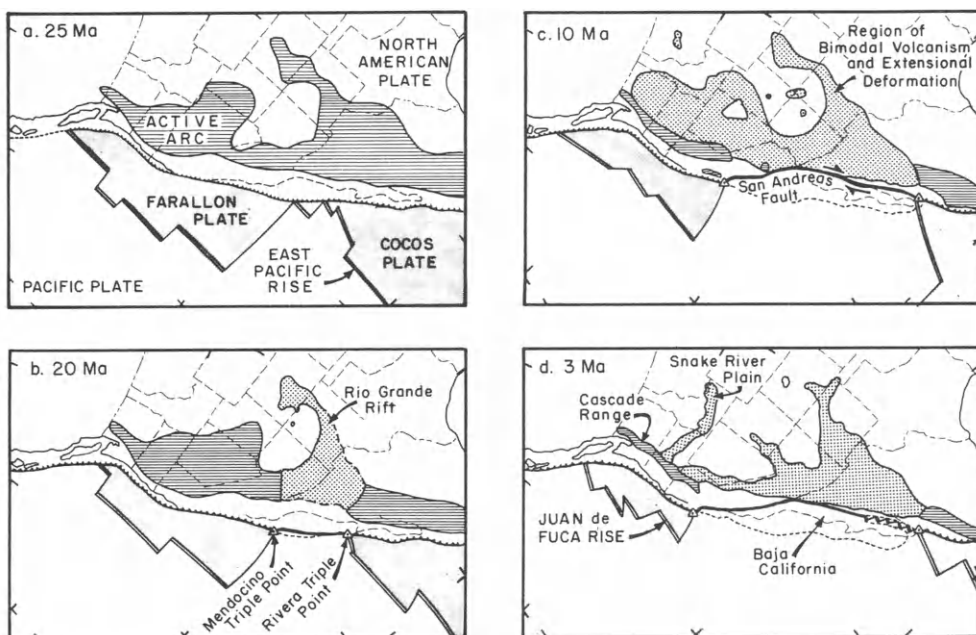


FIG 8.14 Progressive annihilation of the East Pacific rise and changes in volcanic style and tectonic setting in the last 25 Ma. Modified after Atwater (1970), Christiansen and Lipman (1972), and Dickinson (1981).

developed prior to Basin and Range faulting in the time interval of 15–50 Ma ago. Although the origin of core complexes is not well understood, they appear to be related to crustal extension and heating. Beginning 5–10 Ma ago the North American plate cooled sufficiently to allow widespread normal faulting in the crust. At this time the Basin and Range Province and the Rio Grande rift began to develop. Also, the San Andreas fault lengthened as the triple junctions continued to migrate north and south (fig. 8.14c). At 3–4 Ma Baja California decoupled from the North America plate and became part of the Pacific plate as the Gulf of California began to open (d), and subduction gradually stopped along the northwestern coast.

Accompanying subduction of the East Pacific rise and lengthening of the San Andreas transform fault, changes in tectonic and volcanic style occurred in western North America (Christiansen and Lipman, 1972). On land opposite the lengthening transform fault, volcanic activity changed from calc-alkaline to bimodal and deformation from compressive to dominantly extensional (fig. 8.14). At 10–20 Ma ago basalts of the Columbia River Plateau were erupted in an extensional basin behind the Cascade volcanic chain.

Arctic Region

The entire width of Arctic Cordillera in Canada and Alaska is composed of accretionary terranes. Accretion began in the Early Cretaceous and, although continuing today, it was chiefly completed by the Mid-Tertiary. Most terranes appear to represent thin sheets detached from their original lower crust and

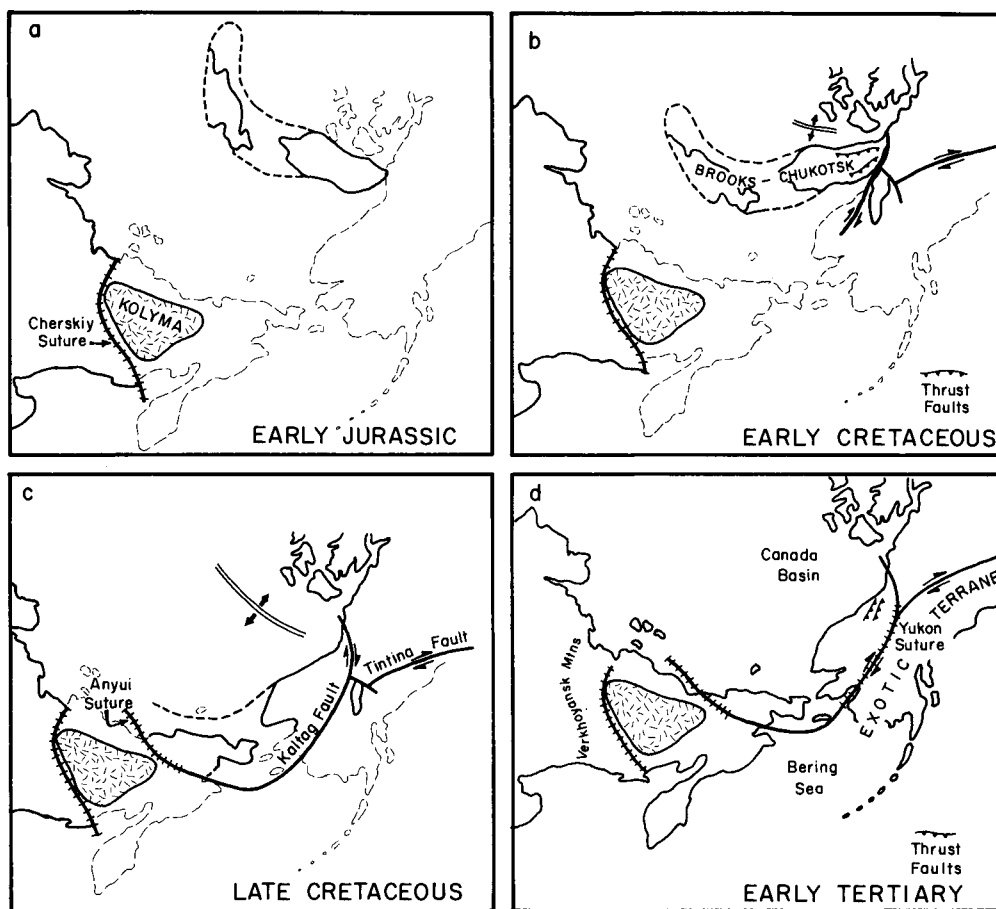


FIG 8.15 Tectonic reconstruction of the Cordilleran Arctic region since the Early Jurassic. Constructed from data given in Sweeney (1981) and Howell and Wiley (1987).

lithospheric basement. After major accretion, Alaska was further telescoped, wrenched, and internally rotated, chiefly by intraplate processes. This occurred in response to continued northwestward motion of terranes in western Canada, eastward motion of Siberia caused by opening of the North Atlantic basin, and perhaps northward compression caused by subduction of the Pacific plate.

Paleomagnetic data indicate that northern Alaska (fig. 8.15) rifted away from Arctic Canada and rotated about 70° counterclockwise to its present position. Because there is not a plate boundary in the Bering Strait, the Chukotsk Peninsula was part of northern Alaska during this rotation. The Kolyma microplate in NE Siberia collided with Siberia in the Early Jurassic and was welded along the Cherskiy suture (fig. 8.15a). It was carried northward by the oceanic Kula plate which was consumed in a paleo-Aleutian subduction zone. The counterclockwise rotation of the Brooks–Chukotsk terrane began in the Early Cretaceous, probably in response to the initiation of opening of the North Atlantic basin (Sweeney, 1981) (b). Paleomagnetic data indicate the rotation was largely completed by Mid-Cretaceous when Chukotsk collided with Kolyma, forming the Anyui suture (c). The rotation resulted

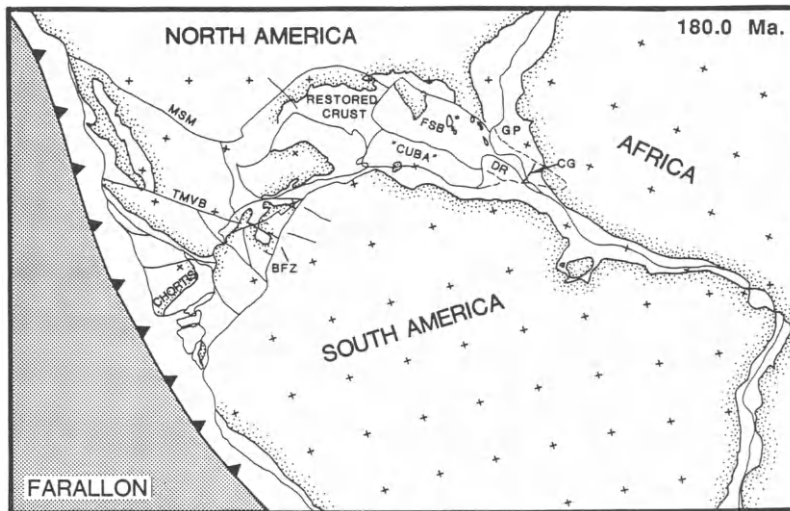


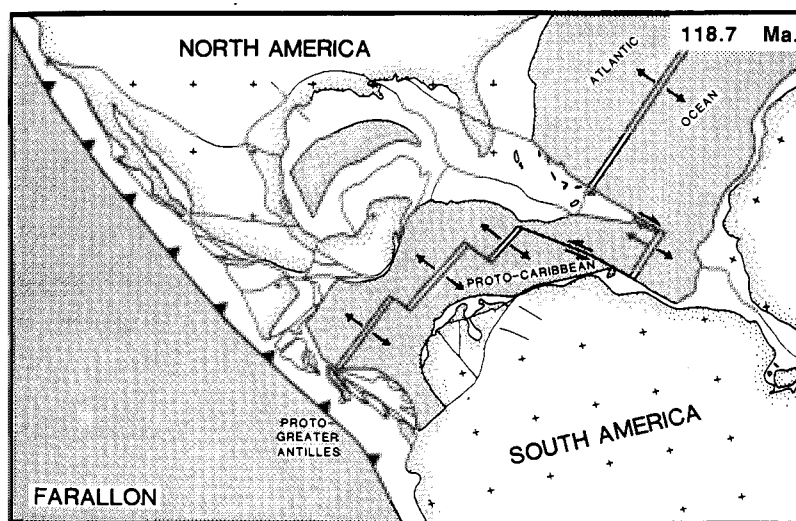
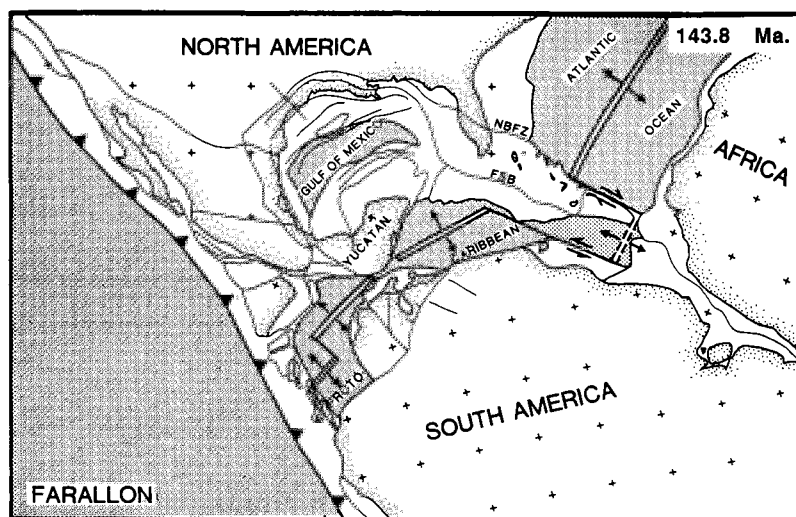
FIG 8.16 Tectonic evolution of the Caribbean region. From Ross and Scotese (1988). Abbreviations: CG, Cat Gap; FSB, Florida Straits block; MSM, Mojave-Sonora megashear; TMVB, Trans-Mexican volcanic belt; NBFZ, Northern Bahamas fracture zone; BP, Bahamas platform; CR, Cayman ridge; YB, Yucatan basin; HE, Hess escarpment; MPJFZ, Motagua fault zone; CT, Cayman trough; AR, Aves ridge; PRB, Puerto Rico basin; NRS, Nicaragua rise shelf; SMFB, southern Mexican fold belt; EPR, East Pacific rise; PFZ, Panama fracture zone.

in compression causing thrusting in the Brooks Range in Alaska (Howell and Wiley, 1987). Continued opening of the North Atlantic in the Late Cretaceous may have caused deformation in the Verkhoyansk foldbelt (west of Kolyma), oroclinal buckling of the Brooks–Chukotsk terrane and right lateral displacement along the Kaltag and Tintina faults (c). Several small terranes were also accreted to central Alaska during the Cretaceous. Finally, during the Early Tertiary (~55 Ma) a single exotic fragment composed of numerous, previously welded terranes was emplaced south of the Kaltag–Tintina fault system (d). Thrusting in NE Alaska during this time suggests continued counterclockwise rotation of northern Alaska. During final suturing of the Brooks–Chukotsk terrane about 50 Ma ago, a portion of the Kula plate was trapped and probably forms the Bering Sea basin.

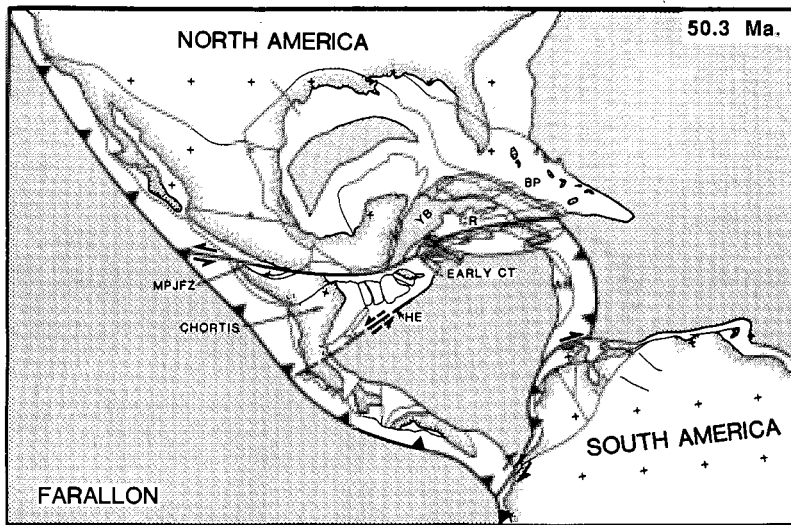
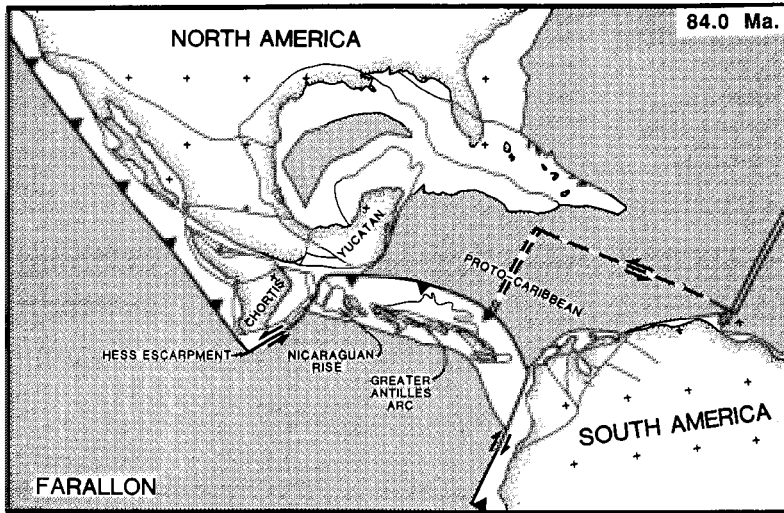
Caribbean Region

The Caribbean region has undergone an exceedingly complex tectonic history involving numerous microplates and large transcurrent displacements between blocks. The Caribbean plate appears to represent a preserved fragment of the Farallon plate and is therefore allochthonous with respect to North and South America (Ross and Scotese, 1988). The evolution of the Caribbean region is controlled by interactions of surrounding major plates. Paleopositions of North and South America can be determined by finite-difference solutions that fit magnetic anomaly pairs, and small plate trajectories can be estimated from offsets along transcurrent faults and other geologic data.

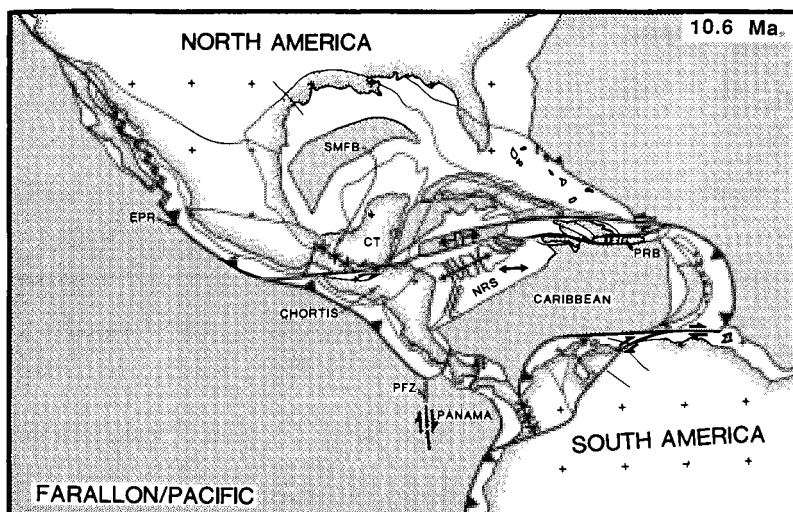
A possible reconstruction of microplates in the Caribbean area in the Early Jurassic is shown in fig 8.16 (180 Ma). Beginning in the Mid-Jurassic, North



America rifted from Africa and South America as the North Atlantic opened. This opening propagated into the Gulf of Mexico along the North Bahamas fracture zone and about 500 km of left-lateral displacement occurred along the Mojave-Sonora megashear and the Trans-Mexican volcanic belt, moving much of Mexico into the Caribbean region (fig. 8.16, 143.8 Ma). The Farallon plate was continuously subducted along the west coast of the American plates during this time. By Late Jurassic, oceanic crust was generated in the Gulf of Mexico. During the Early Cretaceous, South America continued to separate from Yucatan as the proto-Caribbean basin opened (118.7 Ma). By Late Cretaceous, seafloor spreading ended in the proto-Caribbean, and the Farallon plate continued to move northeastward between the American plates carrying the Greater Antilles arc (which included Cuba and Hispaniola) on its leading edge (84 Ma). During the Early Tertiary the Greater



Antilles arc continued to move northeastwards, consuming the proto-Caribbean oceanic crust. The arc also collided with the Yucatan peninsula and the Bahamas platform, resulting in ophiolite obduction in Yucatan. Also in the Early Tertiary, perhaps in response to the collision with the Bahamas platform, the Middle America subduction zone formed, isolating the Caribbean plate. The Farallon plate was not consumed in this new subduction zone. The Yucatan Basin began to open and the Cayman ridge transform fault developed along its southern margin (50.3 Ma). This fault, which became the new northern margin of the Caribbean plate, transferred the Chortis block (Nicaragua, Honduras) from the North American to the Caribbean plate. During the Late Tertiary the Panama arc continued to grow, the Gulf of California began to open, Hispaniola and Puerto Rico continued to move eastward and the Chortis block moved into its present location south of Yucatan (10.6 Ma).



From this time to the present the Caribbean plate has had major strike-slip faults along both its northern and southern boundaries.

The Andes

The Andean Cordillera, which extends for about 10^4 km along the west coast of South America, has undergone a complex and diverse tectonic history (Dalziel, 1986). In much of Peru and northern Chile the belt is underlain by Precambrian basement, while the southern part of the belt is underlain by a collage of Paleozoic accretionary terranes. The Arequipa terrane in southern Peru is a Proterozoic block that appears to have accreted to South America in the Late Proterozoic or Early Paleozoic. The Andean Cordillera formed during the Mesozoic and Cenozoic on an eroded surface of uplifted and eroded Precambrian and Paleozoic rocks.

Volcanism began in the Triassic and Jurassic as a continental-margin arc became established along the western coast of South America. During this time a series of interconnected back-arc basins formed along the entire length of the belt south of Ecuador. These persisted well into the Early Cretaceous. In southernmost Chile ($47\text{--}56^\circ\text{S}$), new oceanic crust formed in the back-arc basin (as represented by; the Sarmiento and Tortuga ophiolites), while in central Chile large batholiths were emplaced in the back-arc basins which were underlain with continental crust (Bartholomew and Tarney, 1984). Beginning in the Mid-Cretaceous, compressive deformation began as the back-arc basins closed. Uplift and deformation is reported all along the Andean chain at this time. The axis of magmatism and deformation migrated eastward, as reflected by the emplacement of Mid-Cretaceous batholiths and the development of foreland basins on the east side of a major fold-thrust belt. There appears to be no evidence for collision of accretionary terranes at this time. The Triassic and Jurassic strata in the Andean belt rest unconformably on Precambrian and Paleozoic rocks with no evidence of thrusting (Dalziel, 1986). Unlike collisional orogens, the amount of crustal shortening along most of the Andes appears to be small. The exception is southernmost Chile, where an oceanic back-arc basin (the Magellan basin) closed with significant crustal shortening.

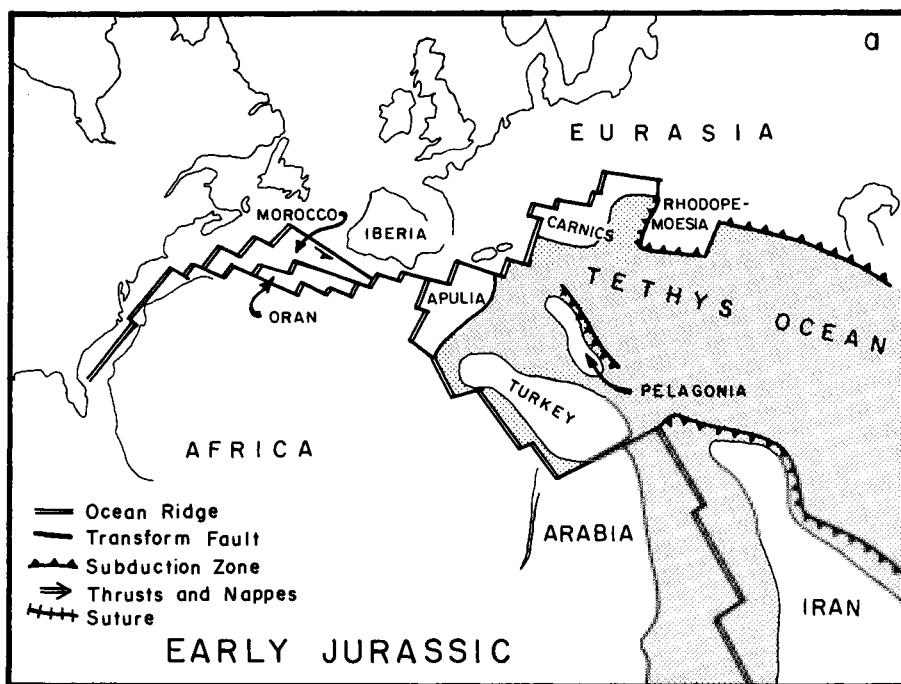


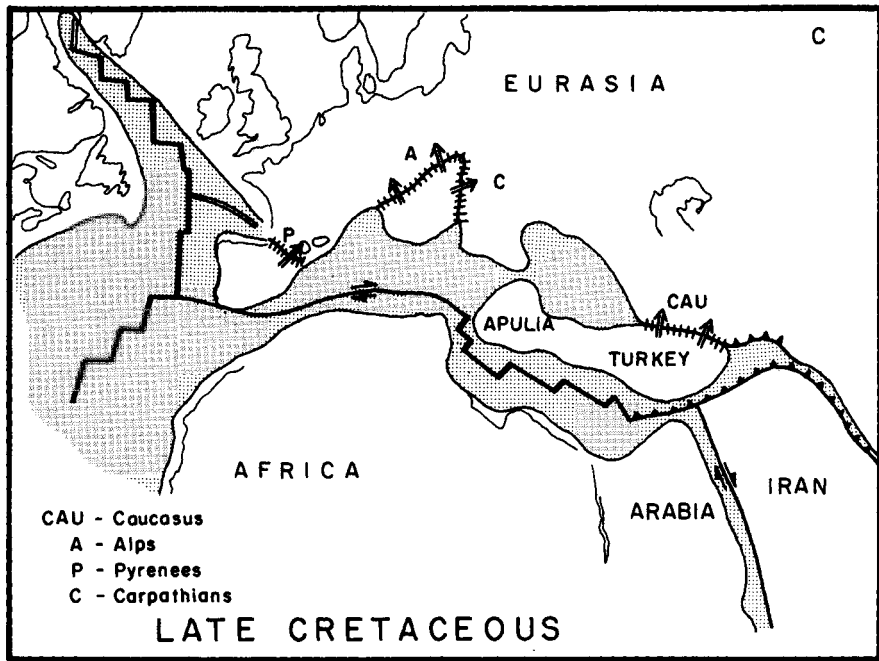
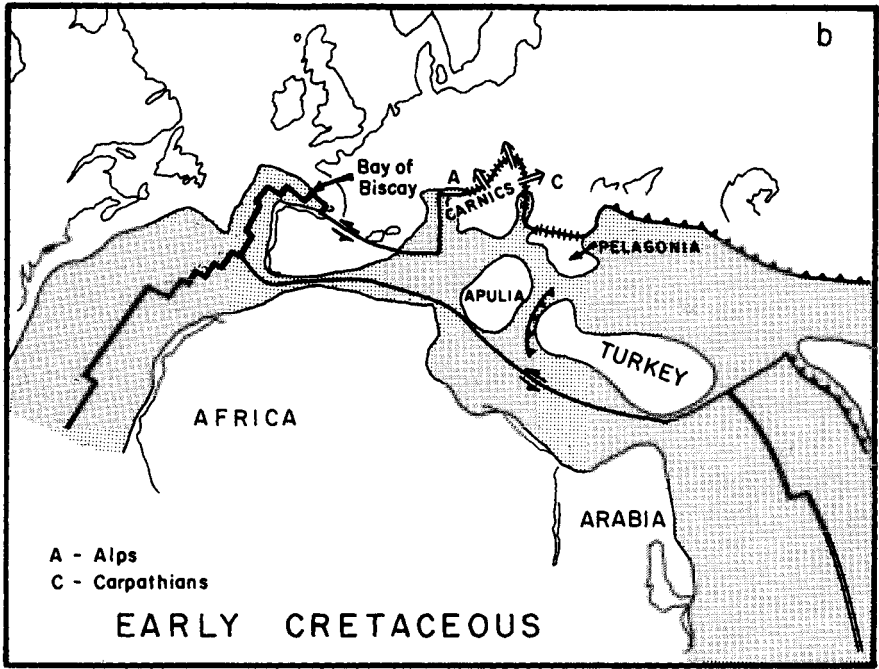
FIG 8.17 Schematic plate tectonic evolution of the Mediterranean region. After Dewey et al. (1973) and Ziegler (1987).

It seems probable that collisional accretion of exotic terranes did not play a major role in the Late Cretaceous and Tertiary Andean orogeny. It is possible that opening of the South Atlantic basin between 110 and 120 Ma may have resulted in increased convergence rates along the Pacific margin of South America. This should result in an increase in the horizontal compressive stress in the overriding South American plate, which in turn may have caused closure of the Andean back-arc basins and be responsible for the Cordilleran orogeny in this belt.

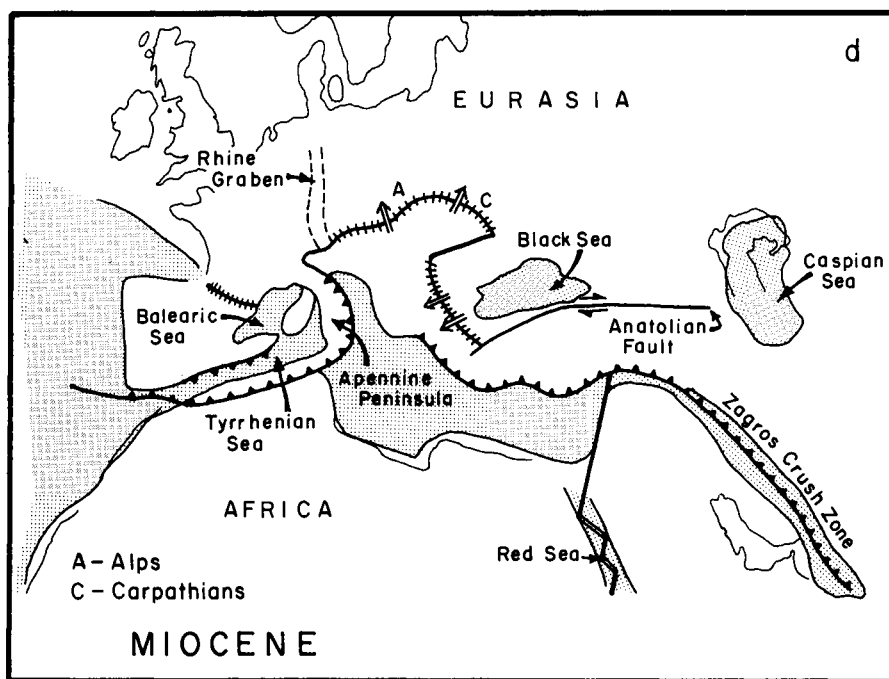
The Alpine Orogenic System

The tectonic history of the Mediterranean region during the last 180 Ma is characterized by fragmentation of minor plates from northern Africa followed by collision of these plates with southern Europe. Such collisions are responsible for widespread compressive deformation that played a major role in formation of the Alpine orogenic belt in western and central Europe. During the last 180 Ma, Africa is characterized by counterclockwise rotation resulting from closure of the Tethys Ocean between Eurasia and Africa. Based on geological and paleomagnetic evidence, the major tectonic events accompanying this closure are summarized in fig 8.17 (after Dewey et al., 1973; Smith and Woodcock, 1982; Ziegler, 1987).

Beginning about 180 Ma ago, the North Atlantic began to open with associated complex rifting in southern Europe and North Africa (fig.8.17a). Morocco, Oran, Pelagonia, Apulia, and Turkey were rifted from North Africa at this time, and Carnics was rifted from southern Europe. The Iran plate, which began to separate



from Africa–Arabia in the Late Triassic, continued to move northeastwards as an ocean basin opened between Arabia and Iran. Marine transgression in the Early



Jurassic around the margins of the western Tethys Ocean supports the existence of active oceanic ridge systems in the area. Major subduction zones are along the northern edge of the Tethys Ocean beneath the present-day Major Caucasus Mountains and along the northeastern edge of Iran. A less certain subduction zone may have existed west of Rhodope-Moesia. By Mid-Jurassic the accreting plate margin along the southwestern boundary of Oran evolved into a transform fault system, and both Oran and Morocco closed on North Africa. In the Late Jurassic the Mid-Atlantic ridge shifted westward toward Africa, and subduction continued along the northern and southern margins of the Tethys basin and west of Rhodope-Moesia. Subduction also appears to have begun along the northeastern margin of the Turkey plate and the southeastern margin of the Iran plate. Morocco and Iran became sutured to North Africa by the Early Cretaceous. During the Late Jurassic the ocean basin north of Pelagonia closed and Pelagonia collided with Rhodope-Moesia during the Early Cretaceous (b). Continued rapid seafloor spreading in the North Atlantic led to the development of a transform fault south of Turkey. Convergence of Carnics on two sides resulted in the onset of compressive deformation in the Alps and Carpathians. Also beginning in the Early Cretaceous, Iberia was rifted away from France as the Bay of Biscay began to open.

Beginning about 80 Ma ago, Greenland started to separate from Canada resulting in a reorganization of plate boundaries (fig. 8.17c). A new subduction zone developed west of Carnics and Africa continued its northwestward rotation, further closing the numerous seaways of the western Tethys Ocean. This rotation also changed the motions of Apulia and Turkey to more westward courses. During the Late Cretaceous an extremely complex pattern of subduction zones, spreading centers, and transform faults existed in the western Tethyan area, and during the

Early Tertiary major deformation and volcanism spread throughout the region. Major convergent boundaries developed on both sides of Iberia. Beginning in the Early Cretaceous and continuing into the Late Cretaceous, Turkey collided with Eurasia and major deformation occurred in the Caucasus. Collision between Iberia and Europe produced major deformation in the Pyrenees and obduction of oceanic crust occurred in Corsica. Collision also continued along the western margin of Carnics, as indicated by nappes in the Alps. Calc-alkaline volcanism is widespread in Iran during the Eocene, reflecting active subduction zones along both margins of the Iran plate. Renewed convergence of Carnics during the Paleocene gave rise to extensive thrusts and nappes in the Alps and Carpathians and was manifest for over 1000 km north of the Alpine front. During the Eocene-Oligocene the Alpine belt began to subside under the load of rapidly advancing nappes and the Molasse basin, a major foreland basin developed in front of the northward-advancing orogen.

During the Miocene the Anatolian transform fault developed as the Turkey plate moved west in response to continued northwestward rotation of Africa (fig. 8.17d). Major convergence is taken up by choked subduction zones along the western and northern margins of Carnics, by subduction beneath the Apennines (in Italy) and by subduction along the southern margins of the Turkey and Iran plates. Corresponding deformations include renewed thrusting and nappe formation in the Alps and Carpathians. Also at this time the Rhine graben developed. In the Late Miocene, Italy as a migrating arc system began to separate from Corsica-Sardinia as the Balearic and Tyrrhenian Seas opened. The Calabria subduction zone (Plate I) is a remnant of the once extensive subduction system around the Apennine Peninsula. Also at about 30 Ma the Red Sea began to open as Arabia was rifted away from Africa. During the Pliocene the northern margin of Arabia collided with Iran, and the Zagros crush zone (Plate I) formed. This collision led to intense folding along the coast of Arabia and the southward spreading of clastic sediments onto the Arabian plate. Remnants of the Tethys Ocean are isolated by collision of Iran with Asia, and today are represented by the Black and Caspian Seas. The modern Hellenic subduction zone south of Greece developed at about 6 Ma ago in response to the choking of subduction zones to the north.

Eastern Asia

Paleozoic-Mesozoic

Most of eastern Asia has grown by terrane accretion beginning in the Late Paleozoic and largely completed during the Middle Mesozoic (Xiong and Coney, 1985; Zhang, 1985). Thirty-three accretionary terranes have been recognized in China and six in Japan (fig. 8.18). Amalgamation of some terranes in central China began in the Silurian, and these collided and were accreted to the Mongolian-Siberian plate in the Late Paleozoic. A Late Paleozoic fold belt extending from western Japan south, including Taiwan and part of SE China, appears to have accreted to the Precambrian South China plate in the Permian or Early Triassic. It is not clear if these terranes accreted to eastern Asia in the Late Paleozoic were part of the same landmass prior to fragmentation in the Early Paleozoic. A large number of terranes were accreted to the southern coast of Siberia during the Triassic, and much of Japan was accreted to the growing craton at this time. The widespread Indonsinian orogeny in SE Asia and

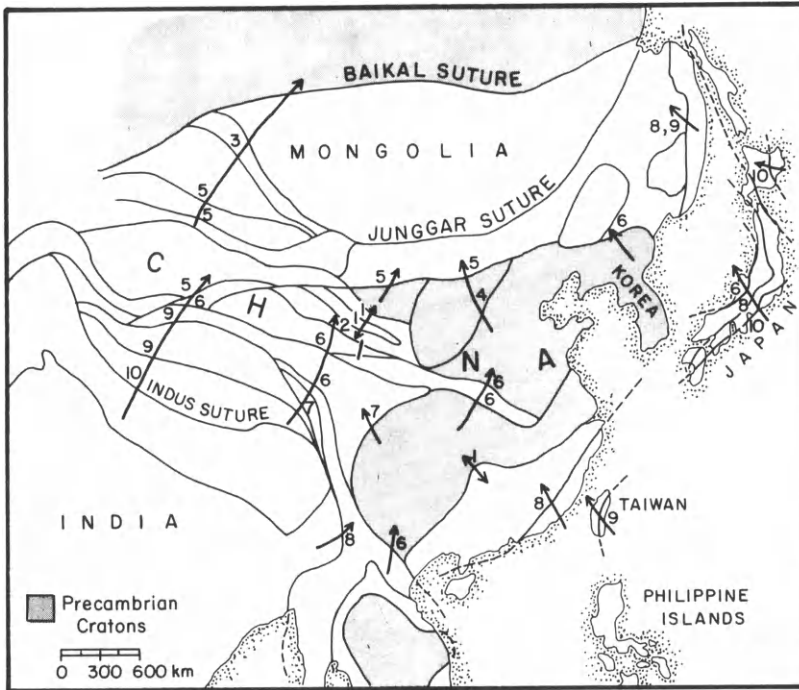


FIG 8.18 Accretionary terranes in Eastern Asia. After Xiong and Coney (1985) and Zhang (1985). Times of terrane accretion or amalgamation given on arrows that indicate directions of accretion. 1, Silurian; 2, Devonian; 3, Mississippian; 4, Pennsylvanian; 5, Permian; 6, Triassic; 7, Early Jurassic; 8, Mid-Jurassic; 9, Cretaceous; 10, Cenozoic.

the Akiyoshi orogeny in Japan reflect collisional deformation associated with these accretions. During the Jurassic and Cretaceous collisional orogenies are recorded in Siberia, Manchuria, northern Japan, and in western China (fig. 8.18). Successive terrane accretions in western China culminated with the India–Tibet collision in the Early Tertiary.

Cenozoic

About 40 Ma ago a change in the motion of the Pacific plate caused it to begin to subduct beneath the Philippine plate. A portion of the Palau–Kyushu arc, which formed over the new subduction zone, is preserved today in eastern Hokkaido (fig. 8.19a). At about this same time Honshu rifted from mainland China as the Sea of Japan began to open. Major opening occurred in the Miocene as the trench–trench–trench triple junction migrated southwest beneath Hokkaido (b). This migration resulted in a collision of the Japan and Palau–Kyushu arcs, the suture of which is exposed in eastern Hokkaido today. This collision fixed the north end of the Japan arc so it could no longer rotate clockwise as a single block. Continued opening of the Sea of Japan, therefore, caused northern Honshu to rotate 40° counterclockwise while southern Japan continued clockwise rotation (Celaya and McCabe, 1987). The net effect is a bend in Honshu (c), and this bending continued until about 12 Ma ago.

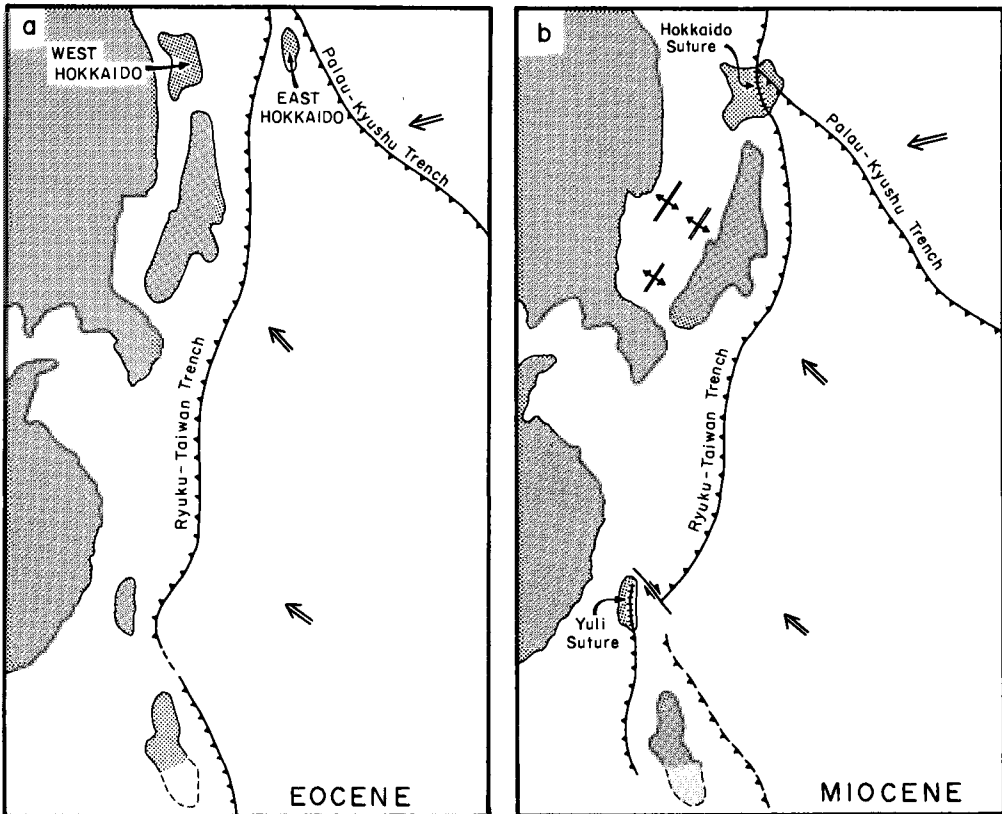


FIG 8.19 Schematic Cenozoic tectonic evolution of eastern Asia. After Celaya and McCabe (1987) and Juan et al. (1983).

During the Early Tertiary the Ryuku-Taiwan arc continued south at least as far as the Philippine Islands (fig. 8.19a). During the Late Miocene and Early Pliocene, subduction began west of Luzon (b) and in the last 5 Ma a major left lateral transform fault developed between Taiwan and Luzon as the Philippine arc collided with the Ryuku-Taiwan arc (c). The collisional zone is exposed on Taiwan as the Yuli suture. Also in the last 5 Ma the east-dipping Luzon subduction zone was initiated in response to continued opening of the South China Sea.

The Himalayas

The rapid northward migration of India during the Tertiary resulted in the India-Tibet collision beginning 50–60 Ma ago. Detailed geological, radiometric and geophysical data from the central Himalayas provide important constraints for the closing of the Tethys Ocean and Himalayan tectonics. Prior to collision, Tibet was a continental-margin arc system characterized by voluminous andesites and felsic ash-flow tuffs, and northern India was a passive continental margin with shelf-facies on the south, passing into a deep-water Tethyan facies on the north. As the collision began, folds and thrusts moved southward from the Indus suture onto the Indian plate (Searle et al., 1987). This resulted in thickening of the crust, high-pressure

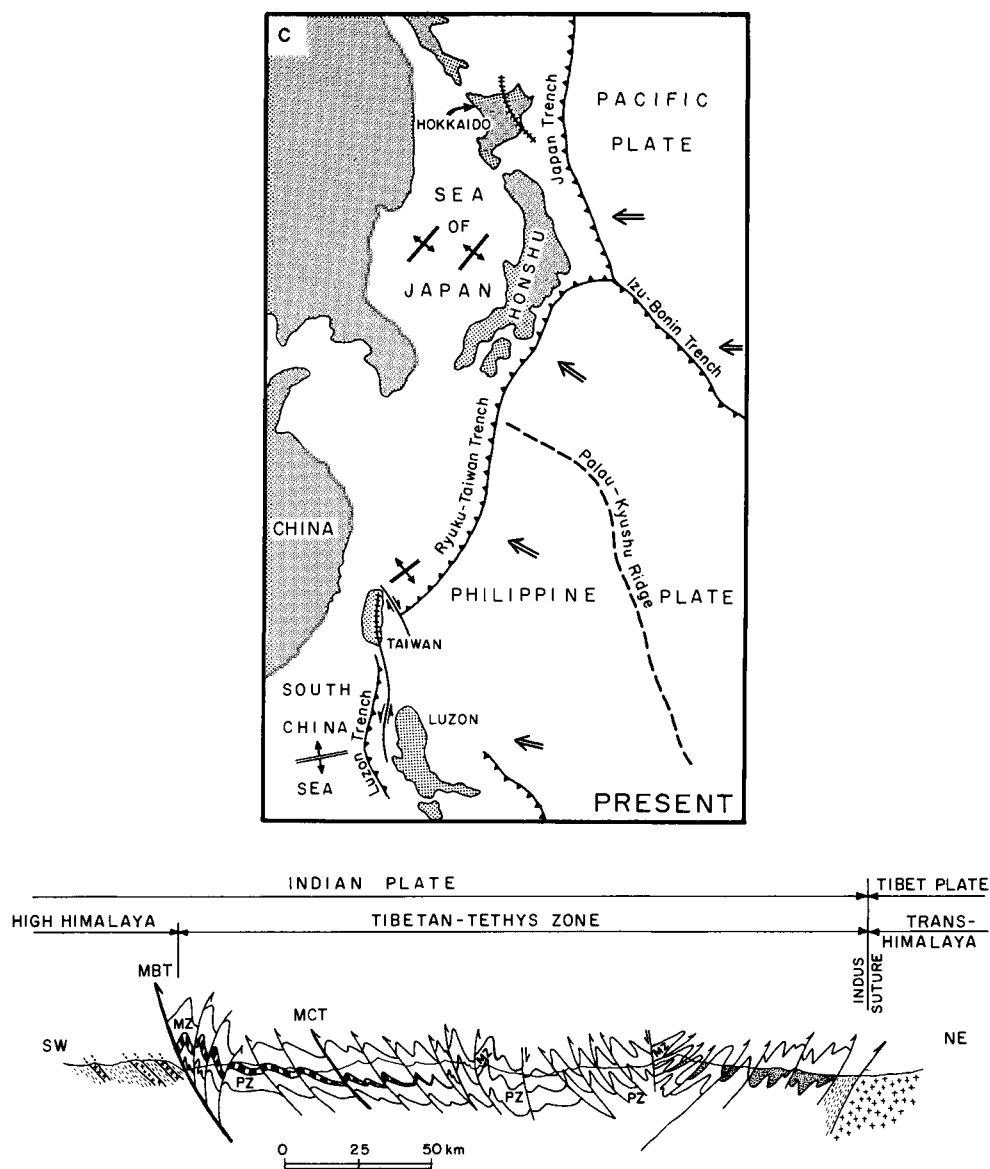


FIG 8.20 Schematic cross-section of the Himalayan orogen. Modified after Searle et al. (1987). MBT, main boundary thrust; MCT, main central thrust; Pz, Mz, Paleozoic and Mesozoic sediments.

metamorphism, and partial melting of the root zones to produce migmatites and leucogranites. Thrusting continued on both sides of the suture as India continued to converge on Tibet. By 40 Ma ago deformation had progressed southward across the Tibetan-Tethys and High Himalayan zones (fig. 8.20). Collapse of the Tibetan-Tethys zone during the Miocene juxtaposed lower Paleozoic shelf sediments north of the Main Boundary thrust against metamorphic rocks and leucogranites south of the thrust. Continued convergence of the two continental plates led to oversteepen-

ing of structures in the Indus suture and finally to backthrusting on to the Tibet plate, as well as continued southward-directed thrusting. The foreland basin continued to move to the south as thrust sheets advanced from the north.

The amount of crustal shortening recorded across the Himalayan orogen is almost 2500 km for a time-averaged compression rate of ~ 5 cm/a (Searle et al., 1987). However, the crust of Tibet is only about 70 km thick, which can account for only about 1000 km of shortening. The remainder appears to have been taken up by strike-slip faulting north of the collision zone (Tapponnier et al., 1986). The Tertiary geological record in SE Asia required 1000–1500 of cumulative strike-slip offset in which India has successfully pushed SE Asia and then Tibet and China in an ESE direction. Most of the Mid-Tertiary displacement has occurred along the left-lateral Red River fault zone (fig. 6.24) accompanying opening of the South China Sea.

Most models for the India-Tibet collision involve buoyant subduction of continental crust. Earthquake data from the Himalayas suggest a shallow ($\sim 3^\circ$) northward-dipping detachment zone extending beneath the foreland basins on the Indian plate. This detachment surface is generally interpreted as the top of the descending Indian plate which probably continues beneath Tibet. Convergence since the Early Miocene has been taken up chiefly along the Main Central thrust (fig. 8.20) and its splays by counterclockwise rotation of India beneath Tibet.

The Western Pacific

The tectonic evolution of the Western Pacific region in the last 150 Ma involves the rapid growth of the Pacific plate at the expense of the Farallon, Nazca, and Kula plates. The Farallon plate is represented today by two small plates, the Juan de Fuca and Cocos plates (Plate I), and the Kula plate is no longer in existence. The Kula–Nazca ridge system originally may have extended into the Tethys Ocean.

Between 200 and 150 Ma the Kula plate was subducted beneath eastern and southeastern Asia and decreased in size as the Pacific plate grew. Data from magnetic anomalies and deep-sea drilling suggest a plate configuration by 135 Ma, as shown in figure 8.21a. The Okhotsk terrane was traveling on the Kula plate and collided with Siberia in the Late Cretaceous. By 100 Ma ago the Pacific plate grew such that only the western end of the Kula plate was consumed beneath Japan (b). The crust of the Bering Sea may represent a remnant of the Kula plate trapped behind the Aleutian arc. The Tethys ocean ridge was subducted beneath Asia about 100 Ma ago, and India began to rift away from Antarctica by 125 Ma. Also, the South China Sea began to open about this time as Borneo separated from mainland China.

By 65 Ma India approached the Tibet subduction zone and the Ninetyeast ridge was an active transform fault between the India and Austral-Antarctic plates (c). By about 50 Ma the ridge on the southern margin of the Pacific plate began to subduct beneath the Austral-Antarctic plate and the Ninetyeast ridge transform became extinct as India collided with Tibet. Between 60 and 80 Ma the Tasman Sea opened, rifting the Lord Howe Rise and New Zealand from southeastern Australia. The Nazca–Pacific ridge migrated rapidly to the southeast and a new plate may have developed in the South Pacific. By 65 Ma only a small amount of the Kula plate

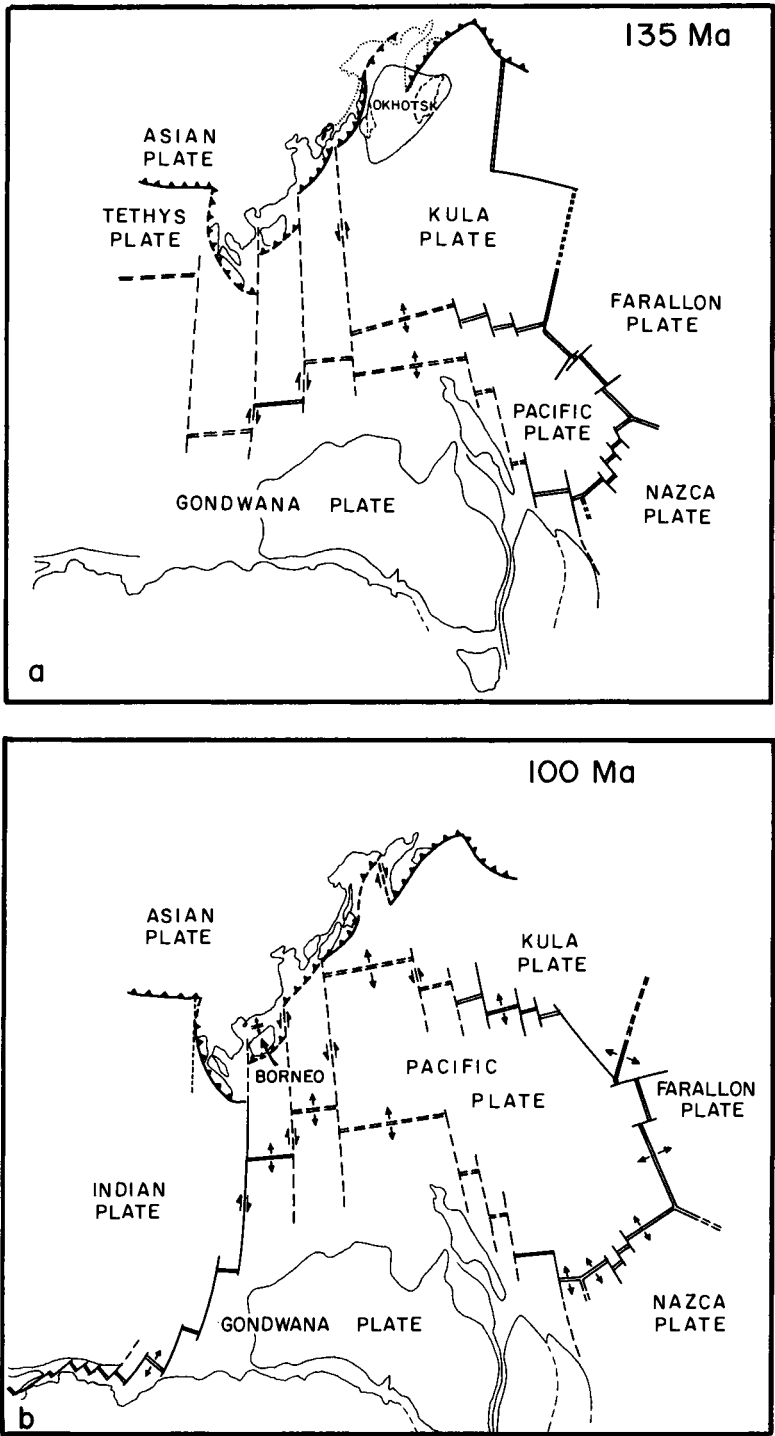
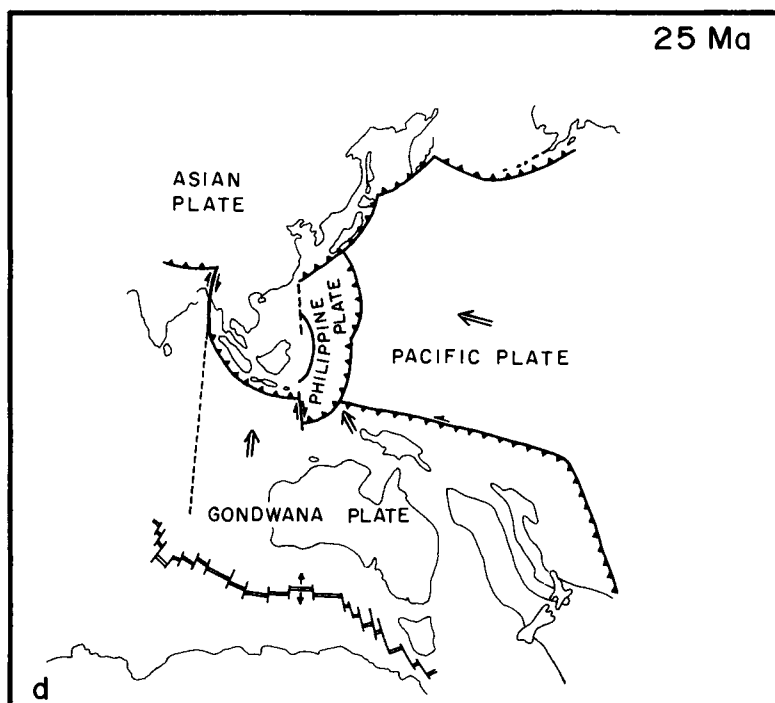
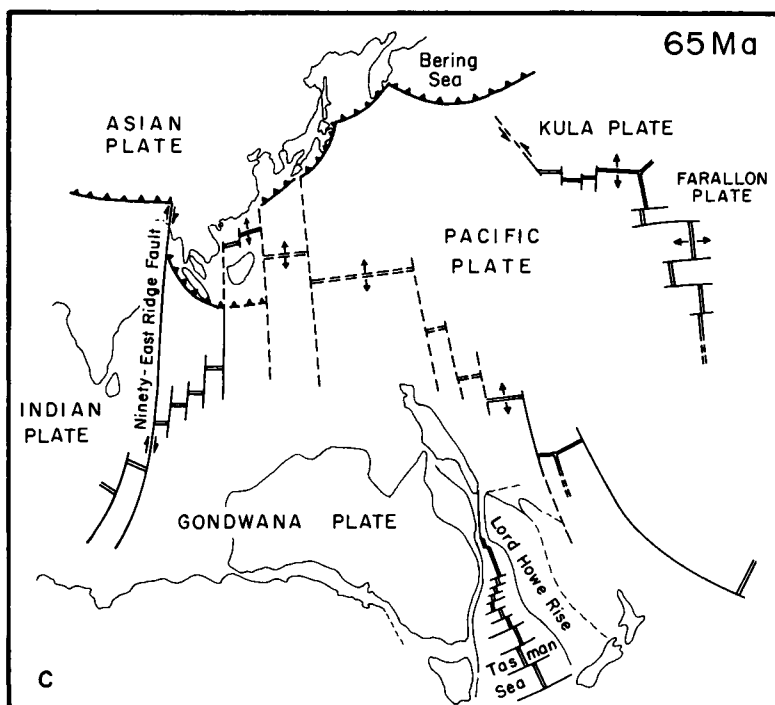


FIG 8.21 Generalized tectonic evolution of the Western Pacific basin during the last 135 Ma. Modified after Hilde et al. (1977).



remained in the northeast Pacific basin. At about 3 Ma, spreading directions in the Pacific plate changed from NNW to WNW, and this change is reflected by the bend in the Emperor–Hawaiian chain (fig. 5.22; Plate I). The ridge system along the southern margin of the Pacific plate was subducted along the northern margin of the Australian plate at about 50 Ma. With the change in spreading direction of the Pacific plate at 43 Ma, the transform faults south of Japan and Korea and south of Tonga became zones of compression, and subduction zones developed. The Philippine, Izu–Bonin, and Tonga–Kermadec arc systems developed along these subduction zones, trapping older oceanic crust in back-arc basins (d). With increased subduction rates at about 25 Ma, back-arc spreading began in some or most of these basins (*viz.* the Philippine Sea) and remnant arc systems were isolated within the basins. Also in the Mid-Tertiary, Australia rifted from Antarctica as the Indian ridge system propagated eastward.

Summary Statements

1. Paleomagnetic and geologic data suggest that a Late Precambrian supercontinent began to fragment by 750 Ma ago forming several major and numerous minor continents. During the Paleozoic the continents collided with each other, producing major orogenic belts in North America and Eurasia, and continued aggregation formed the supercontinent Pangaea by the end of the Permian. During the Mesozoic and Cenozoic, Pangaea was fragmented and the present continents were dispersed. Major collisions occurred, producing the Alpine–Himalayan orogenic belt.
2. The Appalachian–Caledonian orogenic system is characterized by the following simplified history: Late Precambrian opening of the Iapetus Ocean between Africa and North America; development of complex subduction systems along both continental margins of the Iapetus Ocean during the Late Precambrian and Early Paleozoic; the Taconic orogeny (Ordovician) caused by deformation related to closure of back-arc basins; progressive collision of Africa and North America, and North America and Scandinavia, producing the Caledonian–Acadian orogenies (Late Silurian and Devonian); continued convergence, producing thrusts and nappes of the Alleghanian orogeny (Late Carboniferous–Permian); and opening of the Atlantic basin beginning in the Triassic.
3. In Europe the Hercynian orogen developed in response to a complex collision of South Europe with North Europe on the north, and Africa on the south. The continuation of this orogen into the southern United States reflects the South American–African collision with North America. The Urals formed as a result of the European–Kazakhstan collision in the Permian.
4. The Cordilleran orogenic belt in western North America is composed in large part of accretionary terranes added to the continent, chiefly in the Mesozoic and Cenozoic. Some of these traveled >5000 km from their sources in the Southern Hemisphere.
5. Major orogenies in the Cordillera of the western United States reflect terrane collisions (Antler, 300–375 Ma; Sonoma, 200–280 Ma; Nevadan 140–150 Ma), activation orogeny associated with a fold–thrust belt (Sevier, 80–130 Ma) or buoyant subduction (Laramide, 50–80 Ma).

6. During the Mid- to Late Tertiary, core complexes formed in the North American Cordillera in response to mid-crustal extension and heating. The San Andreas fault may have formed during progressive annihilation of the East Pacific rise. In the Late Cenozoic the Basin and Range Province developed opposite the lengthening San Andreas transform system in response to extensional forces and cooling of the North American plate.
7. The Caribbean region began in the Mid-Jurassic with the beginning of opening of the Gulf of Mexico. Small plates were rifted from the southern margin of North America during the Mesozoic to form Yucatan and much of Central America. Major Caribbean Islands formed as part of an eastward-migrating arc system. In the Arctic, northern Alaska and Chukotsk rotated counterclockwise in the Cretaceous, colliding with Siberia, and accretionary terranes were added to form southern Alaska and western Canada in the Early Tertiary. The Andean Cordillera was formed largely by the closing of back-arc basins in the Late Cretaceous and Tertiary.
8. The plate-tectonic history of the Alpine orogenic belt is characterized by a complex and evolving system of ocean ridges, subduction zones, and transform faults that produced many small plates between Africa and Europe as the two continents closed on each other. The major deformation in the form of nappes and thrust faults occurred in the Early and Middle Tertiary in response to the final continent–continent collision.
9. Beginning in the Paleozoic, and largely completed during the Middle Mesozoic, eastern Asia grew by terrane accretion from the south and southeast. Japan was rifted from China during the Miocene. During the last 150 Ma the Pacific plate grew at the expense of the Farallon, Nazca, and Kula plates, the latter of which is nonexistent today.
10. The India–Tibet collision beginning about 50–60 Ma ago was responsible for the Himalayas and the Tibet Plateau. Approximately 2500 km of crustal shortening is recorded across the Himalayan orogen, of which more than half is taken up by strike-slip faulting in SE Asia. During the collision, continental crust was subducted beneath Tibet.

Suggestions for Further Reading

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CHAPTER 9

Precambrian Crustal Provinces

Introduction

A major problem in geology is the role of plate tectonics in the Precambrian. Although many methods of identifying plate boundaries and reconstructing plate motions are available for the last 200 Ma, few are applicable to older crustal provinces. Furthermore, in most Precambrian terranes deformation and metamorphism commonly obliterate original rock characteristics and partially or completely reset isotopic dates and magnetization directions. The most useful constraints on Precambrian tectonic settings are petrotectonic assemblages, broad-scale deformation and metamorphic patterns, and paleomagnetic studies.

There are two extreme schools of thought regarding the role of plate tectonics during the Precambrian. The strictly uniformitarian school suggests that plate-tectonic processes have always operated on the Earth and that the Precambrian can be interpreted in terms of Phanerozoic-type plate tectonics (Burke et al., 1976). The other, a nonuniformitarian view, advocates that the early lithosphere was too hot to be subducted and that thin, perhaps rapidly moving plates were jostled around on the Earth's surface like blocks of ice in the Arctic Ocean. These models call upon vertical or horizontal displacements accompanied by buckling and shearing of plates to accommodate new plate growth.

The cooling of the Earth during the last 4.5 Ga must have been important in governing tectonic processes (Chapt. 10). During the Early Archean the heat production in the Earth was 3–5 times greater than at present (fig. 2.9). Convection would surely have been more vigorous, and the zone of partial melting (i.e. a proto-low-velocity zone) which defines the base of the lithosphere may have been at shallower depths, and thus plates may have been thinner than today.

The simplest way to dissipate the greater heat flow in the Archean is by an increased rate of production and subduction of oceanic lithosphere (Burke and Kidd, 1978). This can be achieved by either faster spreading, increased total ridge length, or both. Today younger lithosphere subducts more slowly than older lithosphere (Abbott and Hoffman, 1984). To dissipate more heat in the Archean, lithosphere may have been younger and hence warmer at the time of subduction, and thus would tend to subduct slowly. This implies that a greater ridge length may have been more important than faster spreading in dissipating the additional heat (Hargraves, 1986). This, of course, does not require that spreading rates have been constant, but that a greater ridge length in the Archean had the major effect on

increased heat loss. Ridge heat loss is proportional to the cube root of ridge length. Thus, for heat flow of three times the present, 27 times as much ridge is required, which suggests that the Earth in the Late Archean was covered by many small plates moving at rates slower than present rates. Another consequence of this model is that angles of subduction would be shallower than today because younger lithosphere subducts at shallow angles. Just what the consequences of these differences are in terms of magma generation, deformation, and metamorphism in the Archean are not yet clear. Some of the differences observed between modern and Archean arc successions, however, may reflect a modified plate tectonic style in the Archean (Abbot and Hoffman, 1984).

Just how subduction occurred in the Archean is not understood. If the major forces were due to negative buoyancy in the lithosphere as in modern plates (Chapt. 5), we immediately encounter the problem that geotherms in subduction zones were probably too steep for eclogite stability prior to the Late Archean (Chapt. 10). One way around this problem is to have an Archean oceanic crust composed chiefly of komatiite. Komatiite has a density (3.2 gm/cm^3) somewhat greater than the upper mantle (3.1 gm/cm^3) and thus would tend to be negatively buoyant and sink (Arndt, 1983). The transition from "komatiite-driven" to "eclogite-driven" plate tectonics may have occurred in the Late Archean.

Archean Provinces

Introduction

Archean crustal provinces are roughly equidimensional in plan view and range from less than 500 km to about 2000 km in maximum dimension. They are composed of rocks ranging in age from about 2.5 to 3.8 Ga. Major Archean provinces of the world are shown in fig. 10.8. The largest preserved province is the Superior Province in eastern Canada (fig. 4.26). North America (including Greenland), in addition, has four more provinces: the Slave, Amitsoq, Nain, and Wyoming Provinces. At least eight provinces are preserved in Africa and several occur in each of the other continents. The original extent of these provinces was probably considerably greater, as evidenced by the abundance of relict Archean dates in younger Precambrian terranes. For example, relict dates from Proterozoic terranes in Canada suggest that the five North American Archean provinces originally may have been part of the same crustal province which was rifted apart, reunited, and in part reworked in the Early Proterozoic (see Chapt. 10).

Most Archean provinces are composed of either or both granite–greenstone or high-grade metamorphic associations (Condie, 1981a). Geochemical studies indicate that granitic rocks (mostly granodiorites and tonalites) and gneiss–migmatite terranes dominate in both associations (Eade and Fahrig, 1971). Structural trends in Archean provinces range from rather continuous to complex and irregular, even within the same geographic area. An example of the complexity of Archean structural trends is shown for the Superior, Slave, and Nain Provinces in North America in figure 4.26. In a few instances, extrapolation of structural trends from one province to another seems to be justified (as for instance between central-west Africa and northeastern Brazil).

Granite–greenstone terranes are characterized by discontinuous greenstone belts

engulfed in a "sea" of granite and gneiss. The major granite-greenstone terranes of the world occur in the Superior and Slave Provinces in North America, in the Zimbabwe and Kaapvaal Provinces in southern Africa, and in the Yilgarn and Pilbara Provinces of Western Australia (Plate I). *Greenstone belts* are linear-to-irregular-shaped volcanic terranes which average 20–100 km wide and extend for distances of several hundred kilometers. They contain thicknesses of supracrustal rocks (chiefly basalts) ranging up to about 20 km (Glikson, 1970; Goodwin, 1968). An idealized map of an Archean granite-greenstone terrane is shown in figure 9.1. Most greenstone belts exhibit a synformal structure with open-to-isoclinal folds and major faults paralleling fold axes. The keel-shaped outcrop pattern so characteristic of some belts results from diapiric granitic intrusions around the edges. Deformational patterns indicate complex polyphase deformation. Rocks within greenstone belts are metamorphosed to the greenschist or amphibolite facies, and primary textures and structures are commonly preserved. Three basic types of granitic rocks are recognized in association with greenstone belts: gneissic complexes, diapiric granitic plutons and discordant granitic plutons. All three types are shown in fig. 9.1. The gneissic complexes and diapiric plutons range from granodiorite to tonalite or trondhjemite in composition, and they are gradational with each other. High-precision U–Pb zircon dating of Archean granite-greenstone terranes indicates that these terranes formed in short periods of time, generally ≤ 50 Ma (Ayres and Thurston, 1985). In some areas more than one volcanic-plutonic cycle may be recorded for a cumulative history of 200–300 Ma.

Studies in the southern part of the Superior Province show that subprovinces with abundant greenstone belts alternate with subprovinces in which metasedimentary rocks (chiefly metagraywacke) dominate (Card and Ciesielski, 1986) (fig. 9.2). Granitic rocks are more abundant than volcanic and sedimentary rocks in both subprovinces, with gneisses and migmatites most abundant in the metasedimentary types. U–Pb zircon dates indicate younging in volcanism and plutonism from the northwest (Sachigo subprovince) to the southeast (Wawa–Abitibi subprovinces). The oldest magmatic events in the northwest occurred at 2.9–3.0, 2.8–2.9, and 2.7–2.75 Ga, followed by major deformation, metamorphism, and plutonism at about 2.7 Ga. In the south, magmatism occurred chiefly between 2.7 and 2.75 Ga, followed by deformation, metamorphism, and plutonism at 2.66–2.7 Ga. The near contemporaneity of magmatic and deformational events along the lengths of the volcanic subprovinces, coupled with structural and geochemical evidence, supports a subduction-dominated tectonic regime in which arcs were successively accreted from northwest to southeast.

Field evidence indicates that most gneissic complexes and diapiric granites are intrusive into greenstone terranes. However, there are some areas in which greenstone volcanic rocks were extruded onto older granite-gneiss basement (Condie, 1981a). In a few greenstone belts, such as Kambalda in southwest Australia, volcanic rocks contain zircon xenocrysts from gneissic basement that is at least 700 Ma older than the host rocks. The fact that gneissosity in Archean gneisses subparallels schistosity in greenstone belts makes identification of intrusive contacts or unconformities in many areas difficult. Some of the gneissic and diapiric granites represent rejuvenated pre-greenstone sialic basement in which daughter isotopes were partially (or completely) remobilized during a later thermal event.

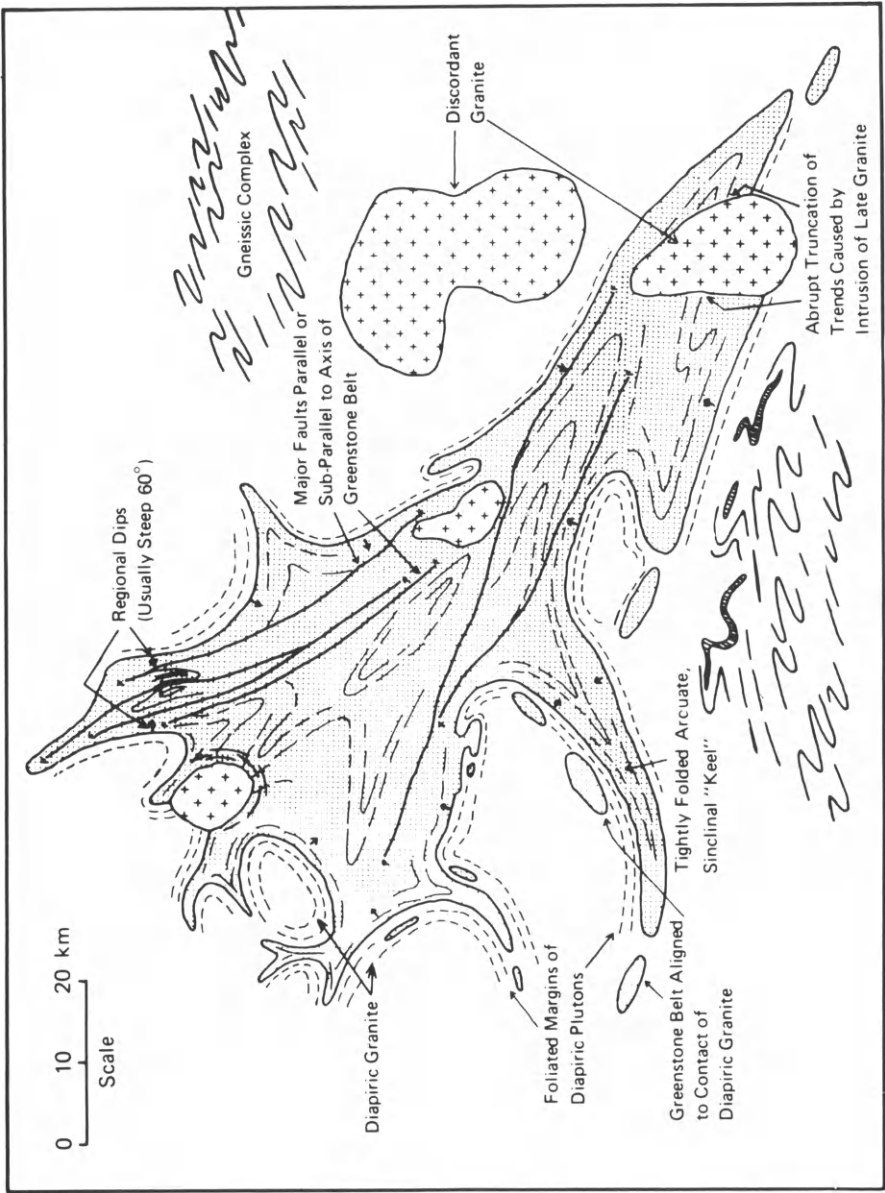


FIG 9.1 Map of an idealized Archean greenstone belt. After Anhaeusser et al. (1969).

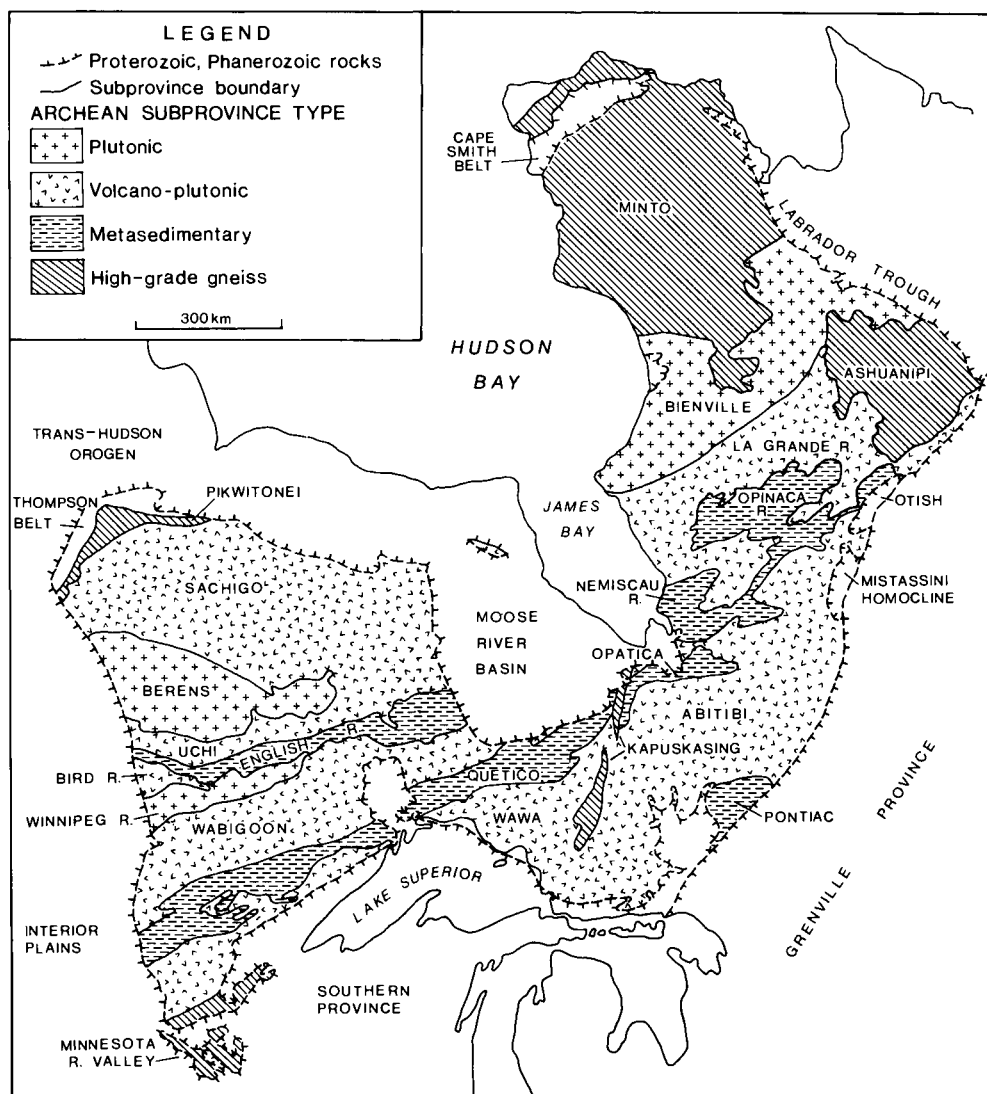


FIG 9.2 Generalized geologic map of the Archean Superior Province, Canada. From Card and Ciesielski (1986). Reproduced with permission of K. Card and the Geological Association of Canada.

Archean *high-grade terranes* are most abundant in Southwest Greenland and in the Archean portions of the Baltic, Ukrainian, Anabar, Aldan, and Indian Shields (Plate I). Archean high-grade terranes are composed chiefly of felsic gneisses and migmatites with minor remnants of supracrustal rocks and layered igneous intrusions. Most have undergone at least one period of granulite-facies metamorphism (650–800°C, 20–35 km burial depth), which appears to represent the culmination of a sequence of igneous, metamorphic, and deformational events (Windley, 1973). The quartzo-feldspathic, gneiss-migmatite complexes formed from 50 to 200 Ma before major granulite-facies metamorphism. Infolded into the gneiss-migmatite

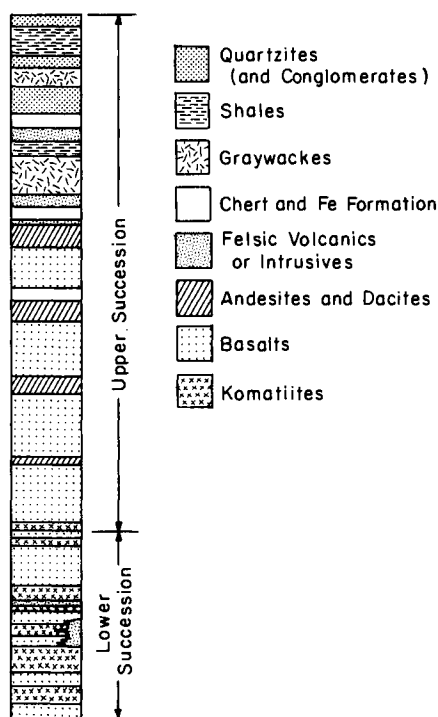


FIG 9.3 Idealized stratigraphic column of an Archean greenstone belt. After Anhaeusser (1971).

terrane are synclinal keels of amphibolites, ultramafic rocks, and metasediments. Although primary structures are rarely preserved in these supracrustal rocks, they are interpreted by some workers as infolded greenstone belts that have undergone higher grades of regional metamorphism. The structural style in Archean high-grade metamorphic terranes is exceedingly complex. It is characterized by fold interference patterns that lack linear regional trends. The close association of folding with the formation of granitic plutons and migmatites suggests that partial melting accompanied deformation. Overall, a high degree of plasticity characterizes the metamorphism and deformation.

One succession of Archean rocks does not fall into greenstone or high-grade categories. It is the Kaapvaal succession in South Africa (Pretorius, 1974), which is composed chiefly of quartzites, shales, and carbonates deposited in shallow marine basins which developed on older (≥ 3.5 Ga) Archean crust. This succession of cratonic and rift sediments includes also major volcanic packages and spans, collectively, about 1000 Ma (3.0–2.0 Ga) of time.

Granite-Greenstone Terranes

Stratigraphy

Archean greenstone successions are structurally and stratigraphically complex. Estimated thicknesses range from 10 to 20 km, but because of shear zones and complex folding, true stratigraphic thicknesses are not well known. An idealized stratigraphic column of a greenstone succession based on studies of the Barberton belt in South Africa is given in figure 9.3 The lower succession is composed chiefly

of subaqueous basalts and komatiites with minor amounts of felsic tuff and layered chert. The lower succession is bimodal in that volcanic rocks of intermediate composition (andesites, dacites) are rare or absent. The upper succession, although generally dominated by subaqueous basalts, contains variable proportions of andesites and felsic volcanics exhibiting tholeiitic or calc-alkaline affinities. Komatiites are rare or absent. Graywackes and volcanoclastic sediments and, in some instances, terrigenous clastic sediments (quartzites, shales) are important in the upper successions. Two general trends observed with increasing stratigraphic height in most Archean greenstone successions are: (1) a decrease in the amount of komatiite, and to a lesser extent basalt; and (2) an increase in the ratio of volcanoclastics to flows and in the relative abundance of andesitic and felsic volcanics. These changes reflect an evolution from widespread submarine eruptions of basalt and komatiite to more localized stratovolcanoes, which may become emergent, and intervening sedimentary basins.

Although it is not yet possible to correlate entire sections between adjacent greenstone belts, the existence of trains of greenstone inclusions in intervening tonalitic gneisses which can be traced from one greenstone belt to another strongly indicates that greenstone belts are remnants of larger Archean basins. Supporting this interpretation is the existence of distinct chert horizons, such as the Manjeri Formation in Zimbabwe greenstones and the Marble Bar Chert in the Pilbara Province in Western Australia, which can be traced or correlated over great distances.

Greenstone belts of more than one age may exist in the same geographic area, commonly separated by an unconformity or a fault. Examples have been described from Zimbabwe, Western Australia, and India (Condie, 1981a). Where two ages of greenstones exist, the older succession is typically bimodal while the younger succession is not. Greenstone belts also exhibit provinciality, in that bimodal types dominate in some crustal provinces such as southern Africa and Western Australia, and tholeiitic or calc-alkaline types dominate in other provinces such as the North American Archean provinces.

Volcanic Rocks

Volcanic and associated hypabyssal rocks dominate in Archean greenstone successions. Although volcanic compositions range from ultramafic to rhyolite, mafic volcanics are most important. Field relationships and volcanic textures and structures such as pillows (fig. 9.4a) indicate that Archean volcanism was largely subaqueous. Many Archean volcanics are highly altered and silicified, probably caused by submarine hydrothermal fluids in a manner similar to that characteristic of modern submarine arcs and ocean ridges (Chapt. 7). Volcanic complexes described from the Superior and Slave Provinces in Canada contain many volcanic centers and range from 100 to 175 km in diameter (Goodwin and Ridler, 1970). Some of these emerged above sea level during the late stages of development, erupting large volumes of felsic subaerial and subaqueous ash-flows. Archean volcanoes were in some respects similar to modern submarine volcanoes in arc systems (Ayres and Thurston, 1985). Similarities include: (1) a general upward change from basalts ($\pm$ komatiites) to calc-alkaline and tholeiitic volcanics, (2) eventual emergence of subaqueous volcanoes to form islands, (3) a linear alignment

(a)



(b)

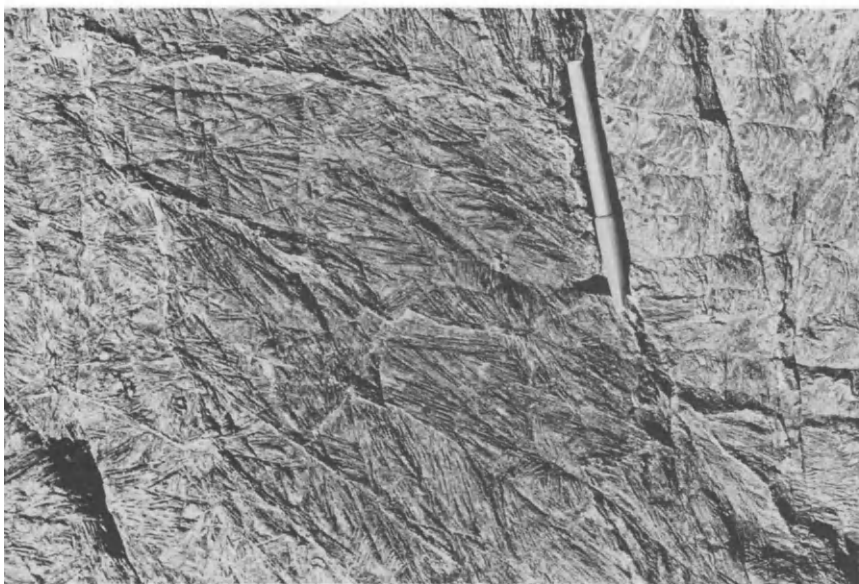


FIG 9.4 (a) Pillow basalt from the Abitibi greenstone belt, Quebec Canada. (b) Spinifex-textured komatiite from Munro Township, Ontario, Canada. Courtesy of C. H. Donaldson.

of volcanoes, and (4) flanking sedimentary aprons leading into basins between volcanoes. Differences between modern arc volcanoes and Archean volcanoes include: (1) the importance of komatiites in many Archean volcanoes, (2) the bimodal character of some Archean volcanics, especially in lower greenstone successions, (3) the paucity of Archean shoshonitic volcanics, and (4) thick, laterally continuous submarine flows which collectively form large submarine plateaus upon which Archean volcanoes grew; these are not found associated with modern submarine volcanoes.

Perhaps the most distinctive volcanic rock of the Archean is komatiite. *Komatiites* are ultramafic lava flows (or hypabyssal intrusives) that exhibit a quench texture known as spinifex texture and contain $\geq 18\%$ MgO. Although komatiites are common in some Archean greenstone successions, they are uncommon in the Proterozoic and very rare in the Phanerozoic. Such a secular distribution probably reflects the falling geothermal gradient in the Earth with time (Chapt. 10). *Spinifex texture* is commonly preserved in the upper parts of komatiite flows. This texture is characterized by randomly oriented skeletal crystals of olivine or pyroxene (fig. 9.4b), and forms by rapid cooling in the near-absence of crystal nuclei. Flows range from those with well-developed spinifex zones that comprise nearly all of the upper half of a flow (fig. 9.5, A) to flows without spinifex textures but with polygonal cooling fractures (C). Komatiite and basalt flows range from about 2 m to >200 m thick, are commonly pillowed, and may be associated with sills or volcanoclastic rocks of similar composition. Intrusive mafic and ultramafic igneous rocks are also found in most greenstone successions, and geochemical studies indicate they are closely related to enclosing volcanic rocks. Archean komatiitic and mafic volcanics are depleted in LIL elements and show variable amounts of light-REE depletion. Most Archean greenstone volcanics exhibit a subduction zone geochemical signature (Chapt. 7) in that they show depletion in Nb, Ta, and Ti relative to REE and LIL elements (Condie, 1988). Such characteristics suggest that Archean greenstones formed in arc systems by processes similar to those operating in modern subduction zones.

Andesites and felsic volcanic rocks in greenstone successions occur chiefly as volcanoclastic rocks and minor flows. Tuffs, breccias, and agglomerates are common, and their distribution can be used to define volcanic centers. Many tuffs have been reworked and exhibit cross-bedding, graded bedding, and other sedimentary structures. Felsic porphyry dikes and plugs are of local importance in some successions. Felsic ash-flow tuffs with well-preserved eutaxitic textures (flattened shards and pumice fragments) have been described from Archean greenstones in Australia and Canada. In terms of most incompatible element distributions, Archean andesites are similar to modern andesites from volcanic arcs. Alkaline volcanic rocks are rare in Archean greenstone successions and, when found, occur chiefly as volcanoclastics and associated hypabyssal intrusives.

Layered Igneous Intrusions

Layered igneous intrusions are found within Archean greenstone successions and within surrounding gneissic complexes. They occur as both post-tectonic, discordant intrusions and as syn- or pre-tectonic deformed intrusions. The largest known Archean body is the Great Dyke in Zimbabwe (2.46 Ga), which has a strike length of

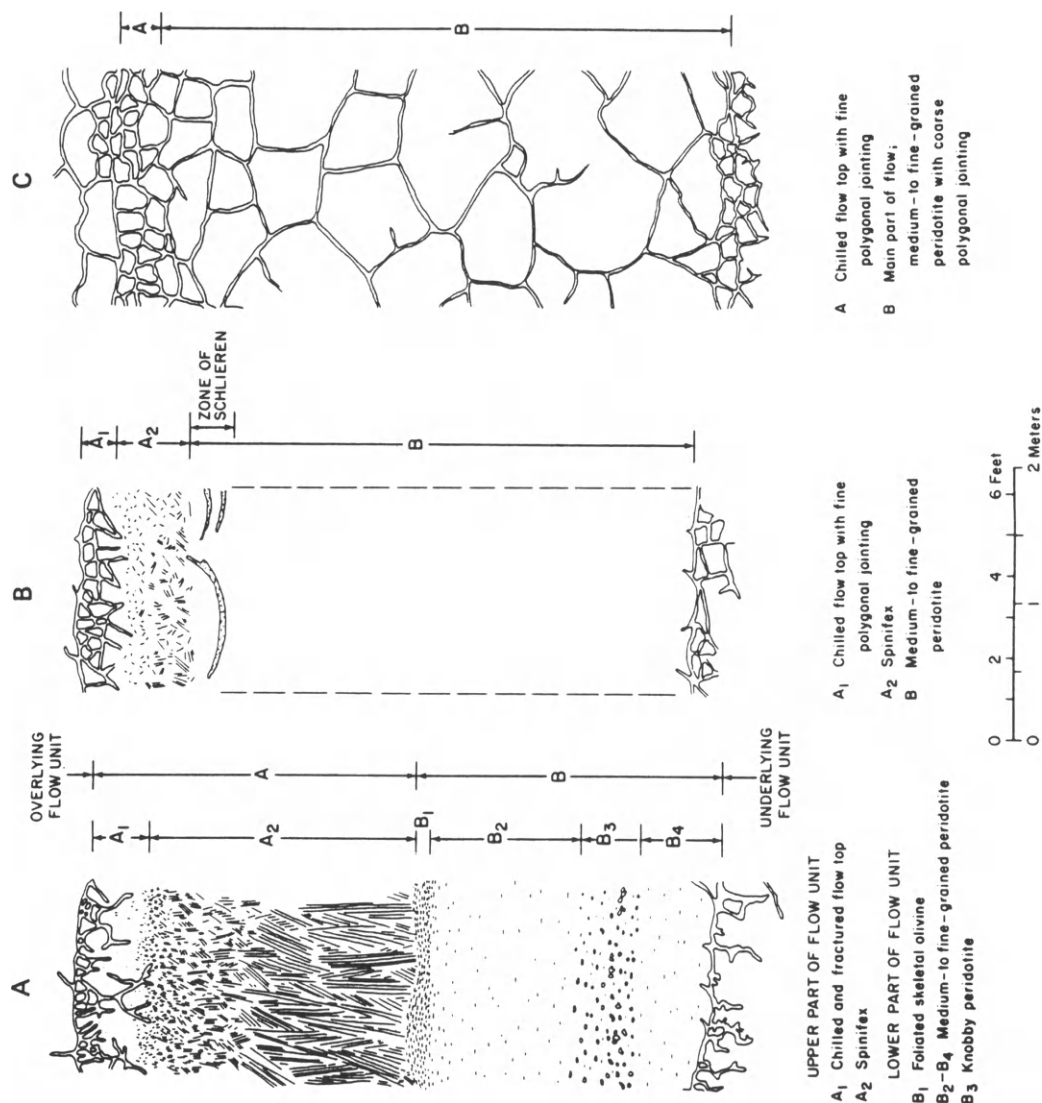


FIG 9.5 Diagrammatic sections through three types of komatiite flows from Munro Township, Canada. From Arndt et al. (1977).

500 km and an average width of 6 km. Stratigraphic thicknesses in large intrusions range from about 5 to 8 km. Geochemical studies indicate an original magma composition of most Archean layered bodies of tholeiite or komatiite. Rhythmic and cryptic layering are characteristic features, and have best documentation in the Stillwater Complex in Montana.

Sedimentary Rocks

Sediments generally comprise between 10 and 30% of Archean greenstone successions. Graywacke and associated pelite are the most important sediments, with quartzite, arkose, conglomerate, and shale of only local significance (Lowe, 1980). Nonclastic sediments, principally chert, are of minor importance but of widespread distribution. Graywacke–pelite successions appear to be, in large part, turbidites, as indicated by the common occurrence of graded bedding and other textures indicative of turbidity-current deposition (Ojakangas, 1985). Archean graywackes have a remarkably constant composition similar to granodiorite.

Greenstone conglomerates are represented by both contact (pebbles touching) and disrupted framework types. The latter is commonly associated with graywackes and appears to have formed during subaqueous slumping. Contact conglomerates, on the other hand, are associated with arkoses and quartzites, and are thought to represent subaqueous alluvial fan deposits formed near rapidly uplifted source areas. Quartzite and arkose are uncommon in most greenstone successions, and when found occur as massive, cross-bedded units. Provenance studies of Archean clastic sediments indicate that volcanic rocks are important in the source areas. Only locally are granite and gneissic sources of importance. In particular, andesitic and felsic volcanic rock fragments are abundant in graywackes and conglomerates.

Layered chert and banded iron formation (BIF) are the most important nonclastic sediments in greenstones. The earliest evidence of life on Earth occurs in the form of microstructures in Archean cherts (Chapt. 11). Relict detrital textures in some cherts, however, indicate that they are chertified clastic sediments or tuffs. Few if any cherts appear to represent pelagic sediments. Most are very local in extent and are probably hydrothermal vent deposits. Low $\delta^{18}\text{O}$ values for many Archean cherts support a volcanogenic origin. Some cherts in Early Archean greenstones are very widespread and contain abundant casts of gypsum crystals (and in some instances, halite) suggesting a shallow-water evaporite origin. Carbonates and barite are very minor sediments in some greenstone belts. The barites appear to represent hydrothermal deposits and the carbonates either hydrothermal or evaporite-related deposits.

Two major environments of sedimentation are recognized in Archean greenstones (Lowe, 1982; Ojakangas, 1985): (1) a shallow-water, relatively low energy environment and (2) an environment characterized by variable water depths associated with active submarine to subaerial volcanoes. The shallow-water environment, which is well represented in some Early Archean greenstones (such as Pilbara and Barberton), is characterized by fine-grained volcanoclastic sediments, biogenic deposits such as carbonaceous shales and carbonates, and chemical precipitates including carbonate, chert, BIF, evaporites, and barites. These rocks, which commonly preserve primary textures (such as mudcracks, oolites, gypsum casts) indicative of shallow-water deposition, are associated with pillowed basalts and komatiites. They

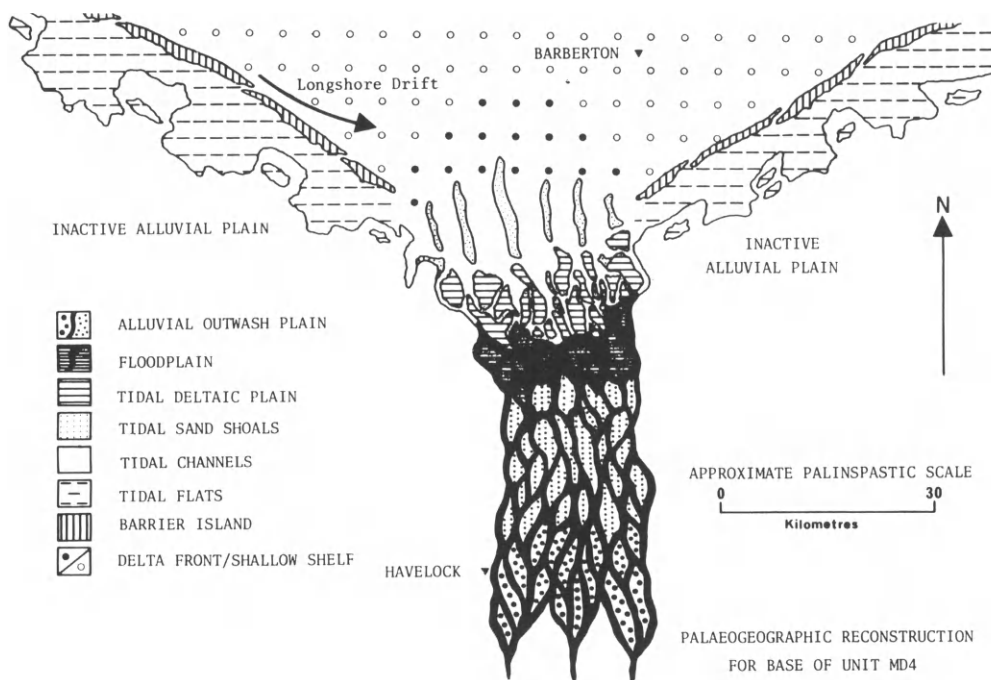


FIG 9.6 Paleoenvironment reconstruction of the Moodies tidal sediments, Barberton greenstone belt, South Africa. From Eriksson (1977).

seem to reflect sources with relatively low relief. The second and most widespread sedimentary environment is characterized by graywackes, pelites, and conglomerates that appear to have been deposited in submarine fans. In some successions, such as the Fig Tree and Moodies Groups in South Africa, deltaic and tidal-flat facies are important (Eriksson, 1977).

A typical sandstone–shale tidalite succession is characterized by three facies representing (from the base upwards) flood-tidal delta, lower-tidal flat, and upper- to mid-tidal flat environments. A reconstruction of the sedimentary environments during the deposition of the Moodies Group shows a southern source area with an alluvial outwash plain followed by a tidal deltaic plain which merges with a submarine delta (fig. 9.6). In comparison to graywacke facies, shelf and deltaic facies are uncommon, indicating a paucity of shallow shelf basins. Graywackes commonly preserve graded bedding and other textures indicative of turbidity current origins in moderate to great water depths. Provenance studies of the turbidite facies indicate a predominance of intermediate to felsic volcanic detritus probably derived from reworking of unconsolidated volcanoclastic material on the slopes of major volcanoes. Subordinate felsic plutonic detritus is derived from local syntectonic plutons or older sialic basement.

Early and Late Archean Greenstones

The existence of at least two sedimentary environments represented by Archean greenstones raises the question of how many tectonic settings exist. The fact that a shallow-water environment is common in Early (≥ 3.5 Ga) Archean greenstones, yet

TABLE 9.1 Comparison of Archean Greenstone Platform and Basinal Environments

	Platform	Basinal
1. Maximum area of exposure	$50 \times 10^3 \text{ km}^2$	$10^3\text{--}10^4 \text{ km}^2$
2. Basement	Not exposed, probably komatiite	Platform greenstones or thinned older sialic basement
3. Tectonic pattern	Equidimensional, with intrusive plutons	Linear, elongated plutons common, transcurrent faults
4. Formation time	100–300+ Ma	$\leq 50 \text{ Ma}$
5. Komatiites	Important	Minor or absent
6. Water depth	Shallow	Variable depths, in part subaerial
7. Sediments	Volcaniclastic, chert, BIF, biogenic shale, carbonate, barite	Chiefly graywacke turbidites
8. Volcanism	Submarine, bimodal	Submarine to locally subaerial, bimodal to tholeiitic or calc-alkaline
9. Mineralization	Minor, Mo–Cu sulfide, barite, BIF	Important, Cu–Ni and Cu–Zn sulfides, Au, Cr, BIF

BIF, banded iron formation.

In part after Groves and Batt (1984).

rare or absent in Late Archean greenstones, supports the possibility that greenstone environments changed with time. The two environments can be characterized as *platform* and *basinal* environments and major differences are summarized in table 9.1. The platform greenstones include the best-known Early Archean examples such as Pilbara (W. Australia), Barberton (S. Africa) and Isua (Greenland). They appear to have formed in shallow water on relatively stable platforms over long periods of time ($\geq 100 \text{ Ma}$). Komatiites comprise about 20–40% of platform successions, while they range from absent to about 10% of basinal greenstones. The nature of the basement upon which platform sequences are deposited is unknown, although it probably was komatiitic. The basinal successions, which dominate in the Late Archean, appear to have been erupted on rifted portions of older greenstones or sialic basement and unconformable relationships are exposed in several areas. In contrast to platform successions, basinal successions formed rapidly ($\leq 50 \text{ Ma}$), contain little if any komatiite, and sediments are chiefly graywacke turbidites.

Structure and Metamorphism

Structural analysis of Archean greenstone belts shows they are considerably more complex than earlier investigators proposed (Schwerdtner et al., 1979; de Wit, 1982; Ayres and Thurston, 1985). It is now recognized that thrust faulting, recumbent folding, and transcurrent faulting are widespread in greenstones. Most successions have undergone two or three periods of major deformation and metamorphism. Greenstones in the Kalgoorlie area of southwestern Australia show repetitions of sections caused by multiple thrust faults. Contacts between stacked units are shear zones characterized by mylonites. In the Michipicoten greenstone in Canada, major recumbent folds facing in opposite directions developed synchronously with major wrench faults. In the Abitibi belt in Canada, wrench faults separate blocks with distinct structural–metamorphic histories that may be allochthonous relative to each

other. Later stages of deformation in greenstones are characterized by vertical forces during which diapiric plutons are intruded. Granitic plutons intrusive into greenstones may contain fabrics of earlier periods of deformation. Results indicate that granite–greenstone terranes are remnants of large basinal or platform sequences that have undergone multiple deformations with compressive forces largely dominating in earlier stages of development changing to vertical forces during intrusion of major plutons.

Metamorphic grade in greenstones ranges from greenschist (or locally zeolite) to amphibolite facies. Higher grades typically occur around margins of greenstone belts, probably caused by intrusion of plutons (Fraser and Heywood, 1978). Metamorphic mineral assemblages are of the low-pressure type (fig. 4.21a) with implied temperature gradients of 30–50°C/km. The major source of heat for later periods of regional metamorphism appears to be from numerous granitic intrusions that surround and probably underlie greenstones.

Granitic Rocks

Granitic rocks in granite–greenstone provinces fall into one of three categories: gneissic complexes and batholiths, diapiric (syntectonic) plutons of variable composition, and late discordant granite plutons. Gneissic complexes and batholiths comprise most of the preserved Archean crust. These terranes are complexly deformed, and range in composition from tonalite to granite. They contain large infolded remnants of supracrustal rocks as well as numerous inclusions. Inclusions increase in abundance near greenstones and occur in various stages of digestion and fragmentation, and trains of inclusions commonly connect greenstone belts. Gneissic complexes contain rocks that range from banded gneisses and migmatites to faintly foliated homogeneous gneisses. Rocks are chiefly tonalitic to trondhjemitic in composition, although granodiorite and granite compositions are of local importance. Contacts between gneissic complexes (and batholiths) and greenstones are generally intrusive, faulted, or strongly deformed. Unconformable relationships are preserved at the base of some greenstone successions. Field, textural, and geochemical relationships indicate that although some Archean gneisses have sedimentary precursors, most have igneous precursors. Gneisses with sedimentary parentage have been described in the southwestern Superior Province, where gneissic layering can be traced along strike into graywacke bedding. Intrusive field relationships, trains of inclusions, and geochemical studies strongly indicate, however, an intrusive origin for most Archean gneissic terranes.

Granitic plutons range from foliated to massive, discordant to concordant, and some are porphyritic. Geophysical studies indicate that most Archean plutons extend to depths less than 15 km. Some plutons, referred to as *diapirs*, have well-developed concordant foliation around their margins and appear to have been deformed during forceful injection. Others, which are post-tectonic, are usually granite (*sensu strictu*) in composition and have discordant contacts and massive interiors (fig. 9.1).

All but the post-tectonic granites are geochemically similar to felsic volcanics in greenstones. Together with their common syntectonic mode of emplacement, this fact suggests that most granitic rocks and felsic volcanics are genetically related. In areas where detailed U–Pb zircon chronology is available, such as in parts of the

Superior Province, entire cycles of volcanism, sedimentation, deformation, and plutonism occur in ≤ 50 Ma. Late discordant granites are generally emplaced during the following 50 Ma. In some Early Archean greenstones (Pilbara, Barberton) plutonism can continue for up to 500 Ma after major volcanism.

Mineral Deposits

Some of the world's major reserves of Cu, Zn, and Au occur in Archean greenstone belts (Condie, 1981a; Groves and Batt, 1984). The platform greenstones, which include most of those ≥ 3.5 Ga in age, generally contain only minor mineral deposits. These include Cu–Mo porphyry and stockwork deposits, and small occurrences of barite and BIF. The Late Archean greenstones contain major Cu–Zn massive sulfides associated with submarine felsic volcanics and Ni–Cu sulfides associated with komatiites. The latter sulfides appear to have formed as cumulates by fractional crystallization of immiscible sulfide melts associated with komatiite magmas. Minor occurrences of BIF occur in most late Archean greenstones. One of the most important deposits in greenstones is Au, which occurs in quartz veins and in late disseminated sulfides commonly associated with hydrothermal chert or carbonate deposits.

High-grade Terranes

Introduction

Compared to granite–greenstone terranes, high-grade Archean terranes are not well known. Major occurrences are found in all Archean provinces, the largest of which are in Antarctica, Siberia, and Southern India. As with greenstones, supracrustal rocks are engulfed with large volumes of granitic rock and both supracrustal rocks and granitic rocks have undergone complex multiple deformation and metamorphic histories (Park and Tarney, 1987). Migmatites are common in high-grade terranes, indicating the onset of partial melting. The major differences between greenstone and high-grade terranes are the supracrustal rocks (chiefly paragneisses and quartzites in the latter) and the granulite–facies metamorphism characteristic of most high-grade terranes. U–Pb zircon and Sm/Nd isotopic studies indicate that high-grade terranes may include substantial amounts of older reworked crust. In the Limpopo belt in South Africa, supracrustals rest with an apparent unconformity on 3.3 Ga basement gneisses. They appear to have been deposited between 3.0 and 3.3 Ga and subsequently underwent a prolonged history of deformation and metamorphism extending to about 2.0 Ga. Unlike most greenstones, many high-grade areas appear to have evolved over several hundred million years.

Supracrustal Rocks

Supracrustal rocks in high-grade areas include large volumes of paragneisses with variable amounts of interlayered quartzite, mafic rocks, carbonate, layered igneous rocks, and BIF. Paragneisses (including various schists and granulites with sedimentary protoliths) comprise more than 50% of most high-grade supracrustal successions. They are often banded and are generally rich in biotite with variable

amounts of hypersthene, cordierite, muscovite, graphite, and sillimanite (or kyanite). In terms of their chemical compositions, including REE patterns, they are very much like Phanerozoic shales from cratonic basins (Condie and Boryta, 1988) and unlike graywackes from greenstones. Geochemical and mineralogical compositions of these rocks suggest shale protoliths. Associated with paragneisses are quartzites which locally preserve cross-bedding. The quartzites appear to be of detrital origin and contain heavy mineral suites, and have trace element characteristics of craton-deposited quartz arenites. BIF is a minor constituent in most high-grade supracrustals, but in NE China it comprises major units where it is mined for iron. Carbonates and associated calc-silicate gneisses comprise minor, but widespread horizons in most high-grade supracrustals. These rocks, which are chiefly dolomites, probably represent shallow marine deposits. Mafic rocks (amphibolites, granulites) occur as dikes and sills, layered intrusions, and perhaps some are metasediments. Minor ultramafic bodies also occur in some successions.

Large layered mafic intrusions occur in some or all high-grade terranes. In general they are intrusive into supracrustals (Windley et al., 1981). They include anorthosites, gabbros, and a variety of ultramafic rocks, all of which may have preserved cumulus textures. Individual intrusions range up to 10 km in thickness and 200 km in strike length, and many preserve original compositional layering. Most evidence suggests that these bodies were intruded into high-grade terranes prior to granulite-facies metamorphism. Unlike layered intrusions in greenstone-granite terranes, the high-grade intrusions often contain igneous hornblende and plagioclase with high An contents (An_{80-100}) and do not show typical cryptic layering. They appear to have crystallized from wet mafic or ultramafic magmas, in contrast to the dry magmas from which layered intrusions in greenstone areas crystallized.

Structure and Metamorphism

Archean high-grade terranes have undergone severe polyphase deformation and metamorphism. Ductile mylonites and augen gneisses are common and indicate major shear zones. Most high-grade terranes have undergone a minimum of three periods of intense isoclinal folding, and some as many as five or six periods (Fripp, 1983). Major compressive forces appear to have dominated in the formation of these terranes.

It is not yet clear whether separate periods of regional metamorphism are common in high-grade rocks or if one prolonged (500–1000 Ma) period of metamorphism is characteristic in a given area. Both prograde and retrograde metamorphic mineral assemblages are found and geobarometry-geothermometry studies indicate temperatures of 700–900°C and burial depths of 25–35 km with some terranes reflecting depths up to 50 km. Detailed studies of Archean crustal sections in southern India and Scotland indicate that granulite-facies metamorphism occurs in the presence of a CO₂-rich fluid phase that purged water from the crust as it moved upward (Condie and Allen, 1984; Newton, 1987). In some areas in India and Sri Lanka the CO₂ wave front has been effectively “frozen” into the rock (fig. 9.7). Breakdown of biotite and loss of water results in removal of such LIL elements as Rb, Cs, U, Th, and Pb from granulites formed at depths of >30 km. The source of the CO₂ is unknown but it may come from the mantle or from the breakdown of carbonates carried to great depths.



FIG 9.7 Incipient granulite formation (gray areas) in tonalitic gneisses from Kabbuldurga in southern India. The granulites formed in response to dehydration by a CO_2 -rich fluid phase derived from greater depths.

Granitic Rocks

Although highly deformed and metamorphosed, granitic rocks in Archean high-grade terranes are similar in composition to those in greenstone terranes. Most are tonalite or granodiorite. They are variably foliated and contain local ductile shear zones. Their intrusive relationship to supracrustals is attested to by exposed contacts and inclusions of supracrustal rocks within the granitic rocks. True granites emplaced at high pressure exhibit concordant, deformed contacts and may be gradational with surrounding tonalitic gneisses (Condie et al., 1986).

Granitic plutons emplaced at lower crustal levels differ from those emplaced at shallow levels by the common presence of positive Eu anomalies and greater concentrations of Ba and Sr. Deep-level plutons appear to be composed of variable amounts of cumulus feldspar, whereas shallow plutons represent largely residual liquids from partial melting or fractional crystallization.

Relationship of Archean High- and Low-grade Terranes

The relationship of Archean high-grade to low-grade terranes (granite–greenstone terranes) is a subject of considerable discussion and controversy. Most models fall into two categories (Condie, 1981a):

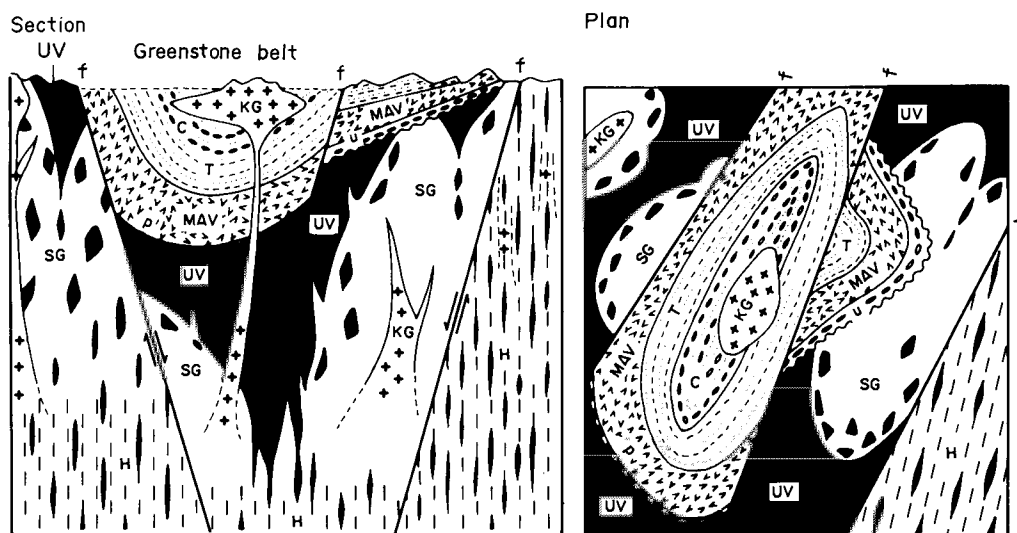


FIG 9.8 Diagrammatic representation of possible depth relationships between Archean granite-greenstone and high-grade terranes (from Glikson, 1976). Symbols: UV = ultramafic-mafic rocks; SG = tonalite-trondhjemite; MAV = mafic to felsic volcanics; T = graywacke-pelite; C = conglomerate, arkose, quartzite; KG = post-tectonic granites; H = high-grade gneisses and granulites; u = unconformity; p = paraconformity; f = fault.

1. high-grade terranes represent the uplifted and eroded root zones of low-grade terranes; or
2. high- and low-grade terranes reflect different tectonic settings.

Radiometric dates from some adjacent high- and low-grade terranes indicate complex but closely related geologic histories. In several areas, such as the Pikwitonei and Kapaskasing subprovinces in the Superior Province in Canada, greenstone belts can be followed from low-grade into high-grade terranes (Percival and Card, 1983). A model dependent upon erosion level relating the two types of terranes is illustrated in figure 9.8. The diagram shows an idealized greenstone belt which, with increasing depth, passes into high-grade gneisses that are equivalent to shallow-level tonalities. These gneisses contain numerous inclusions of greenstone recrystallized at higher metamorphic grade. The high-grade terrane is exposed at the surface by faulting.

Although some high-grade terranes clearly represent the root zones of granite-greenstone terranes, we are faced with three major problems in equating all high-grade terranes to this origin.

1. Most high-grade supracrustal rocks are strikingly different from greenstone successions.
2. In contrast to greenstone terranes, most high-grade terranes have been subjected to long complex histories (>100 Ma) that include, in part, reworking of older continental crust.
3. The metamorphic and deformational styles of the two terranes are quite different, and may not be reconcilable with the same tectonic setting.

For these reasons it appears that many Archean high-grade terranes are not the uplifted root zones of greenstone terranes. They must have formed in different tectonic settings on the Archean Earth.

Origin of Archean Crustal Provinces

Introduction

Many models for the origin and tectonic evolution of the Archean crust have been proposed. Some of the earliest models involve the density-driven inversion of greenstone volcanics and underlying tonalitic crust. Tonalites are “reactivated” and diapirically intrude overlying greenstone belts which settle as synclinal keels between diapirs. One of the first models that attempted to explain both high- and low-grade terranes involved mobile belts (high-grade terranes) that develop over primary convective upcurrents in the mantle, while greenstone belts develop over secondary upwellings. Glikson (1976) has proposed a model for greenstone belts based on evolving oceanic crust. The model involves partial melting of the lower part of an early oceanic crust to produce tonalite–trondhjemite magmas which rise and intrude overlying basalts. This is followed by uplift and erosion and partial melting of the lower tonalitic crust to produce granites.

Perhaps the most unique model for the origin of greenstones is the impact model (Green, 1972a). According to this model, greenstone belts are interpreted as large impact scars, similar to lunar maria, which are initially filled with mafic–ultramafic lavas and are later folded and intruded with tonalites. Two major problems with the impact model, however, have rendered it unlikely. First, most Archean greenstones are younger (chiefly 2.6–2.7 Ga) than mare impact on the Moon (~3.9 Ga), and second, there is an absence of impact textures, structures, and minerals in Archean greenstones.

Platform greenstone successions, which characterize most pre-3.5 Ga greenstones, are perhaps most difficult to characterize in terms of tectonic setting. The broad, shallow-water to locally subaerial environment with relative tectonic stability, yet important submarine volcanism, seems to have no modern counterparts. There is no evidence that platform greenstones developed on or even near older sialic crust. Perhaps they formed on small submarine basaltic plateaus that became emergent from time to time. Because volcanic and plutonic rocks in these greenstone belts carry subduction-zone geochemical signatures it would appear that, if this model is correct, the plateaus must have formed in association with subduction.

Rift Models

Rift models for the origin of Archean greenstone belts have continually been popular with some investigators. Both oceanic and continental rifts have been suggested as sites of greenstone formation (Condie and Hunter, 1976; Groves et al., 1978). Numerous problems are encountered with Archean greenstones formed at ocean ridges. First and foremost is the absence of well-defined ophiolites in greenstone successions. Although a few cases of possible Archean ophiolites have been described (deWit et al., 1986; Helmstaedt et al., 1986), none is very convincing. In

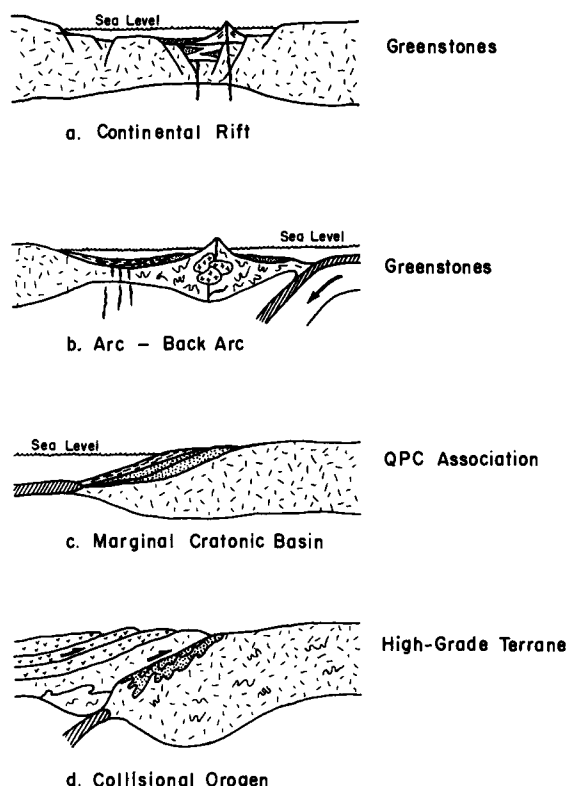


FIG 9.9 Contrasting possible tectonic settings in the Archean. QPC, quartzite–pelite–carbonate.

particular, sheeted dike complexes and tectonized harzburgites have not been documented to exist in any greenstone succession. Furthermore, almost all greenstone basalts exhibit a subduction-zone geochemical signature and less than 5% are geochemically similar to modern ocean ridge basalts (Condie, 1988).

Continental rift models (fig. 9.9a) have met with a greater degree of success (Groves and Batt, 1984; Ayres and Thurston, 1985). Features commonly cited to support continental rifts as sites of greenstone formation are as follows:

1. large volumes of rhyolite in some greenstones;
2. some basinal greenstone successions rest uncomfortably on older continental crust;
3. the bimodal composition of volcanic rocks in some greenstones;
4. linear greenstone basins of the same age that developed over broad geographic areas is suggestive of multiple continental rift systems;
5. some komatiites carry xenocrysts of zircon that are up to 700 Ma older than greenstone volcanism, indicating the presence of older sialic basement.

These features are not unique to continental rifts, however. Large volumes of rhyolite are also found in continental-margin arcs and older continental crust may

also comprise part of the basement in continental-margin arc systems. Although many Archean greenstones are bimodal in terms of volcanic compositions, almost all of these volcanics show subduction-zone geochemical imprints, including depletions in Nb, Ta, and Ti relative to REE (Condie, 1988). Another problem with continental rift models is that they do not provide a mechanism for the production of the vast "seas" of tonalite and granodiorite with arc geochemical signatures that intrude greenstones.

Arc Models

The overall similarity of greenstone successions to modern arc successions has led to a variety of subduction-related models to explain granite–greenstone terranes (Condie, 1981a; Tarney et al., 1976). Features of greenstone belts particularly attractive in terms of arc models are as follows:

1. The overall lithologic assemblage (except for komatiite) in greenstones is similar to that found associated with arc modern systems.
2. The chemical composition of volcanic rocks including incompatible element distributions is similar to that of arc volcanics.
3. Subduction zones provide a means of producing the large volumes of granitic rocks found in granite–greenstone terranes.

Tholeiitic and calc-alkaline volcanic rocks and graywacke–turbidites, which characterize basinal greenstones, are strikingly similar to their modern counterparts in arc systems. Such rocks, which are chiefly of submarine origin in Archean greenstones, may form in back-arc, fore-arc or intra-arc basins (fig. 9.9b). Particularly striking in greenstones volcanics are the depletions in Nb, Ta, and Ti relative to REE, which are characteristic of magmas formed at subduction zones (Chapt. 7). Geochemical studies of Archean tonalites and granodiorites in both greenstone and high-grade terranes indicate they are produced by partial melting of a mafic source with hornblende or/and garnet left in the residue (Condie, 1981a; Martin, 1986). Such a source could be descending mafic crust (as garnet amphibolite) in a subduction zone. Also compatible, although not unique to arc models, are the structural data that indicate major compressive forces during the early stages of greenstone development. These can be interpreted in terms of closure of back-arc basins or by arc collisions. Also compatible with arc models are the linear volcanic and sedimentary belts in the Superior Province (fig. 9.2), which appear to young from the northwest to the southeast. These belts can be interpreted readily in terms of successive arc collisions from the southeast, the net result being the accretion of the Superior microcontinent. As to the origin of bimodal volcanism which does not characterize modern arcs, we may be faced with a sampling bias in the geologic record as described more fully under Proterozoic greenstone assemblages.

Collisional Models

As pointed out above, although some Archean high-grade terranes represent the root zones of greenstones, most probably represent remnants of some other tectonic

setting. High-grade supracrustal successions are similar to those found in cratonic basins. If these successions represent remnants of Archean cratonic basins (fig. 9.9c), how are they buried to 30–40 km depth and then uplifted to the surface? Perhaps collisions of arcs and microcontinents provide the answer (fig. 9.9d). Marginal cratonic basins on leading edges of descending plates could be dragged to considerable depths during collisions where they undergo granulite-facies metamorphism. Subsequent isostatic uplift and erosion remove the overlying arc system (greenstone-granite terrane) and expose the high-grade cratonic terrane on the descending plate. The large volumes of CO₂ accompanying granulite-facies metamorphism may come from the upper mantle. The granitic rocks which intrude high-grade supercrustals also carry subduction-zone signatures and, like the counterparts in greenstone terranes, indicate mafic sources with hornblende or/and garnet left in the residue after melting. In such a collisional model the source of these magmas is not clear. If they come from mafic crust in the descending plate, they should rise into the overriding arc plate before collision. One possible source is mafic rocks in the lower continental crust of the descending plate.

Concluding Statement

Although the specific tectonic environments in which Archean crustal provinces form continues to elude us, there has been a striking convergence toward plate tectonic models. The overall similarities of Archean rock associations to modern ones are, on the whole, more prominent than their differences. Although plates may have been smaller, and ocean ridge lengths may have been greater, than today, most geologists now concur that some form of plate tectonics was in existence by 3.8 Ga.

Proterozoic Provinces

Introduction

Proterozoic crustal provinces are widespread on the continents. Most are stable cratons covered partly or entirely by younger sedimentary or volcanic rocks (Plate I). Some provinces cross-cut older ones at steep angles without offsetting the older provinces. Structural trends within Proterozoic provinces range from linear to complex and swirling (fig. 4.26). Because trends are the result of multiple stages of deformation, careful and detailed structural studies are necessary to interpret their deformational significance. Relict age terranes occur in many orogenic provinces, as for instance in Late Proterozoic mobile belts of southern Africa and in Early Proterozoic provinces of Canada. These terranes, together with widespread relict dates, indicate that much of these provinces is composed of reworked Archean crust. Most of the Proterozoic African provinces less than 2.5 Ga in age appear to represent reworked older crust (Kroner, 1977).

The dominant plutonic rock in most Proterozoic provinces is granodiorite followed by granite. These granitic rocks intrude supercrustal rocks, and may also contain infolded successions of unconformably overlying rocks. Some Proterozoic provinces also contain gabbros, stratiform intrusions, and anorthosites. Mafic dike swarms are common, and are represented by both pre- and post-deformational examples. Phanerozoic-like rock associations become well-established in the Proter-

ozoic, although most (or all) made their first appearances in the Archean. Only the anorogenic granite–anorthosite suite may be unique to the Proterozoic.

Metamorphic grade and facies-series are variable in Proterozoic provinces. The highest-grade zones often accompany extensive migmatite development and occur in the central parts of mobile belts. On the whole, there seems to be a time-dependent trend from low-pressure to high-pressure series in going from Archean to Proterozoic provinces. Provinces that evolve at relatively low temperatures and high pressures are characterized by extensive thrusting and nappe formation, and appear to record substantial shortening of the crust accompanying continental collisions. Examples are the Nagsugtoquidian belt in Greenland and the Pan-African system in Africa and South America. On the other hand, provinces that evolve at relatively high temperatures and low pressures, like the Ketilidian in southern Greenland, are characterized by extensive granitic plutonism and dominantly vertical deformation patterns.

The maximum age of sediments unconformably overlying orogenic provinces can be used to estimate the time involved in uplift and erosion leading to cratonization of an orogenic province (Watson, 1976). Data indicate that uplift and stabilization lasted no more than 200 Ma in provinces older than 700 Ma. Geochronologic studies in Proterozoic provinces indicate variable life spans and complexities of geologic history. Some appear to have formed in <50 Ma with negligible crustal residence times, while mobile belts have undergone multiple periods of deformation, metamorphism, and magmatism remaining episodically active for up to 1000 Ma.

Proterozoic Rock Associations

The Quartzite–Pelite–Carbonate (QPC) Association

The QPC association comprises over 60% of known Proterozoic supracrustal rocks. Representative stratigraphic sections, which attain thicknesses in excess of 10 km, are given in figure 9.10. The overall dominance of quartzite and pelite is evident in the first four sections. Quartzites are typically thick-bedded and massive and commonly show large-scale cross-bedding, and pelites may be recrystallized to phyllites. Stromatolitic dolomite is important in the upper portions of the Kibaran Group and Snowy Range Supergroup and local carbonate-rich horizons occur in the Belt Supergroup. Minor sediments include arkose, conglomerate, tillite and chert. Sandstone provenance studies indicate the importance of granitic–gneiss sources for the QPC associations. Volcanic rocks are of minor importance in most sections and, when found, they are dominantly tholeiitic basalts with minor amounts of felsic or alkaline volcanics. In some successions, like the Animikie Group in Minnesota and the Mt Bruce Supergroup in Western Australia, layered chert and BIF are major rock types. In the Warramunga Group in Australia, subgraywacke is important. Unconformities in QPC successions do not appear to reflect major deformation, but only localized uplift and erosion. Lateral facies changes occur in most successions, which aids in defining basin shapes and source directions. Fluvial, tidal, and shallow marine deposits are most common in this association.

In terms of Phanerozoic occurrences the QPC association is found in three stable tectonic settings: rifted continental margins, cratonic margin of back-arc basins, and intracratonic basins. In rifted continental margin and back-arc basin settings,

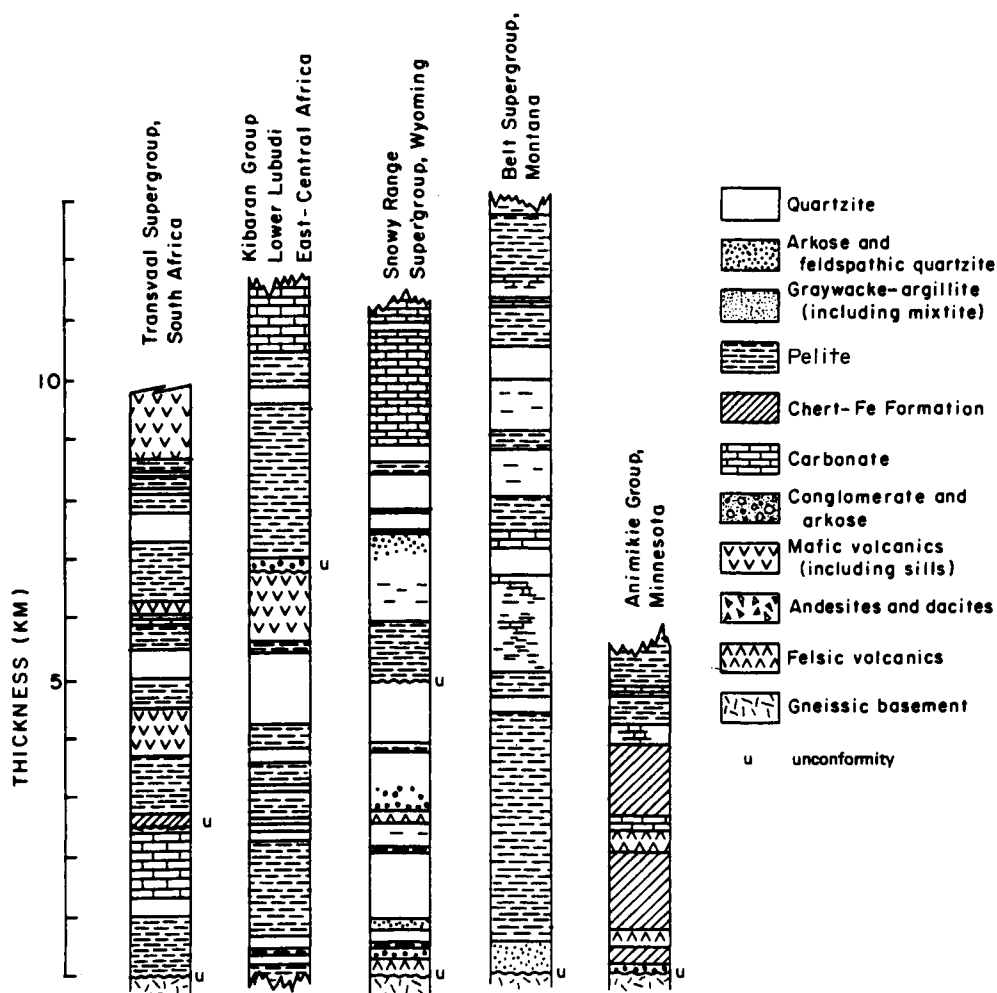


FIG 9.10 Proterozoic stratigraphic sections of the quartzite-pelite-carbonate (QPC) association.

sections thicken toward the ocean, as reflected for instance by the Late Proterozoic Ocoee Series in the southeastern United States. Intracratonic basins are represented by the Athabasca and Thelon basins of Mid-Proterozoic age on the Canadian Shield (Plate I).

One of the most important supracrustal rocks in QPC sections is quartzite. It is of interest to review the possible sources of detrital quartz listed in table 9.2. Mafic and intermediate rocks do not provide adequate sources for quartz; either they do not contain quartz or their quartz is too fine-grained and too minor to serve as a source for quartzites. Derivation of detrital quartz from phenocrysts in felsic volcanics, dikes, and sills requires weathering of vast amounts of volcanic rock, since quartz phenocrysts rarely comprise over 10–20% of such rocks; also, quartz phenocrysts are generally smaller than average grain size of most quartzites. Vein quartz and silicified rocks are very minor in the crust, and hence unlikely to serve as major

TABLE 9.2 Possible Sources of Detrital Quartz

Source Rocks	Comments
1. Basalt, gabbro, nonporphyritic volcanics	Inadequate source
2. Quartz-phenocryst bearing felsic volcanics	Requires extreme weathering and/or selective concentration of quartz; provides only fine-grained quartz
3. Vein quartz	Local source only
4. Silicified source rocks	No evidence
5. Unmetamorphosed chert	Requires polygonization of fine-grained quartz intergrowths; inadequate for monocrystalline and coarse polycrystalline grains
6. Metachert, recycled quartzite	Requires intense weathering; inadequate for monocrystalline and coarse polycrystalline grains
7. Granitic or gneissic terrane	Requires intense weathering or high-energy depositional environment

sources for detrital quartz. Most detrital quartz in Proterozoic quartzites occurs as monocrystalline or polycrystalline grains or as quartzite rock fragments, and hence cannot represent chert. Metachert and recycled quartzite sources also cannot supply the vast amounts of single-crystal quartz found in Proterozoic quartzites. This leaves only granitic or gneissic terranes as major sources for detrital quartz. Such sources may require intense chemical weathering to provide feldspar-free quartz in one cycle of weathering, erosion, and deposition. Alternately, decomposition rates of feldspar may be accelerated in high-energy environments like tidal flats and fluvial systems where relatively pure quartz sands can be produced without intense weathering.

The Bimodal Volcanic–Arkose–Conglomerate (BVAC) Association

The BVAC association is the most diverse of Proterozoic assemblages. Measured sections range from about 5 to greater than 10 km in thickness. Volcanic rocks, which are typically bimodal basalt–rhyolite, dominate in some successions, such as the Sinclair Group in Namibia, while they are minor in others, such as in the East Arm Graben and Pahrump Group in North America (fig. 9.11). Either volcanic end-member may dominate. Basalts are typically tholeiites and occur as both subaqueous and subaerial flows. In some successions, such as the Keweenaw successions in the Great Lakes area, basalts greatly dominate. Felsic volcanics are rhyolites and related rocks and occur chiefly as tuffs. Ash-flow tuffs are particularly common in some successions, such as the Proterozoic successions in the southwestern United States. Sediments are typically immature, terrigenous clastic sediments such as arkose, feldspathic quartzite, and conglomerate, which appear to have been derived from rapidly uplifted terranes of dominantly granitic composition. In some successions, pelites, massive mature quartzites, BIF, and carbonate may be of importance. In the Mt Isa succession in Australia, and the East Arm Graben in Canada, evaporites have been reported. In general, BVAC successions are characterized by rapid facies changes over short distances and mixed subaqueous and subaerial volcanics and sediments. Red beds, reflecting a widespread oxidizing atmosphere, first appear in both the QPC and BVAC assemblages at about 2.4 Ga ago.

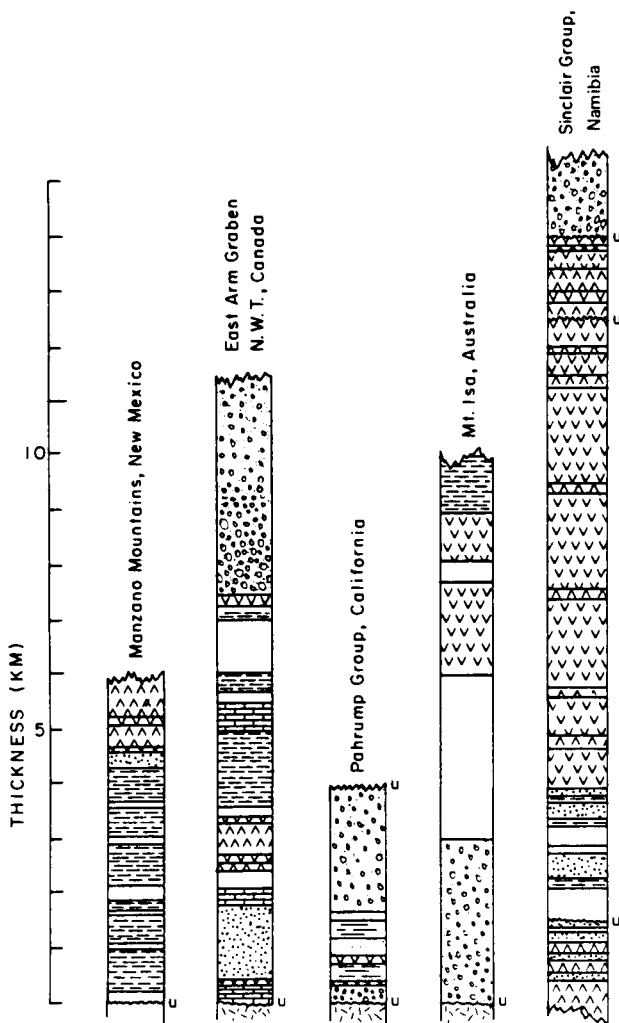


FIG 9.11 Proterozoic stratigraphic sections of the bimodal volcanic-arkose-conglomerate (BVAC) association. Key given in fig. 9.10.

When erosion levels are deep, plutonic rocks related to the BVAC association may be exposed. These are commonly granites (*sensu strictu*) and syenites, and geochemical and isotopic data suggest they represent intrusive equivalents of felsic volcanics. These granitic rocks are typically coarse-grained with uniform textures, and for the most part are post-tectonic.

As described in Chapter 6, modern BVAC associations are typical of continental rifts. Most Proterozoic examples that have been described appear to have formed in cratonic rifts or aulacogens. Geochemical and Sr isotope studies indicate that mafic volcanics from the BVAC association are of mantle origin, and that felsic volcanics and most syn- to post-tectonic granites are of crustal origin. Some Early Proterozoic cratonic rifts appear to be part of large rift systems, as for example the Dewaras and Waterberg successions in southern Africa (1.7–1.8 Ga). Others, such as the

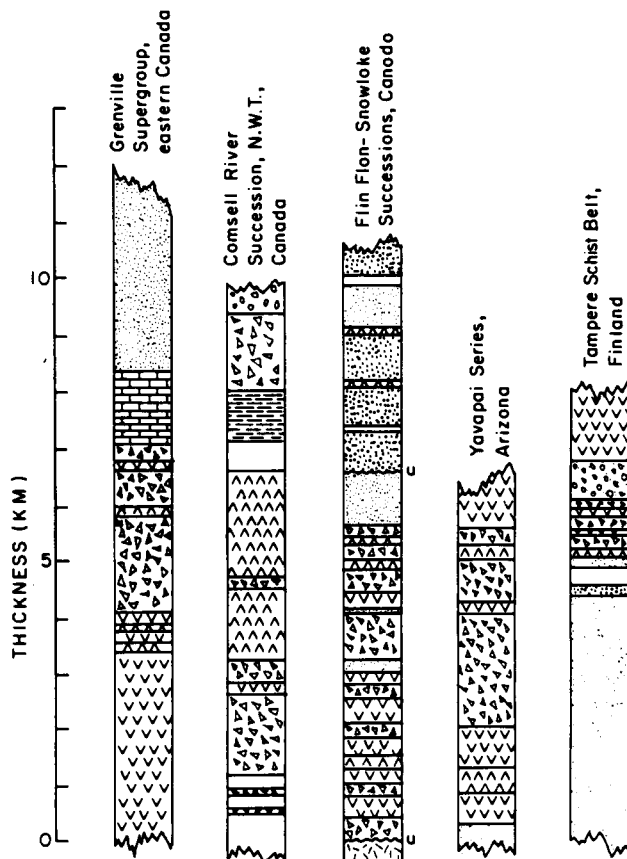


FIG 9.12 Proterozoic stratigraphic sections of the greenstone association. Key given in fig. 9.10.

Pahrump and Uinta Mountain Groups in the western United States, may represent aulacogen successions. Proterozoic rift volcanism and sedimentation are characterized by both subaqueous and subaerial contributions. Sediments represent chiefly fluvial and intertidal deposits, although local marine inundations are recorded by thick carbonate and shale sequences in some successions. The abundance of mature quartzites in some areas, as for instance in the Jotnian successions in Scandinavia and in the Proterozoic successions of New Mexico, may reflect intense winnowing of feldspars in an intertidal environment within the rifts.

The Greenstone Association

With exception of the general absence of komatiites, Proterozoic greenstones are similar to their Archean counterparts. They are characterized by a large proportion of volcanics with both calc-alkaline and tholeiitic affinities (fig. 9.12) (Condie, 1986). Although volcanics typically exhibit subduction-zone geochemical characteristics, they are often bimodal with few if any andesites present.

Basalts are typically pillowed subaqueous flows and associated sills and calc-alkaline volcanics represent a combination of subaqueous and subaerial volcanoclastic rocks. Sections contain variable amounts of graywacke and associated pelite that

exhibit graded bedding and other textures and structures indicative of a turbidite origin. Minor sediments include conglomerate (both matrix- and framework-supported), quartzite, chert, BIF, shale and, rarely, carbonate. Some assemblages, such as the Flin Flon–Snow Lake succession in northern Canada and the Birrimian successions in West Africa, resemble and in some respects are indistinguishable from Archean greenstone successions. Four differences, however, distinguish many Archean from Proterozoic greenstones: (1) komatiitic volcanic rocks are uncommon or absent in Proterozoic successions; (2) volcanoclastic rocks, and in particular potassium-rich felsic volcanic rocks, are more abundant in Proterozoic successions; (3) graywacke is proportionally more important in most Proterozoic successions; and (4) chert and BIF may be less common in Proterozoic successions.

Proterozoic greenstones are intruded by a variety of calc-alkaline plutonic rocks. Most appear to be pre- or syntectonic. Individual plutons vary in size and shape, and some very large batholiths such as the Wathaman batholith in the Trans-Hudson orogen in Canada have been described (Fumerton et al., 1984). Although compositions may range from tonalite to granite, granodiorites greatly dominate in most areas. This is different from typical Archean greenstone–granite terranes where tonalites often dominate. Most Proterozoic granitic rocks also differ from Archean counterparts in that they are not strikingly depleted in heavy REE or Y.

Even more so than Archean greenstones, most Proterozoic greenstones are similar to modern arc successions and in particular to successions from evolved island arcs or continental-margin arcs (Condie, 1986). The major problem in equating Proterozoic greenstones with arcs is the striking bimodal composition of most greenstone volcanics. Although bimodal volcanism is recognized in modern arcs, it is restricted in both space and time. Given volcanic centers may erupt felsic and mafic components at one time and erupt intermediate components at another. When volcanics from a large segment of an arc system are considered collectively, bimodality is generally not apparent.

Perhaps the bimodality in Proterozoic greenstones reflects differences in eruptive style and composition between arcs and back-arc basins with a preservation bias towards back-arc successions. In modern arcs, andesites are produced chiefly as vulcanian-type eruptions and are localized at subaerial stratovolcanoes. Most appear to be the products of fractional crystallization of basaltic magmas that collect in relatively shallow chambers (Gill, 1981). Many felsic volcanics have a similar origin. Basalts are erupted both in arcs and along extensional fractures in back-arc basins. An extensional regime favors rapid eruption and retards fractional crystallization, thus few if any andesites (or more felsic components) form in back-arc basins. The arc proper may not be subjected to large tensional forces and basaltic magmas tend to collect and fractionate, producing andesites and felsic derivatives. Unlike andesites, felsic magmas in arcs undergo plinian eruptions that produce large quantities of airborne pyroclastics. Such pyroclastic material, and especially the finer fraction, settles out in nearby back-arc (and fore-arc) basins, where it is partially reworked by ocean currents. Large amounts of subaerial felsic pyroclastic deposits are also readily eroded and redeposited in adjacent marine basins. Ash-flow tuffs, erupted chiefly at subaerial calderas, may also flow into the sea and travel great distances into adjacent basins. Thus, if most Proterozoic greenstones represent preserved fragments of back-arc basin successions, rather than arcs proper, a

bimodal volcanic suite may be expected. In those few areas where andesites are important, portions of the arc system may have survived. The absence or rarity of andesites in most Proterozoic volcanic terranes may reflect lack of preservation of the arcs, which are susceptible to erosion. It is noteworthy in this respect that most of the Proterozoic greenstones appear to have formed in submarine environments.

Ophiolites

Ophiolites and associated deep-sea sediments are first recognized in terranes about 2 Ga in age. One of the oldest recognized ophiolites that has all the essential components in the correct stratigraphic order (Chapt. 6) is the Jormua Complex in northern Finland (Kontinen, 1987). Although older ophiolites have been reported (deWit et al., 1987; Helmstaedt et al., 1986), most lack a convincing sheeted dike complex and tectonized harzburgites. The thickest most laterally extensive Early Proterozoic ophiolite has recently been described from the Cape Smith Foldbelt in Canada. Although few well-documented ophiolites have been reported that are older than about 1000 Ma, numerous occurrences in the age range 600–1000 Ma are reported from the Pan-African provinces in Africa and South America, and from areas around the North Atlantic. Estimated thicknesses of Late Proterozoic ophiolites range up to 8 km, but most are less than 5 km. They are bounded by thrust faults and appear to have been emplaced by obduction. Although the range in metamorphic grade and degree of deformation is considerable, many Proterozoic ophiolites preserve primary textures and structures. Most, however, represent only partial ophiolite successions.

An example of a complete late Proterozoic ophiolite is the Jabal Ess ophiolite in western Arabia (fig. 9.13). The lower contact of the ophiolite is a melange up to 250 m thick which contains fragments of the ophiolite rock types in a sheared serpentinite matrix. The lowest unit in the ophiolite is a serpentinitized lherzolite with relict cumulus textures, and this is in fault contact with a layered gabbro sequence. Overlying and gradational with the layered gabbro is a sheeted diabase dike complex with individual dikes ranging from 2 to 30 cm wide. Overlying the dike complex is a section ≥ 300 m thick of altered pillow basalts, with a single horizon of shale and laminated chert 50 m thick. A small body of plagiogranite (trondhjemite) is intruded into the section. The ophiolite is folded and has been emplaced by thrusting with a basal melange produced along the low-angle thrust.

As with Phanerozoic ophiolites, most Proterozoic ophiolites carry a weak to strong subduction zone geochemical imprint. This suggests they represent fragments of arc-related oceanic crust from either back-arc or intra-arc basins.

The Anorogenic Granite–Anorthosite Association

A wide belt of Proterozoic granites and associated anorthosites extends from southwestern North America to Labrador, across southern Greenland and into Scandinavia. The granites are massive and undeformed, and appear to have been emplaced in an anorogenic tectonic environment. They typically have rapakivi textures. Granites and anorthosites range in age from about 1000 Ma to 1750 Ma (Anderson, 1983). Most of those in North America are 1400–1500 Ma in age and tend to increase from 1430–1440 in the southwestern United States to 1460–1480 Ma

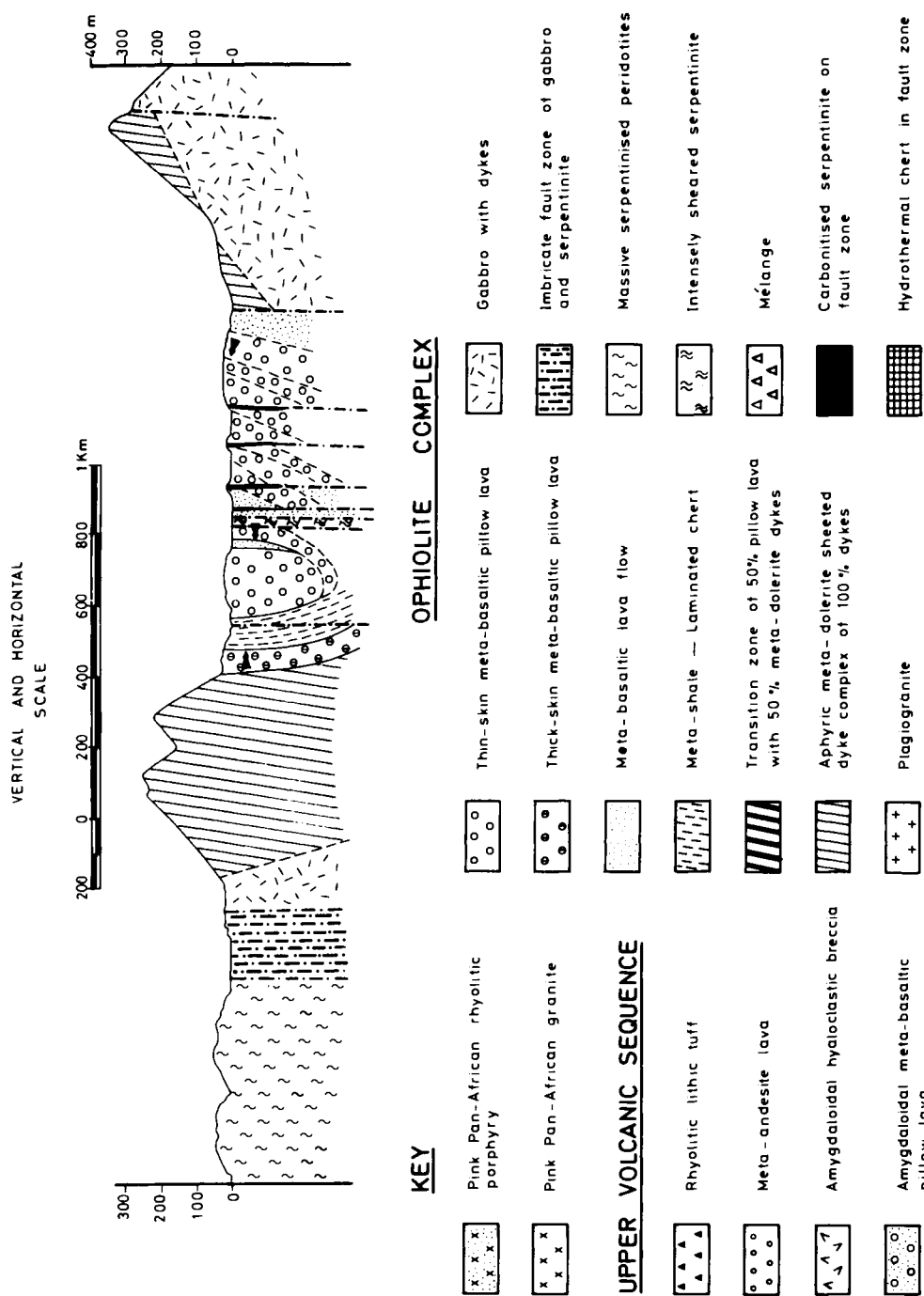


FIG 9.13 Generalized cross-section of the Late Proterozoic Jabal Ess ophiolite, Saudi Arabia. From Shanti and Roobol (1979). (Reproduced by permission of Macmillan Journals, Ltd, copyright 1979.)

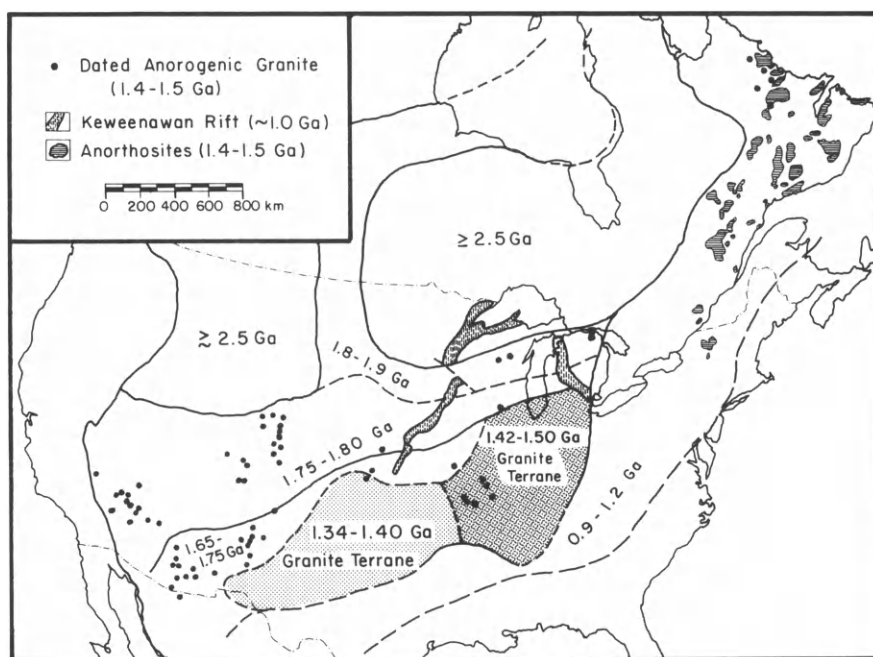


FIG 9.14 Distribution of major Mid-Proterozoic anorogenic granites and anorthosites in North America. After Anderson (1983).

in the mid-continent area. Major subprovinces of 1340–1400 Ma and 1420–1500 Ma granites occur in the mid-continent region (fig. 9.14). The largest and oldest members of the group occur in Finland and date to 1650–1750 Ma. Large anorthosite bodies occur associated with some anorogenic granites, and most occur in the Grenville Province and adjacent areas in eastern Canada.

Proterozoic anorogenic granites are A-type granites enriched in K and Fe and depleted in Ca, Mg, and Sr relative to I- and S-type granitic rocks. They are subalkalic to marginally peraluminous and plot near the minimum in the Q–Ab–Or system at 5–10 kb pressure. These compositions suggest an origin in the lower crust (25–35 km). Relatively high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.705 ± 0.003) are also consistent with a lower crustal origin, as are incompatible-element distributions. The granites appear to have been emplaced under relatively dry conditions at temperatures of 650–800°C and depths of chiefly <15 km. Rapakivi textures may have developed by volatile losses at shallow depths during emplacement. Another characteristic feature of the anorogenic granites is that they crystallized over three orders of magnitude in oxygen fugacities as reflected by their Fe–Ti oxide mineralogy.

Associated anorthosites, which are composed of more than 90% plagioclase (chiefly An_{45-55}) are interlayered with gabbros and norites and exhibit cumulus textures and rhythmic layering. Most bodies range from 10^2 to 10^4 km² in surface area. Many are intruded into older granulite-facies terranes and some are highly fractured. Gravity studies indicate that most bodies are from 2 to 4 km thick and are sheet-like in shape, suggesting they represent portions of stratiform igneous intrusions. The close association of granites and anorthosites suggests a genetic

relationship. Geochemical and isotopic studies, however, indicate that the anorthosites and granites are not derived from the same parent magma by fractional crystallization or from the same source by partial melting. Data are compatible with an origin for the anorthosites as cumulates from fractional crystallization of high- Al_2O_3 tholeiitic magmas produced in the upper mantle (Emslie, 1978). The granitic magmas appear to be produced by partial melting of lower crustal rocks of intermediate or mafic composition (Condie, 1978). The heat for lower crustal melting may be derived from crystallizing basalts in magma chambers in the lower crust. In this respect the anorthosite–granite association is a product of bimodal magmatism.

The tectonic setting of the anorogenic granite–anorthosite suite continues to baffle petrologists. Unlike other Proterozoic rock associations, Phanerozoic counterparts of this suite have not been recognized. Both cratonic rift and convergent-margin models have been proposed. The lack of preservation of supracrustal rocks hinders in characterizing the tectonic setting of the suite. The only well-documented outcrop of coeval supracrustal rocks and anorogenic granite is in the St Francois Mountains of Missouri, where felsic ash-flow tuffs and calderas appear to represent surface expressions of the granitic magmas. Drillcore data suggest that rhyolites are associated with anorogenic granites in the mid-continent area of the United States. The chemical composition, and in particular the incompatible element distributions in anorogenic granites, are suggestive of within-plate tectonic settings. Although our current data base seems to favor an extensional regime for the Proterozoic anorogenic granite–anorthosite suite, we cannot draw a strict parallel to modern (or older Phanerozoic) continental rift systems. It would appear that a unique event (or events) that episodically transcended time from about 1750 to 1000 Ma is responsible for the Proterozoic anorogenic granite–anorthosite association. The nature of this event remains to be determined.

Mafic Dike Swarms

Vast swarms of mafic dikes were intruded into the continents during the Proterozoic (fig. 9.15) (Halls, 1982). Individual dikes range from <1 m to >500 m (averaging 20–40 m) in width and up to 500 km in length. Swarms, which occur on all Precambrian shields, may include many thousands of dikes that range up to 500 km in width and over 3000 km in length. Although some dike swarms were intruded during the Archean, the Proterozoic appears to be the major period of dike emplacement and records widespread extensional forces in the cratons.

Most dikes have rather consistent strikes and dips over great distances. Dips are generally steep, deviating from the vertical by $<20^\circ$. Dike widths tend to be greater where depth of erosion is least, suggesting that dikes widen in the upper crust, and dike spacing ranges generally from 0.5 to 3 km with dikes branching both vertically and along strike. Major swarms appear to have been intruded episodically with important ages of intrusion at 2500, 2370–2390, 2100–2150, 1850–1900, 1250–1270 and 1000–1150 Ma. In terms of composition, dikes are chiefly tholeiites, and from their moderate Mg numbers (50–60) appear to have undergone substantial amounts of fractional crystallization. Most dike swarms older than 2000 Ma exhibit subduction-zone geochemical signatures (Condie et al., 1987). It is only after this

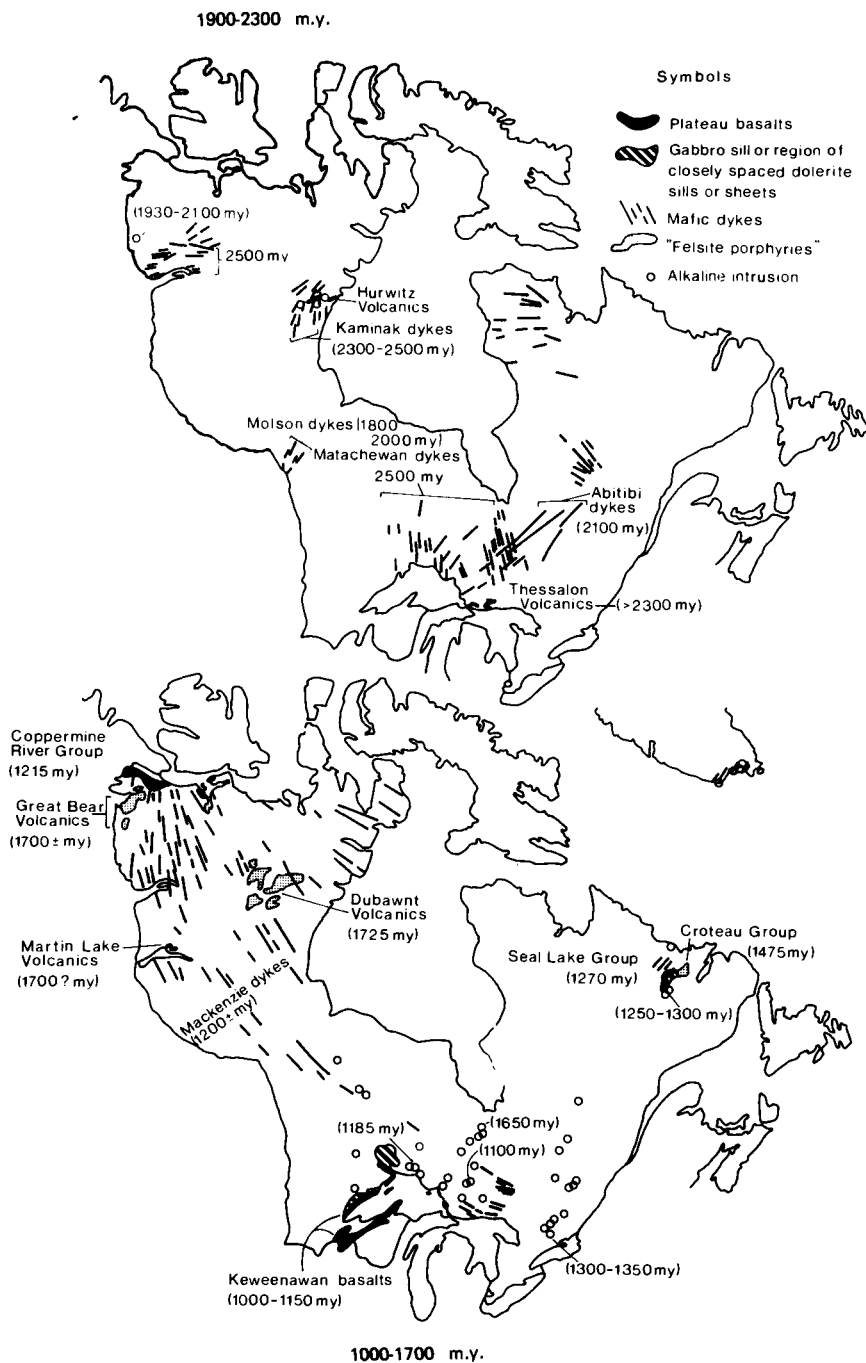


FIG 9.15 Precambrian mafic dike swarms and plateau basalts of the Canadian Shield.
From Baragar (1977).

time, and in particular after 1000 Ma, that within-plate mantle sources became important sites of dike magma generation.

Layered Igneous Intrusions

Layered igneous intrusions are large mafic to ultramafic intrusive bodies that exhibit internal layering resulting from accumulation of crystals during fractional crystallization (Wager and Brown, 1967). Many of these bodies were emplaced during the Proterozoic. The largest is the vast Bushveld Complex in South Africa that is over 8 km thick and covers a minimum area 66,000 km² (Hunter, 1978). It was intruded at about 2000 Ma. Other large bodies with their corresponding ages of intrusion include the Great Dyke in Zimbabwe (2500 Ma), the Widgiemooltha Complex in Western Australia (2400 Ma), the Sudbury Irruptive in Canada (1850 Ma) and the Duluth Complex in Minnesota (1100 Ma). Stratiform complexes are characterized by rhythmic and compositional layering and by cumulus textures. Textural and geochemical data indicate these bodies formed by fractional crystallization of tholeiitic parent magmas. Cyclic layering in the bodies is generally interpreted to reflect episodic injections of new magma into fractionating magma chambers. Large stratiform bodies are major sources of metals, such as Cr, Ni, Cu, and Fe and, in the case of the Bushveld Complex, also Pt. Most layered igneous complexes appear to have been emplaced in extensional environments. They are related either to major continental rifts, such as the Great Dyke and Duluth Complex, or to incipient rifts in cratonic basins, such as the Bushveld Complex.

High-pressure Granulites

Granulite-facies terranes are widespread in Proterozoic provinces (Rudnick et al., 1985). In mobile belts several periods of complex deformation and metamorphism are often recorded. High-P granulites (reflecting depths of >30 km) are also found in Proterozoic mobile belts. Occurrences in which at least one period of granulite-facies metamorphism is Proterozoic in age include the Grenville Province in eastern North America, the Pan-African belts in eastern Africa and South America, the Lapland granulite belt in Scandinavia, the Musgrave and Arunta blocks in Australia, and granulites in Enderby land in Antarctica. These high-P terranes appear to have formed in response to continental collisions, and were brought to the surface by rapid isostatic rebound following collision.

Proterozoic Mineral Deposits

Some of the major occurrences of Cr, Ni, Cu, and Pt are found in Proterozoic layered igneous intrusions. Chromite occurs as primary cumulates within the ultramafic parts of these intrusions, and Cu and Ni generally occur as late-stage hydrothermal replacements. Pt occurs in a great variety of cumulus minerals, the most famous occurrence of which is the Merensky Reef in the Bushveld Complex. In addition, some layered intrusions have magmatic ore deposits of Sn (in late granites) and Ti or V-rich magnetite.

Important Early Proterozoic mineral deposits are the placer deposits of Au and U that occur in quartzites and conglomerates in cratonic successions. The largest deposits are in the Witwatersrand (2800 Ma) and Huronian (2300 Ma) Supergroups (Pretorius, 1976). Detrital Au and uraninite appear to have been concentrated by fluvial and deltaic processes in shallow-water, high-energy environments. Sources for the Au and U are older greenstone-granite terranes. Other important Proterozoic

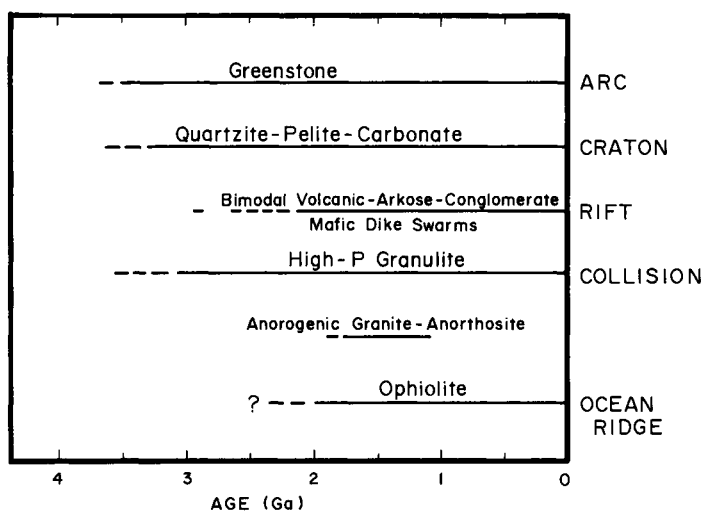


FIG 9.16 Distribution of lithologic associations with geologic time.

sedimentary mineral deposits include banded iron formation (BIF) and manganese-rich sediments. BIF, described more fully in Chapter 11, reached the peak of its development between 2000 and 2600 Ma. It occurs as alternating quartz-rich and magnetite (or hematite)-rich laminae and some was deposited in basins that are hundreds of kilometers across. Most BIF is not associated with volcanic rocks and appears to have been deposited in shallow cratonic basins. Manganese-rich sediments occur also in cratonic successions associated with carbonates. Other Proterozoic mineral deposits include Cu-Pb-Zn massive sulfide deposits, sedimentary Pb-Zn sulfides in carbonates, and U, V, and Cu deposits in red beds. The Cu-Pb-Zn deposits are formed in greenstones, and with exception of the importance of Pb they are like their Archean counterparts. Pb-Zn sedimentary deposits occur in marine carbonates deposited in continental rifts or cratonic basins.

Plate Tectonics in the Proterozoic

Lithologic Associations with Time

The distribution of lithologic associations with geologic time is an important constraint on the role of plate tectonics in the Precambrian. As we have seen, the greenstone, quartzite-pelite-carbonate, mafic dike swarm, layered igneous intrusion and high-P granulite associations have been around since the early Archean (fig. 9.16). However, all but the greenstone association became widespread only during the Proterozoic. Making their first appearances in the Proterozoic are the bimodal volcanic-arkose-conglomerate (BVAC) and ophiolite associations. It is not clear yet, however, whether these first appearances are related to inadequate sampling or they really indicate the first appearances of continental rifts and modern-type oceanic crust. Favoring the first explanation is the fact that ophiolites older than 1000 Ma were unknown before 1987 when several 2000 Ma bodies were described. As we re-examine Precambrian lithologic associations, more similarities to Phanero-

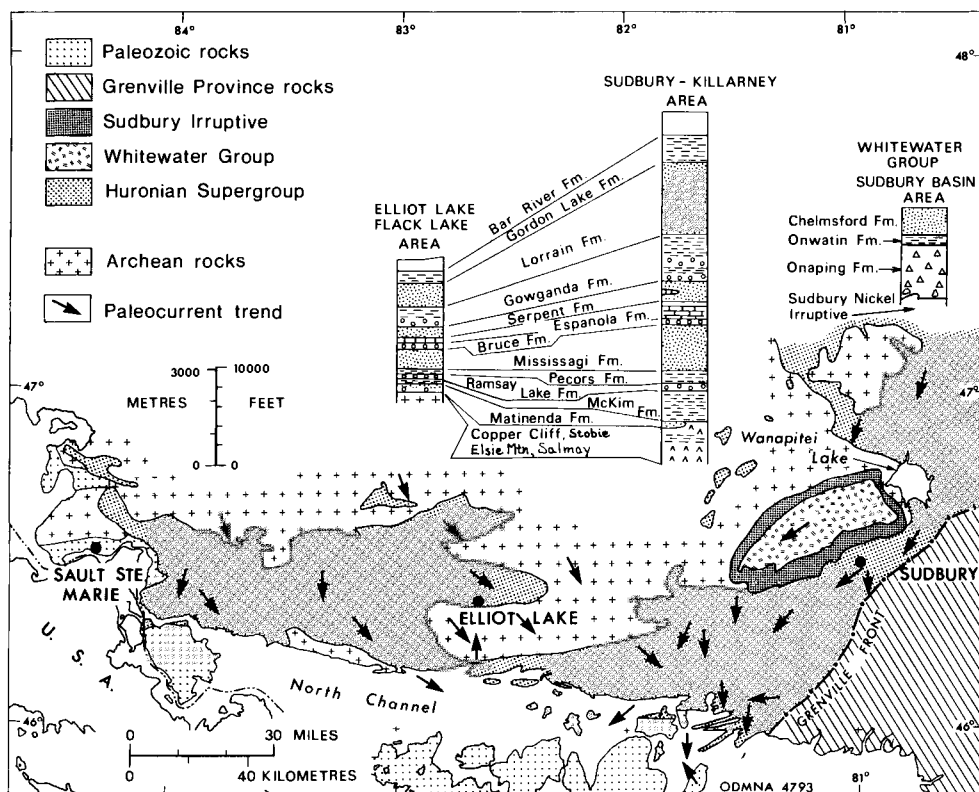


FIG 9.17 Stratigraphy and paleocurrents of the Huronian Supergroup, Ontario, Canada. From Card et al. (1972). Reproduced by permission of K. Card and the Geological Association of Canada.

zoic assemblages become apparent. For instance, accretionary terranes are now described in the Proterozoic. Deformational and metamorphic styles in Precambrian terranes are, on the whole, similar to Phanerozoic orogenic terranes. Differences between the Archean and the Proterozoic, such as the abundance of komatiite in the Archean and the abundance of BIF in the Early Proterozoic, appear to be related either to cooling of the Earth or to changes in the composition of the atmosphere (Chapts 10 and 11). Although the anorogenic granite-anorthosite association presents a particular enigma, on the whole, Proterozoic lithologic associations can be interpreted within a modern plate tectonic framework (fig. 9.16). One cannot, however, equate the number of described assemblages with their areal distribution on the Earth during the Proterozoic. Clearly some associations, such as cratonic associations, may be preferentially preserved while others, such as ophiolites, may be largely destroyed by subduction.

Cratons

Cratonic (QPC) successions were widespread on the continents during the Proterozoic (Condie, 1981b), and some of the best-known successions are from the Canadian Shield. The Huronian Supergroup (2.3–2.4 Ga) in southern Ontario is a succession of coarse clastic sediments up to 12 km thick (fig. 9.17) (Frarey, 1976).

Small amounts of volcanics, carbonates, and pelites also occur in the succession. The sediments are dominated by clean, coarse-grained cross-bedded sands of near-shore marine to fluvial origin. Sources were chiefly to the north in the Archean Superior Province. The succession thickens southward and probably was deposited in a series of marginal cratonic basins along the edge of the Superior Province. Marine tillites occur in part of the succession, indicating continental glaciation in the source areas. Conglomerates in the lower part of the Huronian contain detrital uraninite and pyrite, and suggest an oxygen-poor atmosphere at the time of deposition. The presence of redbeds higher in the succession suggests a change to a more oxygenated environment with time. Deformation is weak in most of the Huronian rocks (gentle warping, some normal faults) and only becomes severe in the south where the succession is involved in the Penokean orogen (1.8–1.9 Ga).

The fact that Early to Middle Proterozoic cratonic basin successions are widespread on the continents indicates that cratons were important by the Early Proterozoic. To accommodate all of the Early Proterozoic QPC basins it would appear that at least 60% of the present continental crust must have been in existence, and must have evolved into stable cratons by 2400 Ma ago.

Arcs

Perhaps the best-documented case for a Proterozoic arc collision is the Wopmay orogen in the Bear Province of northern Canada (fig. 9.18). Hoffman (1980) suggests that at about 2100 Ma the Archean crust in northwestern Canada was rifted apart over several plumes. Thick bimodal volcanics and arkoses are preserved in the East Arm and Kilohigok aulacogens formed during this rifting. The Epworth Basin cratonic succession is deposited along the Wopmay continental margin as a small ocean basin opens. Westward subduction of this oceanic crust (fig. 9.18A) leads to an arc–continent collision at about 1900 Ma (B) (Hoffman and Bowring, 1984). The Hepburn and Wentzel syntectonic batholiths are emplaced during this collision and deformation. The Great Bear plutonic complex is interpreted as part of the overriding plate. In this complex, tonalitic plutons intrude cogenetic calc-alkaline volcanic rocks (LaBine Group). Following the 1900-Ma collision, an east-dipping subduction zone developed, giving rise to calc-alkaline volcanism of the Sloan Group which represents a continental-margin arc (C). Also, the Wopmay fault, a major transcurrent fault, developed in response to oblique subduction. A second deformation, probably resulting from the collision of a Proterozoic microcontinent from the west, is recorded by syntectonic intrusives at about 1870 Ma. Continued collision resulted in transcurrent faulting at about 1820 Ma. The final configuration after the deposition of Middle Proterozoic sediments over most of the Bear Province is shown in figure 9.18D.

Convergent plate margins have also been proposed at about the same time (1.6–1.8 Ga) in the western United States and in Scandinavia. In southeastern Wyoming the Cheyenne shear zone separating the Archean Wyoming Province from Proterozoic provinces in Colorado (fig. 4.31) has been interpreted as a deep-seated manifestation of a suture zone (Hills and Houston, 1979). The volcanic–plutonic terrane south of the shear zone is equated with an arc system that collided with a stable continental margin on the north. Growth of the North American continent

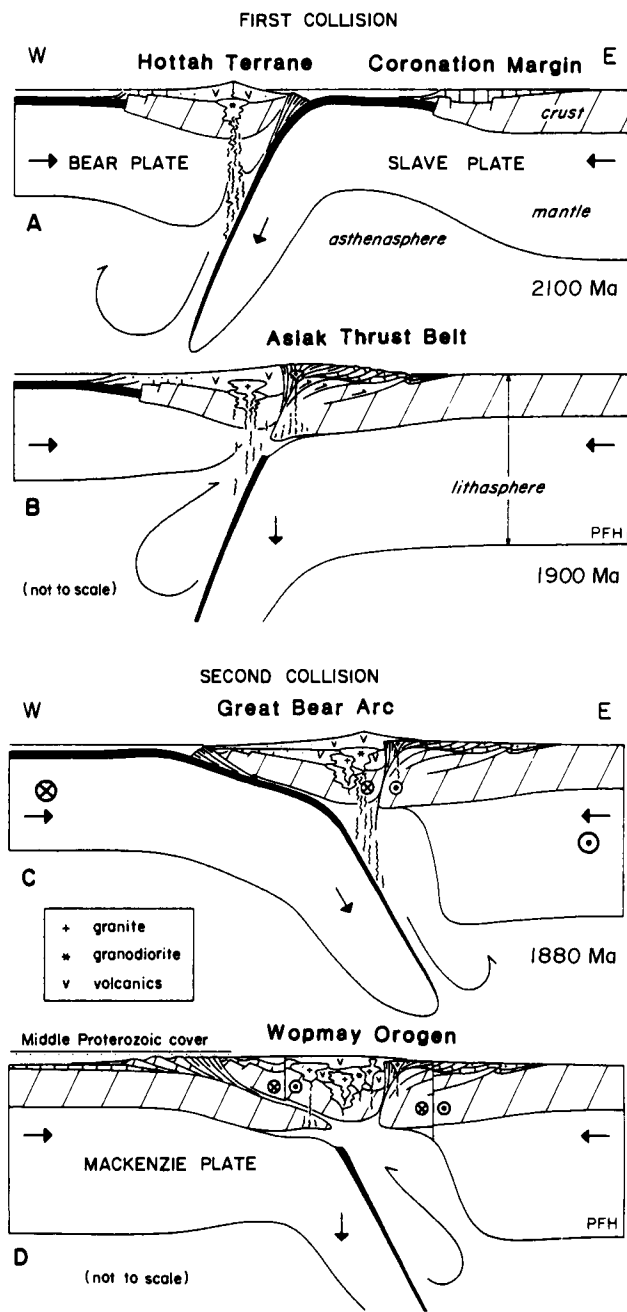


FIG 9.18 Diagrammatic sections showing the evolution of the Wopmay orogen. After Hoffman (1980).

south of the Wyoming province appears to have occurred by successive arc collisions at about 1750, 1710, and 1640 Ma (Condie, 1986). Similar Early Proterozoic arc accretion occurred in Scandinavia as island arcs collided with the Archean Karelia

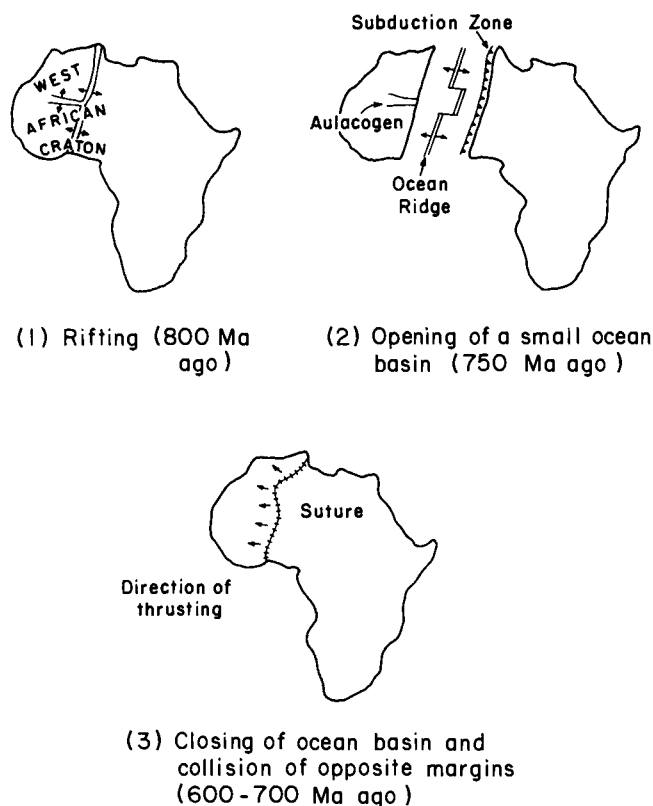


FIG 9.19 Opening and closing of a small ocean basin in the West African craton in the Late Proterozoic.

craton in northern Finland (Park, 1985). Extensive back-arc deposits are described in both the southwestern United States and in Scandinavia. Similar, although less known, Early Proterozoic arc terranes occur in the Birrimian of West Africa (~ 2.0 Ga) and rocks of similar age in the Guiana Shield in South America.

Remnants of island arc systems are widespread in Late Proterozoic terranes in northeastern Africa, Arabia, and in areas around the North Atlantic. The earliest well-documented case of a Wilson cycle is one recorded in West Africa in the Dahomeyan-Pharusian belt. The history of this area can be interpreted in terms of the rifting of the West Africa craton at about 800 Ma to form a small ocean basin (fig. 9.19) (Bertrand and Caby, 1978). Paleomagnetic data will not allow opening of the ocean by more than a few hundred kilometers. A continental-margin arc develops along the eastern margin of the basin by 700 Ma as the basin begins to close. Continental collision at 650–700 Ma is reflected by nappes and thrusts directed to the west, a suture zone with ophiolites, and emplacement of numerous granites of lower crustal origin. The northeastern margin of the Sahara craton underwent a different history, involving several to many island arc systems which became welded to the African craton through collisions between 600 and 1000 Ma. Rocks older than 1000 Ma are not recognized in Arabia, suggesting that Arabia was added to Africa by arc-continent collisions between 1000 and 600 Ma ago.

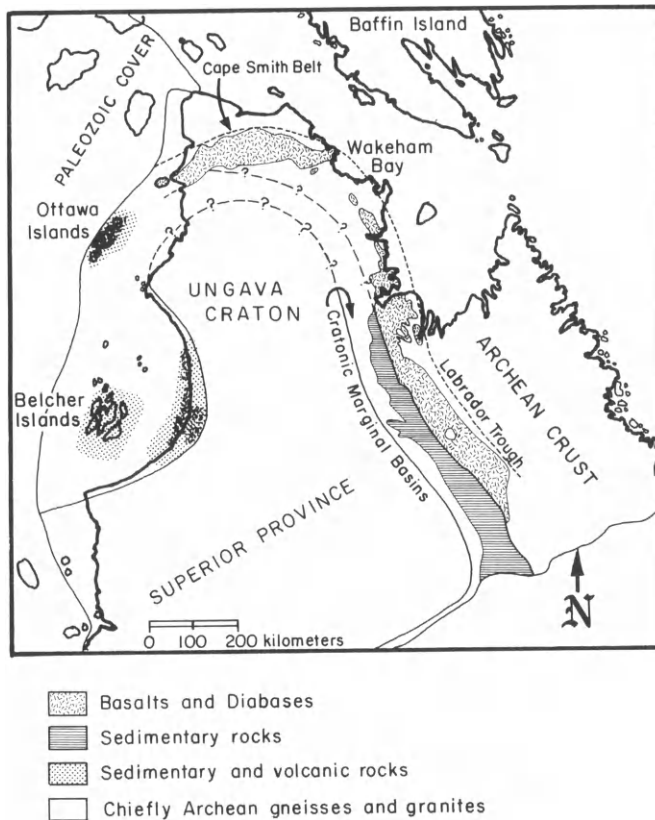


FIG 9.20 Map of the Early Proterozoic Circum-Ungava Province, northeastern Canada. Modified after Dimroth et al. (1970).

Late Proterozoic volcanic rocks in Newfoundland and other areas around the North Atlantic are remnants of a Late Proterozoic arc system and obducted ophiolites (Strong, 1979). The abundance of the BVAC association in the Burin Peninsula and Avalon platform of eastern Newfoundland reflect cratonic rift systems between 600 and 800 Ma. These rifts eventually opened into the Iapetus Ocean basin between 500 and 700 Ma (Chapt. 8). Remnants of back-arc basins associated with these openings are preserved as obducted ophiolites.

Rifts

Although not becoming widespread until about 1000 Ma (Sawkins, 1976), major continental rifting is recorded in the geologic record beginning at about 2000–2300 Ma. One of the earliest documented studies of a cratonic rift system is the Circum-Ungava system in Canada (fig. 9.20). The Circum-Ungava system is composed of the Labrador Trough, the Cape Smith Foldbelt, and the Belcher Foldbelt, which surround the Ungava Craton on the three sides in northern Quebec. Additional portions of the system may extend west along the southern margin of Hudson Bay. The supracrustal rocks are chiefly a QPC succession along the Ungava

craton margin, while mafic volcanics and sills predominate away from this margin. Cyclic sedimentation and volcanism characterize the QPC successions which appear to have been deposited on Archean sialic crust. Rocks in the Labrador Trough begin with a BVAC assemblage and evolve into a QPC assemblage on the west and into a thick succession of basalt, gabbro, and shale on the east. These assemblages are interpreted to reflect a cratonic rift which opens into a small ocean basin (as represented by the mafic rocks and shale) with marginal cratonic basins on the west. A thick obducted ophiolite is well-preserved in the Cape Smith Foldbelt. The entire succession is later deformed and metamorphosed, with major overthrusting and nappes directed toward the Ungava craton.

One model for the development of the Proterozoic Circum-Ungava system involves thinning of the lithosphere and crust above mantle upwellings, producing rift-like basins into which sediments are deposited at 1.9–2.0 Ga. The sediments are derived primarily from the Ungava craton. Extensive heating in the upper mantle produces large volumes of tholeiites which are erupted and intruded in the outer part of the U-shaped basin system. Subsidence of mantle upwellings leads to compression of the basins, and at about 1.8 Ga, supracrustal rocks are deformed and thrust towards the cratonic interior. Minor subduction also appears to have occurred during closing of the rift. The Cape Smith Foldbelt is a large klippe that was emplaced during thrusting of oceanic crust onto the Archean craton.

Perhaps the best-known example of a Precambrian rift system is the Keweenaw rift, which developed between 1.0 and 1.2 Ga in central North America. The mid-continent gravity high (fig. 4.17) delineates this rift system where it is covered with Phanerozoic cratonic sediments. When exposed in the Lake Superior area, the rift contains up to 15 km of basalts and red clastic sediments, filling a trough about 150 km wide and over 1500 km long (Wallace, 1981). The trough is a fault-bounded rift zone which extends as far southwest as central Kansas. Most sections are composed of more than 70% tholeiitic lavas, with the remainder dominantly felsic volcanics, arkose, shale, and conglomerate. Individual basaltic eruptive sequences range up to 7 km thick and are of the order of 33,000 km² in areal extent. Basalts were fed by fissures, examples of which are extensive dike swarms in the Lake Superior area. U–Pb zircon dates suggest the rift was active over a relatively short period of time between 1120 and 1170 Ma.

A summary of the tectonic history of the Keweenaw rift in the Lake Superior area is given in figure 9.21. The rifting cycle began in the stable craton by injection of dikes parallel to the axis of an incipient rift (a). Later these dikes were conduits for basaltic magma to reach the surface. Many hundreds of flows were erupted, all exhibiting reverse magnetization. Sediments were deposited in the rift and basaltic volcanism continued with eruption of normally magnetized basalts (b). Regional subsidence occurred as large volumes of basalt continued to be erupted into the rift, and the stress regime changed from extension to compression and high-angle reverse faults developed along the margins. Vertical movement of 2–4 km formed a central horst that supplied sediments to flanking basins (c). Sedimentation continued in the rift basin well into the Cambrian when subsidence came to an end.

The period from 800 to 1000 Ma was a time of widespread aborted rifting in the continents. Many rifts of this age are known in North America and eastern Europe. Few of these rifts, however, appear to have fragmented continental crust. The best-

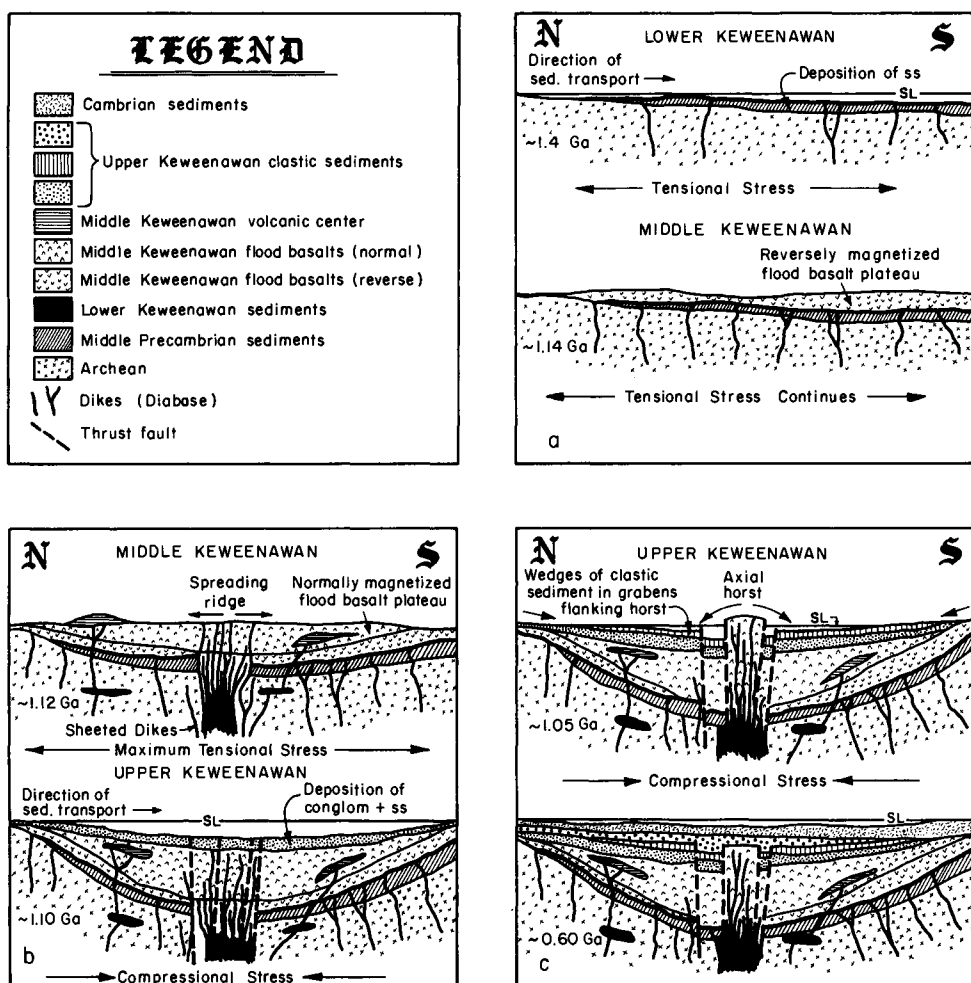


FIG 9.21 Diagrammatic cross-sections showing the evolution of the Keweenaw rift system. From Wallace (1981).

documented example of such fragmentation is the opening of the Iapetus Ocean in the North Atlantic area, and possibly the opening of a proto-Pacific basin as a landmass was rifted away from the western edge of North America. Earlier periods of aborted continental rifting may be recorded in Australia and Africa. Numerous remnants of 1.7–2.0-Ga-old BVAC assemblages are preserved in southern and eastern Africa, and may represent remnants of one or more Early Proterozoic rifts. The fact that many of these lie within the bounds of the present East African rift system suggests they may have been ancestral to this rift system.

Some of the most perplexing lithologic assemblages occur in large linear basins of Early to Middle Proterozoic age in Australia. The rock associations in these basins include primarily sediments (shales, sandstones) with variable amounts of bimodal volcanics that appear to have been deposited on older Precambrian crust. Do these represent large aborted rift systems? It is interesting in this regard that few if any

Proterozoic arc successions have been described from Australia. Perhaps North America and Scandinavia, which show evidence for major arc accretion during the Proterozoic, were different from Australia. Australia may have been subjected chiefly to tensional forces which led to rifting, and in some instances, to fragmentation of micro-continents, whereas North America and western Europe were the sites of subduction and arc collision during this time.

Dikes

Just where do dike swarms fit into Proterozoic tectonic regimes? Dike swarms appear to form in association with major rifting events. During the early stages of rifting, dikes develop radial patterns from point sources (potential triple junctions) and are generally parallel to the rift axis (fig. 9.22). During the spreading stage, an ocean basin may open between two or more of the dike directions and dike intrusion ceases as magmatism is concentrated along the ocean ridge. Later, during closure of the ocean basin and continental collision, earlier dikes are consumed or highly deformed. The swarms intruded in the failed arms are more likely to be preserved if they are on the descending plate. Hence, major dikes that appear to radiate from a point, such as the Mackenzie swarm (1220 Ma) in Canada (fig. 9.15), may indicate the presence of a former ocean basin at approximately right angles to the strike of the dikes.

Proterozoic Collisional Orogens

Reactivation

Large segments of the Proterozoic continental crust represent reactivated (or re-worked) older crust. Suggestive that reactivation — which includes deformation, metamorphism, and partial melting — is more widespread in the Proterozoic than in the Archean is the fact that initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are higher and more variable in granitic rocks from Proterozoic and Phanerozoic terranes than from Archean terranes (fig. 9.23). Initial Nd and Pb isotope ratios, as well as incompatible element distributions and oxygen isotopes, also support this conclusion. The relatively sudden increase in the volume of reactivated crust in the Proterozoic seems to reflect chiefly the rapid growth of continental crust in the Late Archean (Chapt. 10). Prior to the Late Archean there was very little continent to reactivate.

Much has been written about possible tectonic regimes accompanying reactivation, and many *ad hoc* plate tectonic models have been proposed (Wynne-Edwards, 1976; Kröner, 1981). However, there is increasing evidence that reactivation results from continental collisions. Large segments of Phanerozoic crust appear to have been reactivated by such collisions. For instance, much of central Asia, at least as far north as the Baikal rift, was affected by the India–Tibet collision beginning about 50 Ma ago. Widespread faulting and magmatism at present crustal levels suggest that deeper crustal levels may be extensively reactivated during such a collision. In Phanerozoic collisional orogens where deeper crustal levels are exposed, such as the Appalachian and Hercynian orogens (Chapt. 8), there is isotopic evidence for widespread reactivation.

One of the obstacles cited for Proterozoic collisional orogens is the lack of suture

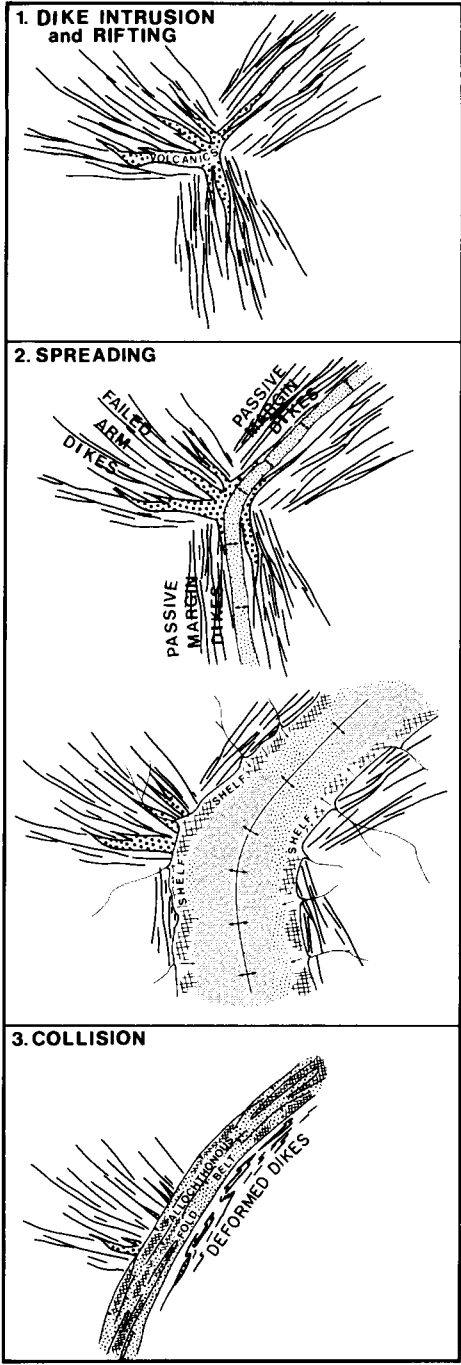


FIG 9.22 Three stages in the tectonic evolution of mafic dike swarms.

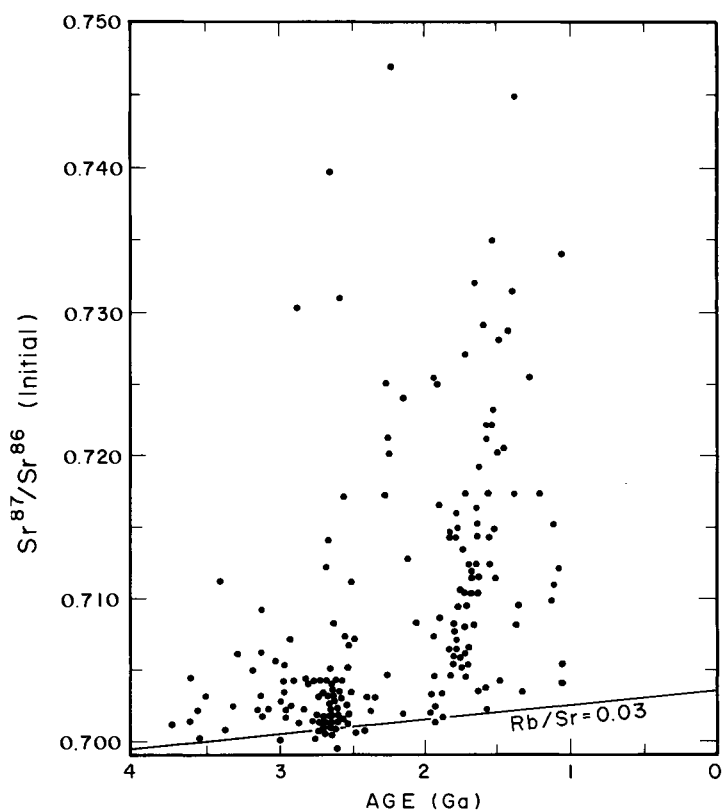


FIG 9.23 Distribution with time of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in granitic rocks. Note the increase in this ratio beginning about 2 Ga ago.

zones. However, at the crustal depths exposed in most reactivated Proterozoic terranes, suture zones are difficult to identify. The classic example of a Cenozoic suture is the Indus suture in the Himalayas. In the Nanga Parbat area, where deep levels of this suture are exposed, it is difficult to define the suture (Coward et al., 1982). Rocks on both sides of the suture are complexly deformed amphibolites and gneisses indistinguishable from each other. Numerous ductile shear zones are also common in reactivated Proterozoic terranes. If we compare equivalent crustal levels of the Indus suture with Proterozoic orogens, there are striking similarities, and deep-seated sutures may very well escape identification.

In terms of our present understanding of Phanerozoic tectonic regimes it does not seem necessary to create new tectonic models for extensively reactivated Proterozoic provinces. These provinces appear to represent complex collisional orogens. From detailed radiometric dating using various isotopic systems, it is now possible to begin to reconstruct the complex histories of reactivated crustal provinces. As illustrative, the following is a brief summary of the histories of two Proterozoic mobile belts.

The Riddle of Two Mobile Belts

The Limpopo mobile belt is composed of highly deformed and metamorphosed rocks (chiefly gneisses and granitic rocks) lying between the Archean Kaapvaal and

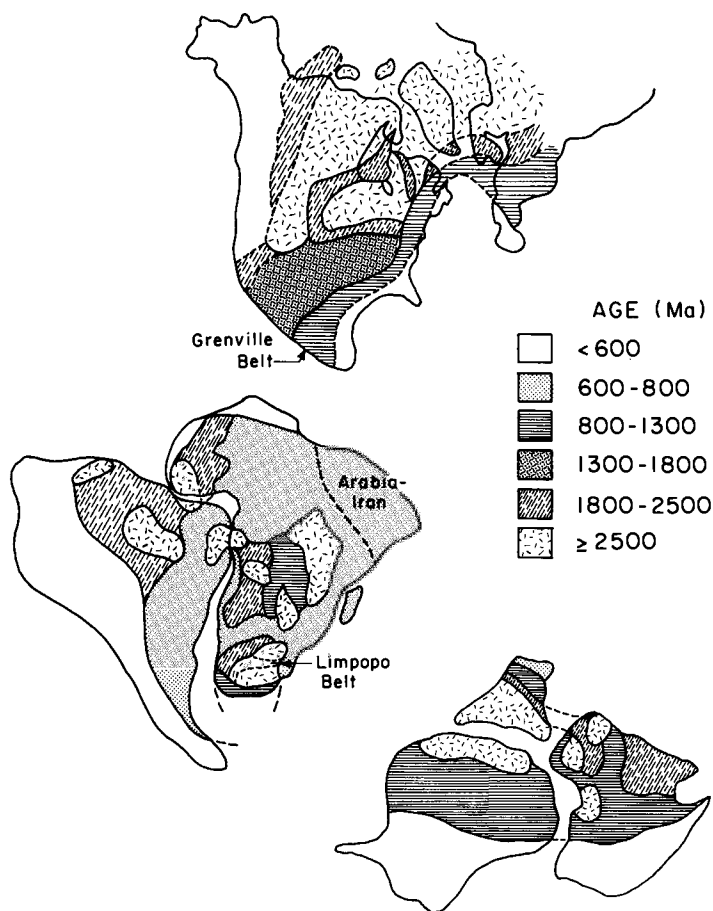


FIG 9.24 Paleomagnetic reconstructions of three mega-continents in the Proterozoic. Relative locations of the three continents are arbitrary. Province ages shown reflect the dominant orogeny.

Zimbabwe Provinces (fig. 9.24). It is approximately 200 km wide and can be traced along strike for 900 km, although geophysical data suggest it is much longer. The belt has been subdivided into three structural zones: a central zone, in which gneisses and other rocks are irregularly deformed with most structures trending oblique to strike of the belt, and two marginal zones, in which structural grain is generally parallel to the trend of the belt. Major ductile shear zones separate the marginal areas from the central zone. Gravity studies suggest that the shear zones dip to the south and reflect thrusting from this direction. Supracrustal rocks in the central zone, which are highly metamorphosed, are QPC assemblages of shale, quartzite, and carbonate. In the Messina area they have been intruded with a thick anorthosite-gabbro complex. Archean greenstone belts extend into the marginal zones of the Limpopo belt both on the north and south, and boundaries of the belt are gradational with the adjacent Archean granite-greenstone terranes.

A summary of the tectonic history of the Limpopo belt, based on field,

TABLE 9.3 *Summaries of Tectonic Histories of Two Proterozoic Mobile Belts*

Age (Ga)	Limpopo	Age (Ga)	Grenville
1.9	Intrusion mafic dikes	0.95	Intrusion mafic dikes
2.0	Renewed movement along shear zones	1.0	Deformation, syntectonic plutonism (second Grenville event) (D ₃)
2.2	Intrusion mafic dikes	→1.2	Deformation, metamorphism (first Grenville event) (D ₂)
2.46	Intrusion, Great Dyke	1.3	Intrusion granite
2.65	Major shear zones (D ₄)	1.4–	Intrusion anorogenic granite, anorthosite and gabbro
→2.7	Major deformation, syntectonic plutonism, metamorphism (D ₃)	1.5	
3.0	Intrusion mafic dikes	1.53	Deposition sediments and volcanics
→3.2	High-grade metamorphism, plutonism (D ₂)	1.6–	Eruption volcanics; intrusion granites
3.3	Intrusion layered igneous complex; deformation	1.65	
3.3	Deposition Beit Bridge sediments	→1.75–	Deformation, metamorphism, syntectonic plutonism (D ₁)
	Intrusion mafic dikes	1.85	Deposition volcanics and sediments
3.3	Archean crust formation (D ₀)	≥1.9	Archean crust formation (D ₀)
		2.6–	
		2.7	

Arrows indicate possible continental collisions; Limpopo.

After Barton (1983); Grenville after Gower (1985).

metamorphic and geochronologic studies, is given in table 9.3. Some of the oldest rocks exposed in the belt are the Sand River gneisses which yield U/Pb zircon dates of about 3300 Ma. The Beit Bridge sediments represent one of the oldest QPC assemblages known, and they may have been deposited on the Sand River gneisses. In addition to the Archean deformation (D₁), three major deformations (D₂–D₄) are recognized in the central Limpopo belt, the first two of which involve complex isoclinal folding and syntectonic plutonism and metamorphism. Possible continental collisions and extensive crustal reactivation accompany D₂ and D₃ (Van Reenen et al., 1987). The last major deformation was transcurrent movement along ductile shear zones (D₄) separating the central from the marginal zones of the mobile belt.

The Grenville Province is exposed in eastern Canada and probably underlies much of the western Appalachian Mountains to the south, as evidenced by sparse exposures and basement samples retrieved from drillcores. How far south the province extends, and what its relationship is to similar-age rocks in Texas and Mexico, are intriguing questions that are currently unanswered. In a pre-drift reconstruction of the North Atlantic, the Grenville Province extends into western Europe, exposed as the Gothics Province in Scandinavia (fig. 9.24). In eastern Canada the Grenville Province has been divided into seven subprovinces, based chiefly on structural and metamorphic styles. It is notable that the Grenville Province is thrust against the Superior Province on the west along the Grenville Front, and that the crust beneath the Grenville Province is thicker than that beneath the Superior Province.

Extensive radiometric dating in the Grenville Province indicates that it records approximately 1500 Ma of geologic history (Easton, 1981). Much of the western and

central parts of the province and an unknown amount of the eastern part of the province represent reworked Archean crust (fig. 9.25). At 1.8–1.9 Ga much of the province was deformed and intruded with granitic rocks, which affects large portions of the Canadian Shield, and the Grenville Front fault zone was initiated at this time. Between 1.4 and 1.5 Ga most of the anorthosite and related granite bodies were emplaced, and granulite-facies metamorphism occurred in the vicinity of the anorthosite massifs. Between 1.2 and about 1.4 Ga, thick successions of chiefly quartzite, shale, and carbonate were deposited unconformably on reworked Archean gneisses. Some arc-type volcanism and sedimentation occurred in the southern part of the province at about 1.3 Ga. At 1.1–1.2 Ga the Grenville orogeny, the major and most widespread period of deformation, occurred throughout the province. This event is characterized by isoclinal folding, regional metamorphism, and minor granitic plutonism. K–Ar ages indicate that the Grenville Province was uplifted and eroded to approximately its present level by 800 Ma ago.

The tectonic history of the Grenville Province varies with geographic location (fig. 9.25) and an unknown amount of the province may be composed of allochthonous terranes. A summary of the major events in the Labrador area is given in table 9.3. Supracrustal rocks deposited at 1.9 and 1.53 Ga appear to have rift affinities and were, in part, deposited on Archean basement. Two major orogenies are recorded at 1.75–1.85 Ga and 1.0–1.2 Ga, respectively. The first Grenville event at 1.2 Ga probably reflects a continental collision.

In comparing the Limpopo and Grenville chronologies, both involve at least two periods of intense deformation that may be caused by plate collisions. Stress patterns changed from compressive to extensional several times, and the continental crust was rifted at least once in each belt. Archean basement is present in both belts, and both belts were episodically active for ≥ 1000 Ma during the Proterozoic.

Precambrian Paleomagnetic Studies

The problems discussed in Chapter 5 in relation to defining ancient pole positions in the geologic past by paleomagnetic methods become more critical in Precambrian terranes. Radiometric dates that represent the age of magnetization are difficult to obtain in most Precambrian rocks due to deformation and metamorphism. Metamorphism may partially or completely reset either or both radiometric dates and magnetization directions, and it is often difficult to correlate the correct age with a measured pole direction. Recent studies indicate that $^{40}\text{Ar}/^{39}\text{Ar}$ dates in hornblende correspond most closely to magnetization blocking temperatures in magnetite, and little if any remanent magnetization can survive upper greenschist-facies metamorphism. Structural orientation of samples becomes more uncertain in deformed and metamorphosed terranes, and it is difficult to remove soft magnetization without affecting hard magnetization. In many successions, for instance basalts in the Belcher Island volcanics and sediments of the Huronian Supergroup in Canada, two periods of magnetization are recorded. One is pre- and the other post-folding. Three episodes of magnetic overprinting are recorded in the Archean Superior Province at 2600, 2100, and 1800 Ma (Schutts and Dunlop, 1981).

Because of these problems, Precambrian pole positions from a given locality are commonly more scattered than from comparable Phanerozoic sites. Some investi-

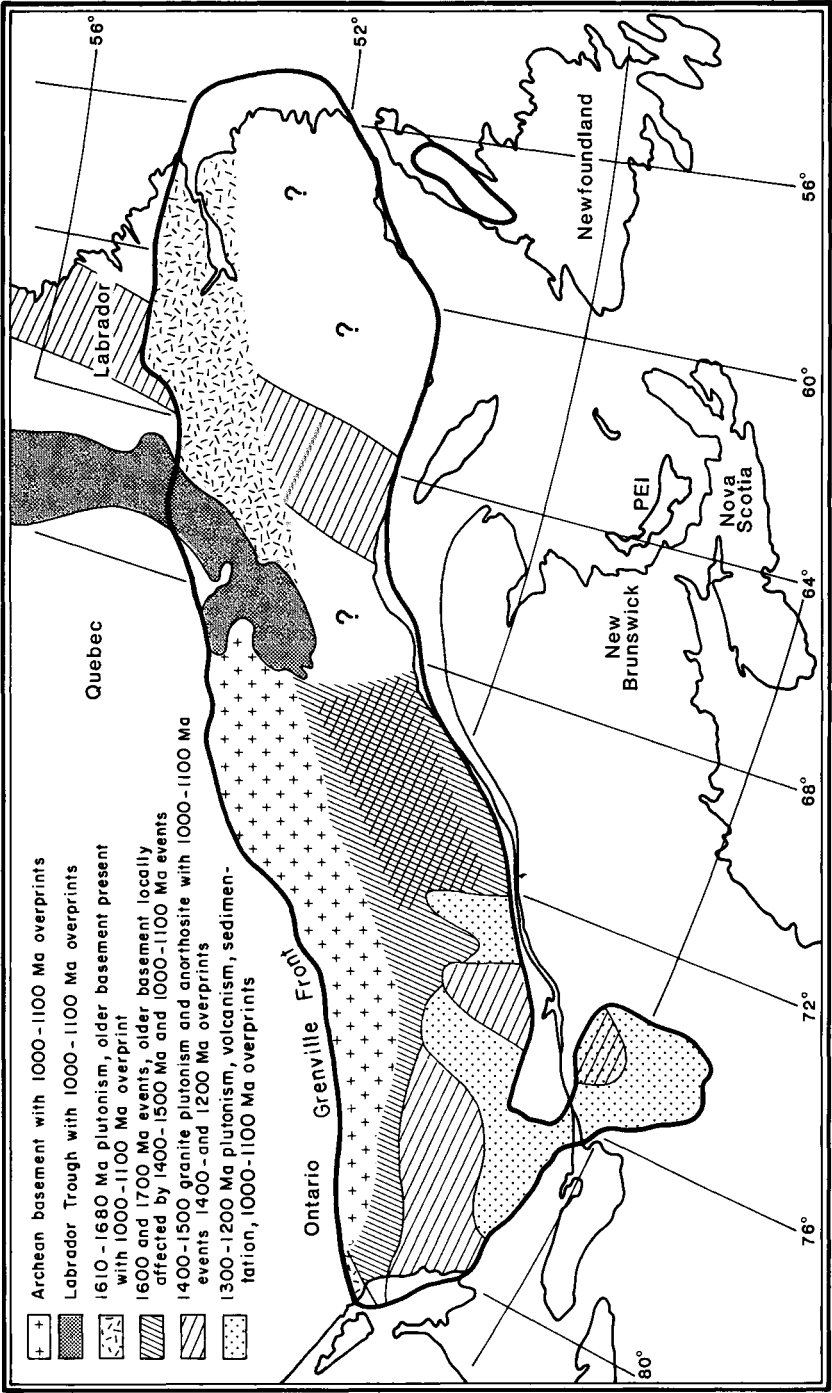


FIG 9.25 Geochronologic divisions of the Grenville Province. The nature of division boundaries is not well known. From Easton (1981).

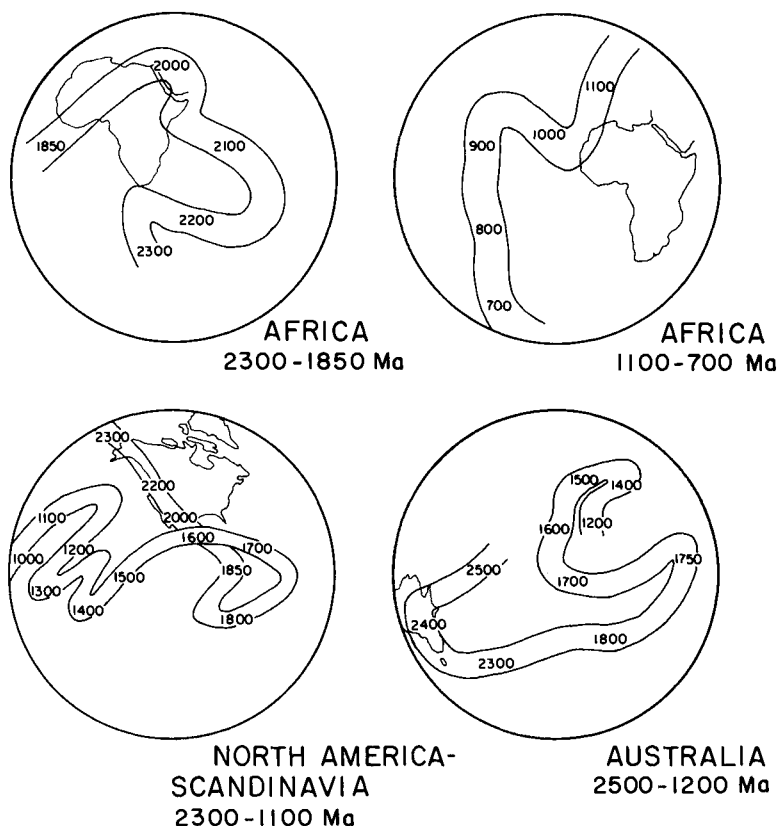


FIG 9.26 Schematic examples of Proterozoic APW paths. After Morgan and Briden (1981).

gators reject and accept data based on sources of error that are difficult to evaluate, and this leads to diverse and subjective apparent polar wandering curves. To minimize errors, Precambrian APW paths are usually drawn as wide bands or tracks encompassing many pole positions (fig. 9.26). APW tracks define curvilinear paths with approximately 10^8 a between loops.

The problem of whether loops exist in Precambrian polar wandering paths is dependent upon the length of the increments of time over which pole positions are averaged. When data from North America are averaged for time intervals less than 200 Ma, loops are resolved at 1150, 1300, 1350, 1400, 1750, 1800, and 1850 Ma (fig. 9.26). The two younger loops are defined from rocks from different geographic areas. The 1300 Ma loop is recorded by basalts from the Keweenaw rift and samples from stratigraphically lower horizons define the descending limb (1350–1300 Ma) whereas samples from progressively higher horizons define the ascending limb. Results indicate rapid motion of the North American plate in a northeasterly direction during a relatively short period of time (<20 Ma). The younger loop at 1000 Ma characterizes rocks from the Grenville Province and probably reflects cooling after a major collision of parts of the Grenville Province with the Superior craton at 1100–1200 Ma. Paleomagnetic studies from sedimentary

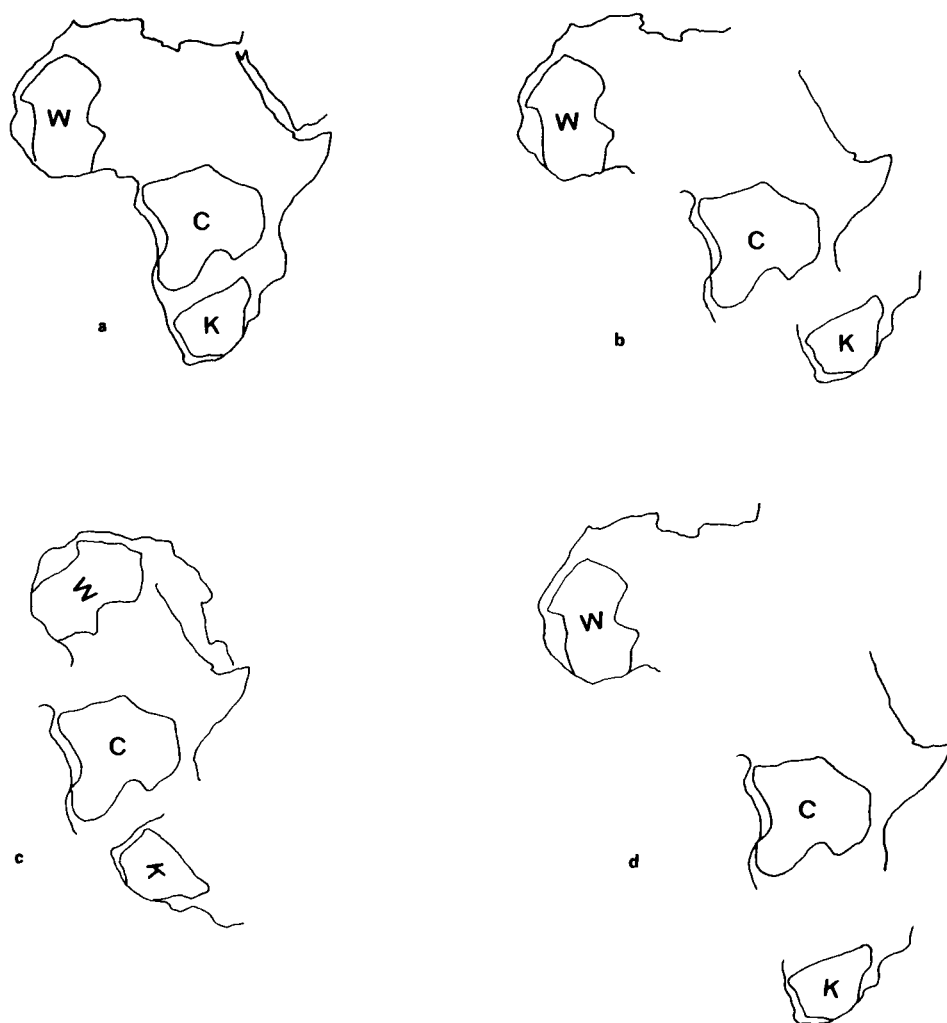


FIG 9.27 Diagrammatic illustration of motions between African Archean provinces which are allowed (a, b) and which are not allowed (c, d) by the tolerance levels of paleomagnetic dates. From Briden (1976).

and volcanic rocks of the Grand Canyon Supergroup in northern Arizona (1000–1200 Ma) indicate significant apparent polar wandering along a complex multiloop path as a function of stratigraphic height (Elston et al., 1973). Data suggest that loops in polar wandering curves with periods of 200 Ma are characteristic features for at least the last 2000 Ma. It is noteworthy that some of the Proterozoic loops correspond to major orogenies (1150, 1750, and 1850 Ma in North America; 1100 and 2150 in Africa). Perhaps these loops reflect rapid changes in plate motions following significant continental collisions.

Precambrian paleomagnetic data, despite inherent errors in interpretation, constrain the amount of independent motion between adjacent Archean provinces (Morgan and Briden, 1981). This is illustrated in figure 9.27 for African provinces.

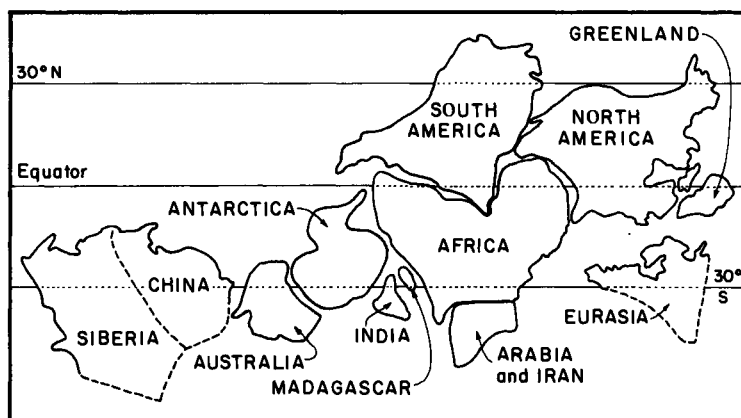


FIG 9.28 Paleomagnetic reconstruction of a Late Proterozoic (800–600 Ma) supercontinent. After Morel and Irving (1978).

It is possible, allowing for the various sources of error described above, to open and close small ocean basins less than 1000 km across, provided adjacent Archean provinces return to their approximate starting positions (McElhinny and McWilliams, 1977). Similar limitations on the amount of interprovince movement are documented in North America, Africa, and Australia. Results from North America and western Europe suggest that these continents were part of the same supercontinent during most of the Precambrian (Piper, 1980). When Scandinavia is returned to its Proterozoic position, crustal provinces can be matched from northeastern Canada across southern Greenland into the Baltic Shield (fig. 9.24). On such a reconstruction the 1.6–1.8 Ga accretionary terrane in southern North America appears to continue eastward into central and southern Scandinavia. Paleomagnetic results also indicate that Africa and South America were joined during the Proterozoic. With exception of about 1000 km of right lateral offset, which must have occurred after 2000 Ma ago, the Proterozoic reconstruction of these two continents is similar to the Pangaeic reconstruction (Onstott et al., 1984). Paleomagnetic evidence also indicates that Australia, India, and Antarctica were part of the same continent during most of the Proterozoic. What is not known from available data is: (1) how many other continents existed during the Proterozoic (including most of Asia), and (2) how the Proterozoic continents were distributed relative to each other. The fact that their APW tracks are remarkably different (fig. 9.26), however, indicates they drifted independently of each other during most of the Proterozoic.

In the Late Proterozoic (800–700 Ma), paleomagnetic data suggest that three Proterozoic mega-continents (fig. 9.24) may have collided to form a supercontinent (fig. 9.28). Although the geometry of this Late Proterozoic reconstruction is not well constrained, it is clear that it bears a resemblance to Pangaea. The positioning of China and Siberia west of Australia is speculative. As shown on the reconstruction, much of the supercontinent appears to have been located at low latitudes. Both paleoclimatic and paleontologic (Ediacaran fauna) data are consistent with this reconstruction. Another supercontinent reconstruction proposed by Piper (1982) is untenable both in terms of paleomagnetic and geologic data. The Late Proterozoic supercontinent began to break up about 750 Ma ago, and this break-up continued into the Early Paleozoic, completely fragmenting the supercontinent (Chapt. 8). It is

quite remarkable that, when Pangaea began to form in the Middle and Late Paleozoic, continental fragments returned to approximately the same positions they occupied in the Late Proterozoic supercontinent.

Minimum rates of Proterozoic continental drift can be obtained from APW tracks. This applies only to the rate of change of latitude relative to a fixed pole. Results indicate that Proterozoic plate velocities (3–10 cm/a) often exceeded those of present-day continents (≤ 3 cm/a) and were equivalent to present-day oceanic velocities (Ullrich and Van der Voo, 1981). Major peaks in velocity do not always correspond to major orogenies. Prior to 1 Ga, North American velocities are significantly greater than those of Asia or Europe.

Summary Statements

1. Increased heat flow in the Earth during the Precambrian was probably dissipated by increased rates of production of oceanic lithosphere at ocean ridges. This can be achieved by faster spreading rates, increased total ridge length, or both.
2. Archean crustal provinces are of two major types: granite–greenstone terranes and high-grade terranes. These two types of provinces probably reflect different tectonic settings.
3. Archean greenstone belts are composed of submarine volcanic-sediment successions and with increasing stratigraphic height exhibit decreases in the amount of komatiite (ultramafic lava) and basalt with corresponding increases in intermediate to felsic volcanics and volcanoclastic sediments. Greenstones exhibit low to moderate grades of metamorphism and polyphase deformation, in which compressional forces dominated during early stages and vertical forces during later stages of development.
4. Most Early Archean (≥ 3.5 Ga) greenstones differ from Late Archean greenstones in that they contain more komatiite, represent longer periods of time, and contain shallow-water sediments indicative of a relatively low-energy environment. They appear to have formed on relatively shallow submarine platforms in a tectonic setting unlike any found on the Earth today.
5. Late Archean greenstones appear to have formed in deep basins adjacent to major volcanic chains. Overall similarity of rock types, structural styles, and geochemical signatures of these greenstones to modern arc systems suggest a similar tectonic setting.
6. Archean high-grade terranes are composed of small amounts of supracrustal rocks engulfed by intrusive granitic rocks. These terranes reflect intense polyphase deformation and metamorphism, most of which record at least one period of granulite-facies metamorphism involving a CO_2 -rich fluid phase.
7. Archean high-grade supracrustal rocks are chiefly paragneisses (protoliths = shales) with small amounts of quartzite, carbonate, mafic to ultramafic intrusive rocks, and banded iron formation. These rocks appear to have been deposited in shallow-water cratonic basins.
8. Granitic rocks in Archean provinces are largely tonalite (greenstone terranes) or granodiorite (high-grade terranes). They appear to have formed by partial melting of mafic sources, probably at subduction zones.

9. Both continental rift and arc models are proposed for Archean greenstones with the arc models having the greatest success in meeting both geologic and geochemical constraints. Supracrustals in high-grade terranes may have formed in marginal cratonic basins that were deformed and buried during continental collisions.
10. Seven major rock associations are recognized in Proterozoic crustal provinces: (1) quartzite–pelite–carbonate (QPC), (2) bimodal volcanic–arkose–conglomerate (BVAC), (3) greenstone, (4) ophiolite, (5) anorogenic granite–anorthosite, (6) mafic dike swarms, and (7) layered igneous intrusions.
11. Most Proterozoic rock associations appear to have formed in tectonic settings similar to modern settings: (1) QPC, cratonic basins; (2) BVAC and mafic dikes, continental rifts or aulacogens; (3) greenstones, arc systems; and (4) ophiolites, ocean ridges.
12. Anorogenic granites and associated anorthosites occur in a broad belt extending from the southwestern United States to Labrador and across southern Greenland into Scandinavia. They exhibit an age range of 1350–1500 Ma in North America and range up to 1750 Ma in Finland. Although Phanerozoic examples of this association are unknown, geologic and geochemical data are consistent with an extensional tectonic setting.
13. Arc and craton lithologic assemblages are recognized back to the Early Archean, whereas continental rift and ophiolite assemblages are rare or perhaps non-existent prior to about 2 Ga ago.
14. Precambrian mobile belts often record several periods of intense deformation and metamorphism, and are episodically active for ≥ 1000 Ma. They appear to record one or more continental collisions.
15. Paleomagnetic data indicate during most of the Proterozoic that North America–Scandinavia, South America–Africa, and Australia–India–Antarctica comprised three mega-continents that drifted independently of each other. In the Late Proterozoic the mega-continents collided to form a supercontinent with relative continent positions similar to those found in Pangaea 400 Ma later.
16. With perhaps minor evolutionary changes, plate tectonics appears to have operated on the Earth since the Early Archean.

Suggestions for Further Reading

- Condie, K. C. (1981) *Archean Greenstone Belts*. Amsterdam: Elsevier. 434 pp.
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- Windley, B. F. (1984) *The Evolving Continents*, second edition. New York: J. Wiley. 399 pp.

CHAPTER 10

Origin and Evolution of the Crust and Mantle

Introduction

One of the unique features of the Earth is its crust. None of the other terrestrial planets seems to have a crust similar to that of the Earth. The origin of the Earth's crust appears to be related to plate tectonics and can be addressed in terms of several questions:

1. Was the first crust of local or worldwide extent?
2. When and by what process did the first crust form?
3. At what rate and by what processes did the early crust grow?
4. What was the composition of the early crust?
5. When and how did oceanic and continental crustal types develop?

The oldest preserved fragments of continental crust are about 3.8 Ga in age and are composed chiefly of tonalitic gneisses. These gneisses, in turn, contain fragments of komatiite and basalt (amphibolite), some of which may be remnants of the early oceanic crust. Model lead ages from the Earth and radiometric dates from meteorites suggest that the earliest terrestrial crust may have formed just after or during the late stages of planetary accretion at about 4.5 Ga. Although the original areal extent of Early Archean continental fragments is not known, they comprise less than 10% of preserved Archean crust (Plate I). The sparsity of rocks older than 3.0 Ga may be related to losses resulting from partial recycling of early crust into the mantle. Alternately, the first stable continental crust may not have formed until about 4 Ga. It may be possible to constrain the extent of the first crust by comparison with the lunar crust and crusts of other terrestrial planets. The lunar highlands, for instance, appear to represent remnants of the early lunar crust (4.4–4.5 Ga), which covered most or all of the Moon's surface (Taylor, 1975). Studies of topographic features on Mercury and Mars also suggest the preservation of widespread primitive crusts. If the early history of the Earth was similar to these planets, it also may have had an early crust that covered most of its surface.

Theories for the origin of the Earth's crust fall into three broad categories: (1) *inhomogeneous accretion* of the Earth, (2) *impact models*, and (3) *terrestrial models*. In the inhomogeneous accretion model, the last compounds to condense from the solar nebula produce a thin veneer on planetary surfaces rich in alkali and other volatile

elements, which may form or evolve into the first crust. A major problem with this model is that many nonvolatile elements, such as U, Th, and REE, which are concentrated in the core and lower mantle in an inhomogeneously accreted Earth, are today concentrated in the crust. This necessitates magmatic transfer from within the Earth, thus producing a crust of magmatic origin. Also, as discussed in Chapter 2, inhomogeneous accretion of the Earth is faced with other geochemical problems and is not a likely means to produce an early crust.

Several models have been proposed for crustal origin either directly or indirectly involving impact of accreting objects. These models include the infall of granitic asteroids that evolve into continents (Donn et al., 1965), and surface impacting which leads to melting within the Earth and the production of either mafic (oceanic) or granitic (continental) crust. Large impacts may have produced mare-like craters on the terrestrial surface that were filled with impact-produced magmas (Salisbury and Ronca, 1966; Grieve, 1980). If the magmas or their differentiation products were felsic, continental nuclei may have formed and continued to grow by magmatic additions from within the Earth. Alternately, if the impact craters were flooded with basalt, they may have become oceanic crust. Although initially attractive, impact models are faced with many difficulties in explaining crustal origin. For instance, most or all of the basalts that flood lunar mares formed later than the impacts and do not appear to be related directly to impacting; also, relatively small amounts of magma were erupted into lunar mares. Textures and structures so characteristic of impact, which should be preserved in Archean greenstones, are absent, as are impact breccias and high-pressure phases of SiO_2 . The timing of impact is also a problem in that most (or all) of the large impacting in the inner solar system occurred before the age of the oldest preserved continental crust (~ 3.9 Ga).

The terrestrial models, which call upon processes operating within the Earth, have been most successful in explaining the origin of the Earth's crust. The fact that textures and geochemical relationships indicate that the early lunar crust is a product of magmatic processes favors a similar origin for the Earth's early crust. It is likely that enough heat was retained in the Earth after or during the late stages of accretion that the upper mantle was partially or entirely melted. Complete melting of the upper mantle would result in a magma ocean, which upon cooling should give rise to a primitive widespread crust (fig. 10.1). Even without a magma ocean, extensive melting in the early upper mantle should produce large quantities of magma, some of which rise to the surface forming an early crust. Whether plate tectonics was operative at this time is not known. However, some mechanism(s) of plate creation and recycling must have been operative to accommodate the large amounts of heat loss and vigorous convection in the early mantle. A summary of the major characteristics of the early terrestrial crust, as discussed in later sections, is given in table 10.1.

The Possibility of a Magma Ocean

The cumulus textures and uniform composition and age of the lunar highland crust are suggested as evidence for a widespread magma ocean early in the Moon's history (Taylor, 1982). Crystallization of a well-mixed magma ocean should produce a crust of uniform composition as well as a layered structure in the outer part of the Moon.

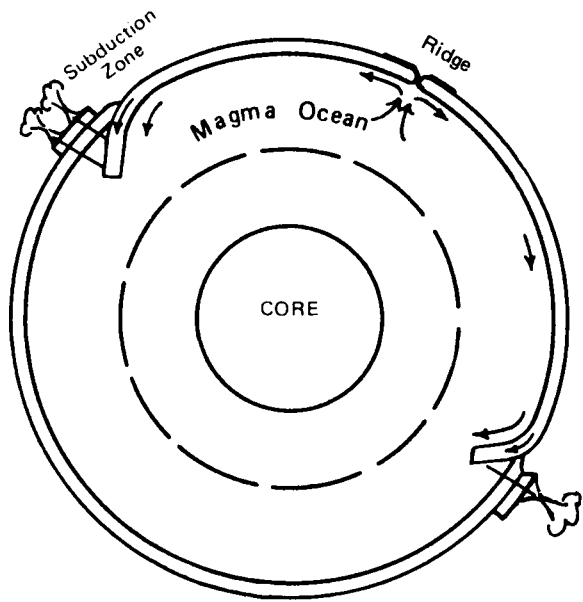


FIG 10.1 Model for the production of an early, probably komatiitic crust above a magma ocean on the Earth.

TABLE 10.1 *Summary of Characteristics of the Earth's Early Crust*

	Oceanic Crust	Continental Crust
First appearance	~ 4.5 Ga	~ 4.0 Ga
Where formed	Ocean ridges	Subduction zones (sinks)
Composition	Komatiite–basalt	Tonalite–granodiorite
Lateral extent	Widespread	Local
How generated	Partial melting of ultramafic rocks in upper mantle	Partial melting of wet mafic rocks in descending slabs

The possible existence of a terrestrial magma ocean also has been suggested, based on the large amount of initial heat available in the Earth (fig. 2.9). Furthermore, estimates of the temperature regime in the accreting Earth allow the outer part of the Earth to melt during accretion. Depth estimates for a terrestrial magma ocean range from 100 to over 1000 km, and the composition is generally assumed to be ultramafic (komatiitic) (Ohtani, 1985). Because of vigorous convection in a magma ocean, solidus and liquidus temperature gradients are steeper than the adiabatic gradient. Hence, crystallization will begin at the bottom where the liquidus intersects the adiabat and progress upwards. Because of rapid heat loss due to convection, crystallization should be essentially complete in <100 Ma. Although such crystallization should lead to a layered upper mantle, mixing of cumulate minerals with residual liquids may give rise to a compositionally uniform upper mantle.

If a terrestrial magma ocean existed, the earliest crust probably formed on its surface (fig. 10.1). Such a crust would be komatiitic in composition and subject to

rapid break-up and recycling into the magma ocean, both from vigorous convection and major impacts on the Earth's surface. If surfaces of magma in modern volcanic craters can be used as scaled-down models for Archean crust, plate tectonics would probably exist in an early crust overlying a magma ocean. Active ridges, subduction zones and transform faults have been described from the crust on ponded magma in Kilauea crater in Hawaii (Duffield, 1972).

Regardless of whether a magma ocean existed on the early Earth, the earliest crust was probably widespread due to extensive melting of the mantle.

Composition of the Primitive Crust

Numerous compositions have been suggested for the Earth's earliest crust. In part responsible for diverging opinions are the different approaches to the subject. The most direct approach is to find and describe a relict of the primitive crust. Although some investigators have not given up on this approach, the chances that a remnant of this crust are preserved seem very small. Another approach is to deduce the composition of the early crust from studies of the preserved Archean crust. However, compositions and field relations of rock types in the oldest preserved Archean terranes may not be representative of earlier crust. The oldest known supracrustal rocks in the Isua greenstone belt (~3.8 Ga) contain a mixture of volcanic rocks (mafic, komatiitic, felsic), quartzites, iron formation, carbonates and pelitic rocks. Another approach has been to assume that the Earth and Moon have undergone similar early histories, and hence to go to the Moon, where the early record is well preserved, to determine the composition of the Earth's primitive crust. Geochemical models based on crystal-melt equilibria and a falling geothermal gradient with time have also been used to constrain the composition of the early terrestrial crust.

Felsic Models

Some models for the production of a primitive felsic or andesitic crust rely on the assumption that low degrees of partial melting in the mantle will be reached before high degrees, and hence felsic magmas should be produced before mafic ones. Other models call upon fractional crystallization of basalt to form andesitic or felsic crust. Ramberg (1964) develops a model based on laboratory studies using various substances of different densities and viscosities, and concludes that felsic magmas, once formed, should rise to the surface buoyantly before basaltic magmas, which have higher densities. Shaw (1976) presents a geochemical model for the formation of a widespread anorthositic-sialic crust. He proposes that a molten mantle cools and crystallizes from the centre outwards, concentrating incompatible elements into a near-surface basaltic magma layer. This layer undergoes fractional crystallization, resulting in the accumulation of an anorthositic scum in irregular patches, and in residual granitic magmas that crystallize to form the first stable crust by about 4 Ga.

Two main obstacles face the felsic crustal models. First, the high heat generation in the Early Archean probably produced large degrees of melting of the upper mantle, and hence it is unlikely that felsic melts could form directly. A felsic or andesitic crust, however, could be produced by fractional crystallization of basaltic

magmas. If widespread felsic crust were produced in this manner, and it is essentially nondestructible because of its low density, why are not remnants of the early crust preserved today?

Anorthositic Models

Studies of lunar samples indicate that the oldest rocks on the lunar surface are gabbroic anorthosites and related high-alumina basalts of the lunar highlands (Taylor, 1982). These rocks are remnants of a widespread lunar crust formed between 4.4 and 4.5 Ga. This crust appears to have formed in response to catastrophic heating that led to widespread melting of the lunar interior and the production of voluminous basaltic magmas forming a magma ocean. Near the surface, the magma oceans cooled and underwent fractional crystallization. Pyroxenes and olivine sank and plagioclase (and some pyroxenes) floated, forming a crust of gabbroic anorthosites. Impact disrupted this crust and produced mare craters, and these craters were later filled with basaltic magmas (3.1–3.9 Ga). The highland areas of the Moon represent remnants of this primitive gabbroic anorthosite crust.

Windley (1970) has pointed out that Early Archean anorthosites in Greenland are similar in composition (i.e. high An content, associated chromite) to lunar anorthosites and not to younger terrestrial anorthosites. It is clear from field relationships, however, that the Greenland anorthosites are not remnants of an early terrestrial crust, since they commonly intrude tonalitic gneisses. If, however, the Earth had an early melting history similar to that of the Moon, the first crust may have been composed dominantly of gabbroic anorthosites. The terrestrial Early Archean anorthosites may represent the last stages of anorthosite production, which continued after both mafic and felsic magmas were also being produced.

The increased pressure gradient in the Earth limits the stability range of plagioclase to depths considerably shallower than on the Moon. Experimental data suggest that plagioclase is not a stable phase at depths >35 km in the Earth. Hence, if such a model is applicable to the Earth, the anorthositic fraction, either as floating crystals or as magmas, must find its way to very shallow depths to be stable. The most serious problem with the anorthositic model is related to the hydrous nature of the Earth. Plagioclase will readily float in an anhydrous lunar magmatic ocean but even small amounts of water in the system cause it to sink (Taylor, 1979b). Hence in the terrestrial system, where water was probably abundant in the early mantle, an anorthositic scum on a magma ocean would not form.

Komatiite–Basalt Models

In terms of our understanding of the Earth's early thermal history (Chapt. 2) and from the geochemical and experimental data base related to magma production, it seems likely that the Earth's primitive crust was komatiitic or basaltic in composition. If a magma ocean existed, cooling would produce a komatiitic crust. Without a magma ocean (or after its solidification), basalts also may have composed an important part of the early crust. The importance of these two rock types in Early Archean greenstone successions attests to their probable importance on the surface of the Earth during the Archean.

Origin of the Crust

Oceanic Crust

If the Earth was rapidly degassed the first 50 Ma after accretion, as suggested by some Xe isotopic data (Chapt. 11), the oceans would appear very early in geologic history. As mentioned above, it is likely that widespread melting in the upper mantle gave rise to magmas that produced the first crust, which was probably komatiitic in composition. This crust, which may have begun to form during the late stages of accretion, would underlie the early oceans, and represent the earliest oceanic crust. As with modern oceanic crust, it would have been generated at ocean ridges. Because of the greater amount of heat in the Archean upper mantle, the oceanic crust may have been produced 4–6 times faster than at present. Alternately, if this excess heat was lost chiefly by a longer total ridge system rather than by faster seafloor spreading, rates of crustal production may have been comparable to present-day rates. In either case, oceanic lithosphere and crust would be recycled into the mantle at “sinks.” These sinks may have had different geometries from modern subduction zones. For instance, if lava-lake plate tectonics (Duffield, 1972) is more analogous to Archean than to modern plate tectonics, two plates may have been subducted at sinks (i.e. dual subduction). Surface impacts, which were important up to about 3.9 Ga, may also have contributed to recycling of plates (see fig. 10.10).

If komatiites were important in the earliest oceanic crust, which seems likely, they may have provided the density contrast needed to initiate plate subsidence. It would appear that a komatiitic crust ($\rho = 3.2 \text{ gm/cm}^3$) is slightly more dense than a hot, partially molten upper mantle ($\rho = 3.1 \text{ gm/cm}^3$) and should sink (Nisbet and Fowler, 1983). Hence, komatiites may have provided the major driving force for the onset of the Earth's first plate tectonics.

Continental Crust

The Earth may be the only terrestrial planet with continental crust. If so, what is unique about the Earth that gives rise to continents? Two factors immediately stand out: (1) the Earth is the only planet with significant amounts of water, and (2) it may be the only planet on which plate tectonics has been operative for a significant amount of time. Let us review some of the characteristics of Archean continental crust and then return to the question of how water and plate tectonics are necessary for continent formation.

The oldest preserved continental crust (3.5–3.9 Ga) occurs as small crustal provinces (see fig. 10.8). These provinces are generally <500 km across, (although the Pilbara Province in Western Australia is larger) as typified by the Amitsoq terrane in SW Greenland (fig. 10.2.). Nd and U–Pb isotopic studies indicate that these provinces were probably not much larger than their present sizes, and that surrounding later Archean terranes were added to their perimeters after 3.5 Ga. As discussed in Chapter 9, Early Archean crust is composed largely of felsic gneisses (tonalite to granodiorite in composition) containing remnants of supracrustal rocks representing at least two different tectonic settings. The oldest well-documented occurrences of continental crust are in Enderby Land in Antarctica (3.9 Ga), and

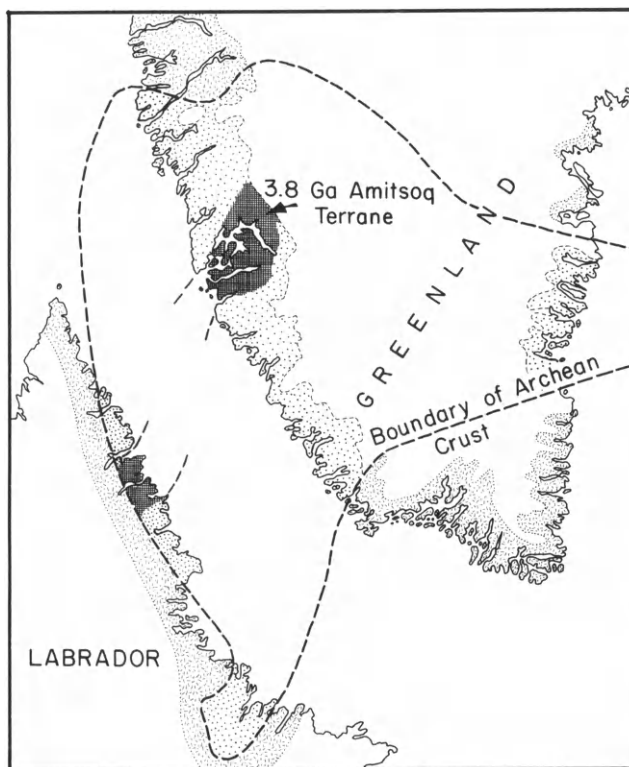


FIG 10.2 Distribution of Archean continental crust in Southwest Greenland and northeastern Canada.

the Amitsoq terrane of SW Greenland (3.8 Ga) (fig 10.3) (Moorbath et al., 1986; Black et al., 1986). The existence of detrital zircons in early Archean sediments that give ion-probe U–Pb dates of 4.0–4.2 Ga (Froude et al., 1983), however, suggest but do not prove the existence of still older continental crust which is not preserved. If these old detrital zircons come from granitic sources, which seems likely, at least small islands of continental crust must have been present by 4 Ga ago.

An important constraint on the origin of Archean continents is the composition of Archean tonalites and granodiorites. These rocks are similar to post-Archean counterparts in terms of most incompatible element distributions and especially in terms of the negative Ta–Nb anomalies on primitive-mantle normalized plots (fig. 10.4). Experimental data favor an origin for Archean tonalites by partial melting of amphibolite or eclogite in descending slabs in the presence of significant amounts of water. On the other hand, post-Archean felsic magmas appear to be the products of basalt fractional crystallization under relatively dry conditions. The production of large amounts of Archean continental crust requires subduction of large quantities of hydrated basalt or komatiite and large quantities of water. Hence, with the possible exception of Venus, the absence of continental crust on other terrestrial planets may reflect the small amounts of water and the absence of plate tectonics on these planets. It is likely that the earliest terrestrial continents were produced at

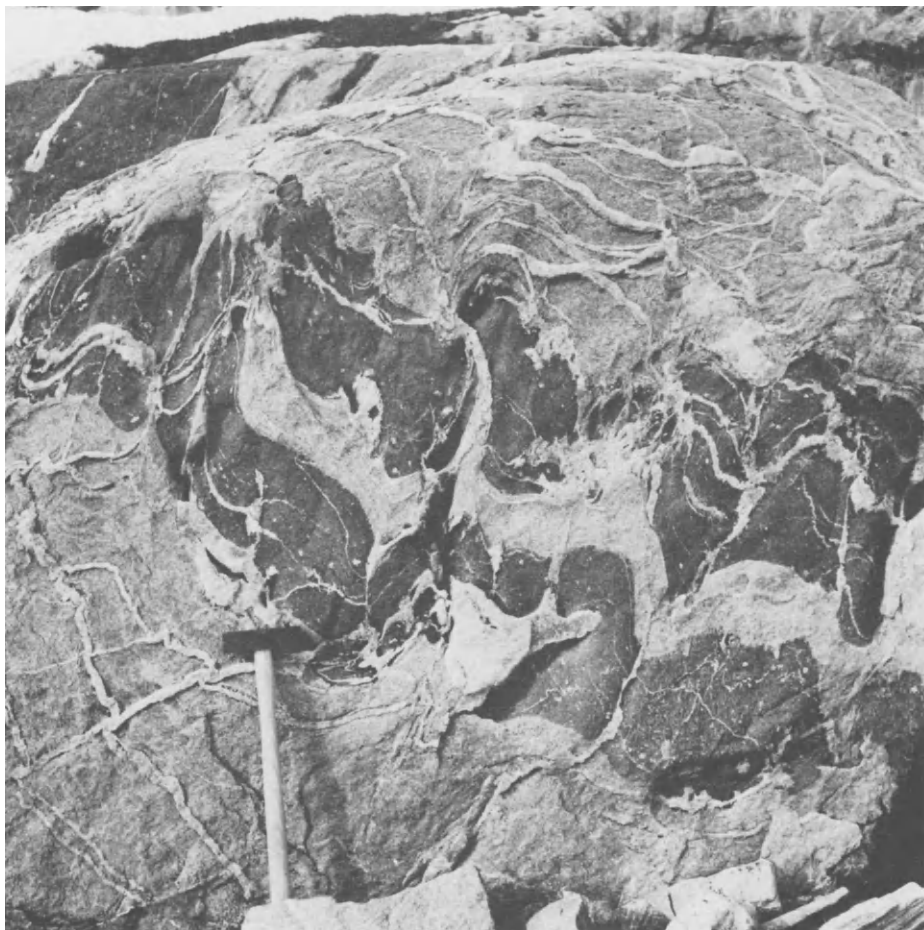


FIG 10.3 Deformed amphibolite and tonalitic gneisses from the 3.8-Ga-old Amitsoq gneiss terrane in Southwest Greenland. Courtesy of V. McGregor.

subduction zones (or other sinks) as temperatures fell to levels where descending wet mafic ($\pm$ komatiitic) crust underwent relatively small degrees of melting, producing felsic magmas. These magmas underplated mafic and komatiitic arc systems, some of which are preserved today as greenstone belts. True granites do not appear in the geologic record until about 3 Ga, and do not become important until after 2.6 Ga. Geochemical and experimental data suggest that these granites are produced by partial melting or fractional crystallization of tonalite (Condie, 1986), and it is not until tonalites are relatively widespread that granites appear in the geologic record. Thus, the story of early continental crust is the story of four rock types: *komatiite*, *basalt*, *tonalite*, and *granite* listed in the general order of appearance in the Archean geologic record. Field relations in most Archean granite–greenstone terranes also indicate this order of relative ages. Early Archean komatiites and basalts may be produced at ocean ridges and basalts at subduction zones. These rocks are hydrated by reaction with the oceans, and as the mantle cools beneath subduction zones, they

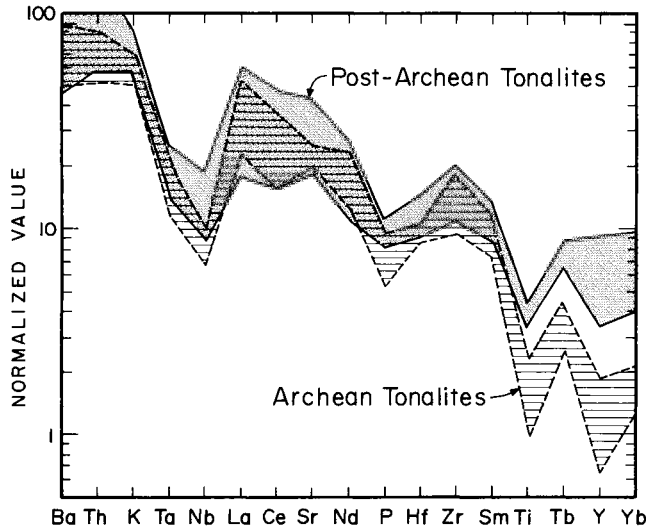


FIG 10.4 Primitive-mantle normalized distribution of incompatible elements in Archean tonalites compared to post-Archean tonalites. Normalization values from Wood (1979).

are partially melted to produce tonalites, which in turn are partially melted or undergo fractional crystallization to produce granites.

Studies of the modern seafloor indicate that ocean depth falls as the square root of the age of oceanic crust (eq. 4.4). In the Archean, when it is likely that plates were smaller and subducted earlier than today, the depth to the seafloor should be less than today, and for an equivalent amount of seawater the Archean continents should be flooded more than the present continents. The near-absence of terrestrial sediments in Archean supracrustal successions is consistent with this observation.

Mechanisms of Continental Growth

Various mechanisms have been suggested for the growth of the crust. *Crustal growth* can occur by additional of new material from the mantle, by readdition of crustal material that has been cycled through the mantle, and by redistribution of crustal rocks due to sedimentary and tectonic processes. *Net crustal growth*, however, includes only additions to the crust of new (unrecycled) material from the mantle. Sr, Pb, and Nd isotopic systems are useful in distinguishing new from recycled crustal rocks. The major mechanisms of crustal growth are magma additions, thrusting and stacking of crustal rocks, aggregation by microcontinent and arc collisions, and addition of sedimentary prisms to continental margins.

As erosion removes material from the continents it accumulates on continental shelves and slopes and in ocean basins near continental margins. Thick sedimentary prisms may form in this manner. Burial leads to metamorphism and perhaps to partial melting, the net result being lateral growth of the continents. How important this mechanism of growth was in the geologic past is difficult to evaluate, because ancient continental margins have been intensely reworked during later deformation. However, it may have been important adjacent to high-grade Precambrian provinces, where evidence indicates 15–30 km of uplift and erosion has occurred.

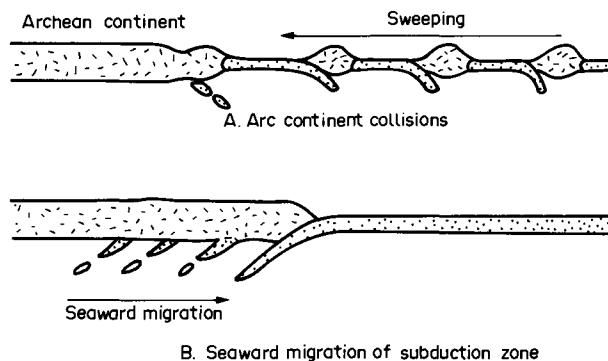


FIG 10.5 Possible growth mechanisms of continents at convergent plate boundaries.

Magma from the mantle may be added to the crust and lithosphere by underplating, involving the intrusion of sills and plutons, and by overplating of volcanic rocks. Such additions may occur in a variety of tectonic environments. Field relationships in exposed lower crustal sections, such as the Ivrea Zone in Italy (fig. 4.23), suggest that many mafic granulites are intrusive gabbros, and that additions of mafic magmas to the lower continental crust may be important.

Growth of continents and arcs may occur by additions of magma from descending lithospheric slabs, and continents and oceanic islands may thicken by addition of magmas from mantle plumes. Investigations of Archean high-grade terranes suggest that the early crust thickened by thrusting and stacking of thrust sheets and nappes of both oceanic and continental rocks. Such intimate intermixing of these rocks implies intense horizontal compression, which is probably caused by collisions of arcs and continents. Continental growth occurs when an arc collides and becomes sutured to another arc or a microcontinent (fig. 10.5A), and also during seaward migration of subduction zones (B). As discussed in Chapter 8, terrane accretion has been an important mechanism of continental growth during the Phanerozoic. Most of the Cordilleran and Appalachian orogenic belts in North America represent collages of terranes added by arc collisions either at convergent plate boundaries or along transform faults. Similar collisions were also important during the Precambrian. It would appear that collision of arcs and microcontinents is the most important mechanism by which continents have grown.

Thickness of Early Crust

The chief factor controlling the thickness of oceanic crust is the temperature of the upper mantle beneath ocean ridges. The higher the temperature, the greater the degree of melting and the larger the region that is melted (Sleep and Windley, 1982; Campbell and Jarvis, 1984). For a given volume of mantle, higher degrees of melting result in more magma for extrusion and thus in a thicker oceanic crust. Melting of a larger part of the upper mantle has the same effect, and hence higher mantle temperatures during the Archean should have resulted in a thicker oceanic crust than at present. Models suggest that the early oceanic crust may have been as thick as 20 km, and even thicker over subduction zones. This crust may have been largely komatiitic in composition due to higher Archean mantle temperatures.

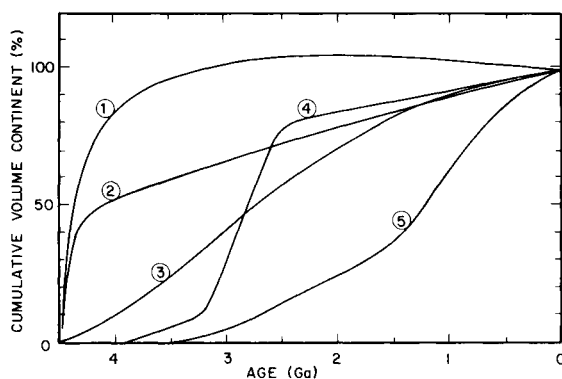


FIG 10.6 Examples of proposed continental growth rates through geologic time. See text for explanation.

Metamorphic mineral assemblages in exposed Archean crustal sections indicate that Archean continents were equal or greater in thickness to the present continents. These mineral assemblages, which imply depths of 30–50 km (Chapt. 4), are underlain today by 30–40 km of continental crust. To explain this observation, it would appear either that underplating has kept pace with uplift and erosion, or that at least some Archean continental crust was 60–80 km thick. Although mafic underplating has certainly occurred, the fact that radiometric dates do not generally become younger with increasing level of crustal exposure suggests that such underplating occurred at or soon after (in <100 Ma) the time of crustal formation. Thickening of the crust to >50 km is a natural consequence of continental collisions and provides an attractive explanation for Archean high-*P* granulites. Another interesting consequence of this model is that Archean mountain ranges developed in response to collision should have as great of relief as modern collisional mountain ranges like the Himalayas.

Continental Growth Rates

The Problem of Recycling

Continental growth is the net gain in mass of continental crust per unit time. Because continental crust may be extracted from the mantle or returned to the mantle by subduction, crustal growth rate can be positive, zero or even negative. Many different models of continental growth rate have been proposed (fig. 10.6). These are based on one or a combination of data sources including: (1) Pb, Sr, and Nd isotopic data from igneous rocks; (2) Sr isotope ratios of marine carbonates; (3) the constancy of continental freeboard through time; (4) Phanerozoic crustal addition and subtraction rates; (5) the areal distribution of radiometric dates; and (6) estimates of sediment recycling rates. The earliest models for continental growth were based chiefly on the geographic distribution of radiometric dates on the continents. These models suggested that continents grew slowly in the Archean and rapidly after 2 Ga ago (curve 5, fig. 10.6). We now realize that this is not a valid approach to estimating crustal growth rates because many of the Rb–Sr and K–Ar dates used in such studies have been reset during later orogenic events, and the true crustal formation age is older than the dates suggest. On the opposite extreme are

models that suggest very rapid growth early in the Earth's history, followed by extensive recycling of continent back into the mantle (curve 1, fig. 10.6) (Reymer and Schubert, 1984). Such recycling is necessary because we do not see this old (>4 Ga) continental crust on the Earth today. Perhaps the easiest way to recycle continent is by subduction of continental sediments, although other mechanisms have been proposed. Other growth models fall in between these extremes and include approximately linear growth with time (curve 3), rapid early growth followed by linear growth (curve 2) and rapid growth in the Late Archean (curve 4).

There are several major problems with models that call upon rapid continental growth prior to 4 Ga followed by extensive crustal recycling into the mantle. The first is that the constancy of continental freeboard (mean elevation of continents) together with decreasing heat flow over the last 4 Ga implies that the continents must have grown during this time (Reymer and Schubert, 1984). If not, as the ocean basins subsided due to mantle cooling, continents should become more emergent, which is not supported by geological data. A second problem is how to avoid melting of continental materials before they are carried deep into the mantle by convection (Condie, 1986). If hydrous continental sediments were subducted in the Archean with only moderate geotherms of $30^{\circ}\text{C}/\text{km}$, they should completely melt at depths <100 km. Even if the sediments were dehydrated before melting began, they should undergo large degrees of melting ($>50\%$) between 100 and 150 km, and if geotherms were steeper beneath Early Archean subduction zones, which seems likely, sediments should melt at even shallower depths. Because sediment melts are granitic in composition, they would rise into the crust, and it is difficult to see how significant quantities of sediment could be recycled into the deep mantle. This observation is also consistent with geochemical and isotopic data from modern arc volcanics, which suggest that $<10\%$ of subducted sediment contributes to the production of arc magmas (Chapt. 7). Models that involve no growth of continents after the Early Archean also require that all sediments delivered to ocean basins (≤ 1 km^3/a) be subducted and recycled to sustain such a model (McLennan and Taylor, 1983). This is clearly unacceptable, as we see large quantities of ancient sediments preserved in the continental crust.

Still another problem with recycling is provided by Nd isotopic data from sediments. If the true stratigraphic age of sediments is plotted against the Nd model age (relative to chondrites), progressively younger sediments deviate farther from the equal age line (fig. 10.7). This reflects reworking of older continental crust and a progressively greater proportion of older crust entering the sediment record with time (Miller et al., 1986). Chemical and biochemical sediments show a similar pattern (Miller and O'Nions, 1985). The fact that most sediments older than 3 Ga do not greatly deviate from the equal age line, indicates that very little continental crust existed prior to this time. However, ϵNd values from Isua sediments in SW Greenland (with a stratigraphic age of 3.7–3.8 Ga) are slightly positive (+1 to +3), which requires a pre-3.8-Ga depletion event in the mantle. Although this event could be a continent-forming event, it could also be an event that formed an enriched mantle reservoir (Condie, 1986b; Jacobsen and Dymek, 1988). Thus, pre-4.0-Ga continental contributions to Archean sediments are not required by Nd isotopic results. Although Nd isotopic data do not favor early recycling of continental crust, the low initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in many post-Archean igneous

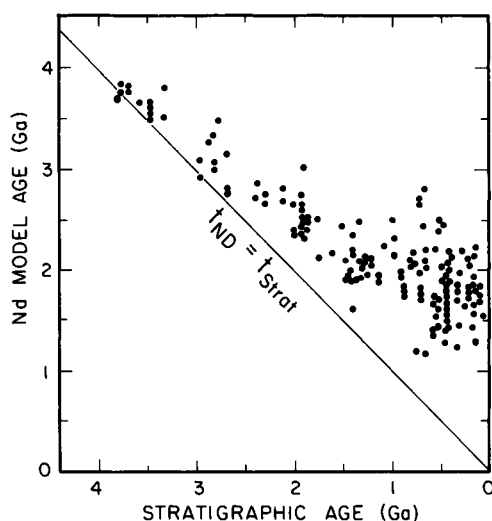


FIG 10.7 Comparison of stratigraphic and Nd model ages (relative to chondrites) of fine grained terrigenous sediments. Data compiled from many sources.

rocks (negative ϵNd values) are consistent with recycling of some enriched component into the mantle after 2.5 Ga (DePaolo, 1983). One candidate for this component is continental sediments. The melting argument discussed above may not be a problem after 2.5 Ga, because mantle temperatures are lower after this time.

Nd Isotopic Constraints

As discussed in Chapter 9, large segments of the continental crust have been reworked several times since their first extraction from the mantle. Because reworking does not appreciably fractionate Sm and Nd (Chapt. 4), Nd isotopes may be capable of seeing through later orogenies to the time when continental crust is first separated from the mantle. This time is known as the *crustal formation age*. Two approaches are used to estimate crustal formation ages—Sm–Nd isochron dating and Sm–Nd model ages. Although whole-rock Sm–Nd isochron dates are generally equivalent to crustal formation ages (Patchett and Arndt, 1986), they usually have large errors due to the relatively small range in Sm/Nd ratios in most granitic rocks. In many instances, Sm/Nd ratios are so similar among samples that it is not possible to obtain meaningful isochrons. The second approach employs U–Pb zircon dates, which provide an igneous crystallization age, coupled with the average Nd isotopic composition and Sm/Nd ratio of a rock or a crustal segment. Based on the Sm/Nd ratio, a growth curve is constructed for the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio. This growth curve is extrapolated to a model mantle composition, the intersection of which is assumed to represent the crustal formation age (Arndt and Goldstein, 1987). Because of the uncertainties in mantle composition and Sm/Nd ratio, caution should be used in estimating crustal formation ages in this manner.

The distribution of reasonably well-established crustal formation ages shows an episodic pattern, much like that exhibited by U–Pb zircon dates (see fig. 10.22) but



FIG 10.8 Distribution of Archean continental crust on a Pangaeic reassembly of the continents. This distribution does not reflect the original distribution of Archean continents, and paleomagnetic data do not support the existence of one supercontinent in the Archean.

with a lag time of up to several hundred million years. It is noteworthy that a large number of crustal formation ages fall in the range of 2.7–3.0 Ga. The widespread geographic distribution of these ages suggests that portions of the continental crust formed during the Late Archean (fig. 10.8). Estimates of the amount of continental crust produced between 2.7 and 3.0 Ga range from a minimum of 50% to as much as 80% (Taylor and McLennan, 1985). Clearly, much of the crust classified as Proterozoic mobile belts is reworked Archean crust. A model for rapid growth of the continents in the Late Archean is illustrated by curve 4 in figure 10.5.

Rates of Post-Archean Continental Growth

Rates of Cenozoic continental growth are estimated from rates of magma addition to arcs and continents, and rates of loss primarily by subduction and subduction-zone erosion. Despite uncertainties in estimating gains and losses, most calculations suggest net growth rates of 0.5–1.5 km³/a. Because of the secular decline in heat production in the mantle, oceans must deepen with time. Together with the constant freeboard of continents, this necessitates continental growth since the end of the Archean of about 1 km³/a (Reymer and Schubert, 1984). Nd isotopic dates from North America and northwestern Europe indicate that about 35% of the pre-1.6 Ga continental crust was extracted from the mantle between 1.7 and 1.9 Ga (Patchett and Arndt, 1986), which corresponds to a crustal production rate of about 1.2 km³/a,

similar to estimated Phanerozoic rates. This can be equated to total growth rate only if most of Early Proterozoic continental growth was concentrated in North America and northwestern Europe. Data from other continents, however, indicate that substantial continental additions occurred elsewhere, and thus Early Proterozoic growth rates are at least twice the calculate Phanerozoic rates. Corresponding estimates for the Archean suggest continental growth rates of $3\text{--}4\text{ km}^3/\text{a}$, occurring principally in the late Archean. In both the Late Archean and the Early Proterozoic, continental growth rates appear to have exceeded Phanerozoic growth rates.

Towards a Growth Rate Curve

In addition to Nd crustal formation ages there are several other lines of evidence that support a period of rapid continental growth in the Late Archean. The large volume of cratonic sediments between 2000 and 2400 Ma in age requires widespread continents to serve as source areas. The rather abrupt appearance of large quantities of sediment in the Early Proterozoic is consistent with rapid continental growth in the Late Archean. In addition, the following lines of evidence, each of which is discussed more fully in later sections, support rapid continental growth between 2700 and 3000 Ma:

1. increase in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of marine carbonates beginning in the Early Proterozoic;
2. geochemical and Nd isotopic evidence for relatively undepleted mantle in the early Archean but widespread depleted mantle by 2700 Ma; and
3. recycling models for sediments.

In addition, Nd isotopic evidence from North America and the Baltic Shield in Scandinavia strongly suggests that large quantities of continental crust were produced between 1700 and 1900 Ma and especially between 1850 and 1900 Ma (Patchett and Arndt, 1986). In general, Nd crustal formation ages suggest episodic continental growth with the Late Archean and early Proterozoic representing the times of most rapid growth (see fig. 10.13).

The origin and significance of episodic continental growth is not understood and may be related to episodic orogeny. The rapid continental growth in the Late Archean as implied by various lines of evidence, appears to represent a unique, perhaps catastrophic event in Earth history. One model that may explain such rapid continental growth is the eclogite transition model proposed by Condie (1986). In this model, basalt, which is relatively buoyant, congregates over mantle sinks prior to 3.0 Ga where it forms thick ($>50\text{ km}$) submarine plateaus (fig. 10.9). As cooling continues beneath sinks, geotherms eventually pass into the eclogite stability field, and the root zones of the basaltic plateaus invert to a dense eclogite mineral assemblage that acts as a gravitational anchor pulling the plateaus into the mantle. As they sink, the wet plateau roots undergo partial melting, producing tonalitic magmas which rise, intruding the overlying basalts and forming small continents that resist subduction. If the transition to eclogite occurred over a relatively short time interval in the Late Archean, tremendous volumes of basalt could have been pulled into the mantle, producing large quantities of tonalitic magma. The net result

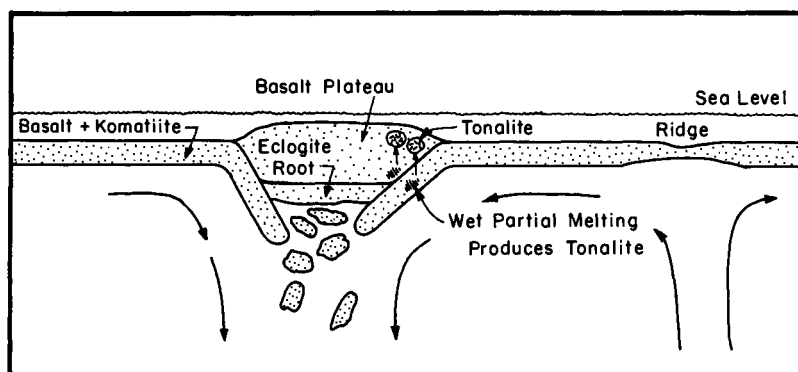


FIG 10.9 Schematic cross-section showing development of eclogite roots beneath basaltic plateaus during the Early Archean.

is catastrophic continental growth. The continental growth rate should decrease after the basaltic plateaus have descended into the mantle, perhaps at about 2.6 Ga. If 70% of the continental crust formed in the Late Archean, this requires subduction of a basaltic layer averaging 50 km thick and covering 75% of the Earth's surface.

Although the eclogite transition model is attractive to explain Late Archean continental growth, it only works once, and cannot explain the rapid Early Proterozoic growth of continents.

Early Cratons

Just when the first continental cratons appeared is not known. However, the quartzite–pelite–carbonate association, which characterizes cratons, is found in Archean high-grade terranes (Chapt. 9), the oldest known of which is ≥ 3.3 Ga in the Limpopo belt of South Africa. Early cratons were probably very small (≤ 500 km across) and relatively few in number, and it was not until after the rapid continental growth in the Late Archean that cratons became important.

What processes lead to the production of cratons? Evidence suggests that devolatilization of the lower crust and lithosphere and intracrustal melting may be important (Campbell and Jarvis, 1984; Pollack, 1986). Both of these processes redistribute LIL elements, including the heat-producing isotopes of U, Th, and K. Rising magmas and fluids transfer these elements from the lithosphere and lower crust into the upper crust. Following the rapid 2.7–3.0 Ga continental growth, high-K granitic magmas generated in the lower crust were intruded at shallow levels, principally between 2.4 and 2.6 Ga. The mantle lithosphere beneath continents also may be depleted in LIL elements by loss of volatiles, and probably plays an important role in the stabilization of cratons. Loss of volatiles from both the subcontinental lithosphere and lower crust raises their solidus temperatures and makes subsequent melting more difficult. Such losses also result in increased shear strength, enhancing the mechanical stability of both the crust and lithosphere. With time, devolatilization of the upper mantle becomes less efficient as viscosity increases with cooling, and the process of cratonization slows down.

The length of time necessary to form a craton varies from one region to another.

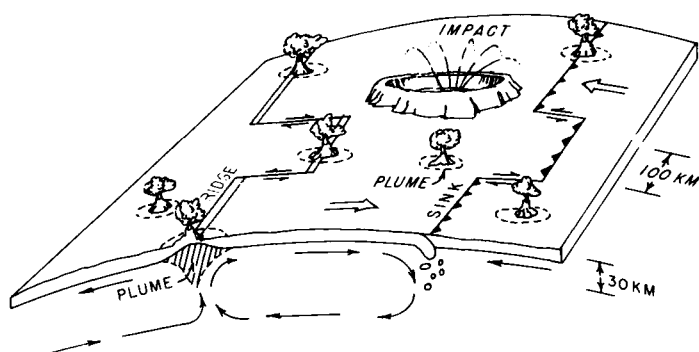


FIG 10.10 Schematic diagram showing early development of a primitive komatiitic terrestrial crust. After Condie (1980).

The emplacement of post-tectonic granites, which seems to be the first signal of the completion of cratonization, ranges from 10 to 100 Ma after deformation and syntectonic plutonism. Following emplacement of these granites by 200–400 Ma, widespread uplift occurs and erosion removes major topographic features. Derivative sediments are deposited in marginal and intracratonic basins, representing one of the last stages of cratonization.

A Model of Archean Crustal Development

A tentative model for the development of the crust and lithosphere is based on the following assumptions (after Condie, 1980, 1986): (1) the geothermal gradient in the Earth prior to 4.0 Ga was adiabatic; (2) heat production decreases with time; (3) mantle convection begins early, probably during the late stages of planetary accretion; (4) the first crust is komatiitic in composition; (5) the Earth is subjected to intense impact cratering until 3.9 Ga ago; and (6) continents are aggregated into a few supercontinents by 2.7 Ga, probably by collisions of microcontinent fragments.

Extensive melting occurs in the outer part of the mantle during the first 50 to 100 Ma of Earth history. During this time the core forms, an adiabatic geotherm is established, and the mantle is largely devolatilized, giving rise to the atmosphere and oceans. Loss of heat by radiation and volatile escape cools the surface of the Earth and a thin unstable komatiite crust forms at ocean ridges (fig. 10.10). As cooling continues, basaltic magmas contribute to this oceanic crust. The oceanic crust and lithosphere by 4.0 Ga are composed of an array of ridges and sinks, and plates are driven by the sinking of komatiite into the mantle. This early crust is rapidly recycled through the mantle and major impacting until about 3.9 Ga fragments the crust and aids in recycling. Some time between 4.0 and 4.2 Ga, geotherms at subduction zones decrease such that descending crust is only partially melted, producing tonalite magmas which rise to form the earliest continental crust as arc systems (fig. 10.11). Submarine basaltic plateaus, which tend to resist subduction, grow in size and eventually cover large portions of the Earth's surface. Cooling at sinks leads to eclogite stability at about 3 Ga and the roots of the basaltic plateaus invert to eclogite and catastrophically sink into the mantle. Wet partial melting occurs as they sink, and large volumes of tonalite rise, forming 50–70% of the

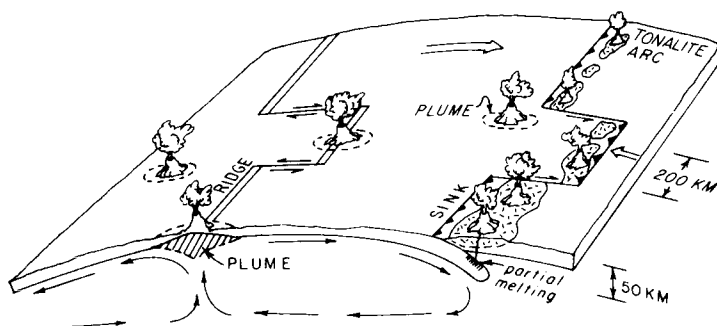


FIG 10.11 Schematic diagram showing development of early continental crust (tonalite arcs) at convergent plate margins. After Condie (1980).

present continental crust as tonalitic arcs. These arcs collide to form a relatively few large continents by 2.5 Ga. Partial melting of the lower crust at 2.4–2.6 Ga produces granites which are intruded into the dominantly tonalitic crust, signaling completion of the first large cratons.

North America: The Growth of a Continent

North America provides an example of the birth and growth of a continent through geologic time. As previously emphasized, continental growth can be considered in terms of crustal formation age and accretion age. The former is the time that continental fragments are extracted from the mantle, and the latter is the time of collision and accretion of fragments to make a continent. The crustal formation provinces of North America are shown in figure 10.12 from available Nd isotopic data and interpolation of these data (Nelson and DePaolo, 1985; Patchett and Arndt, 1986; Hoffman, 1988). These provinces, in turn, are composed of smaller accretionary terranes. It is important to remember that the present locations of many of these provinces are not the same as they were when the provinces first formed. Noteworthy is the small amount of continent formation prior to 3.5 Ga as represented by four small provinces. Also apparent from the map is the large amount of crust formed in the Late Archean, comprising a minimum of 50% of the continent. Approximately 30% of the continent appears to have formed in the Early Proterozoic, <10% in the Mid-to-Late Proterozoic, and <10% in the Phanerozoic.

Field and geophysical data from the Canadian Shield, as well as results from boreholes in platform sediment, indicate that North America is an amalgamation of plates, recently referred to as the “United Plates of America” (Hoffman, 1988). The Archean crust is composed of at least six separate provinces joined by Early Proterozoic orogenic belts (fig. 10.12). The systematic asymmetry of stratigraphic sections, structure, metamorphism, and igneous rocks is consistent with an origin by subduction and collision. Such asymmetry is particularly well displayed along the Trans Hudson, Labrador, and Penokean orogenic belts. In these belts, zones of foreland deformation are dominated by thrusts and recumbent folds, whereas hinterlands typically show transcurrent faults. Both features are characteristic of subduction zones. Some Proterozoic orogens have large accretionary prisms, whereas others do not. For instance, the Rae and Hearne Provinces involve only

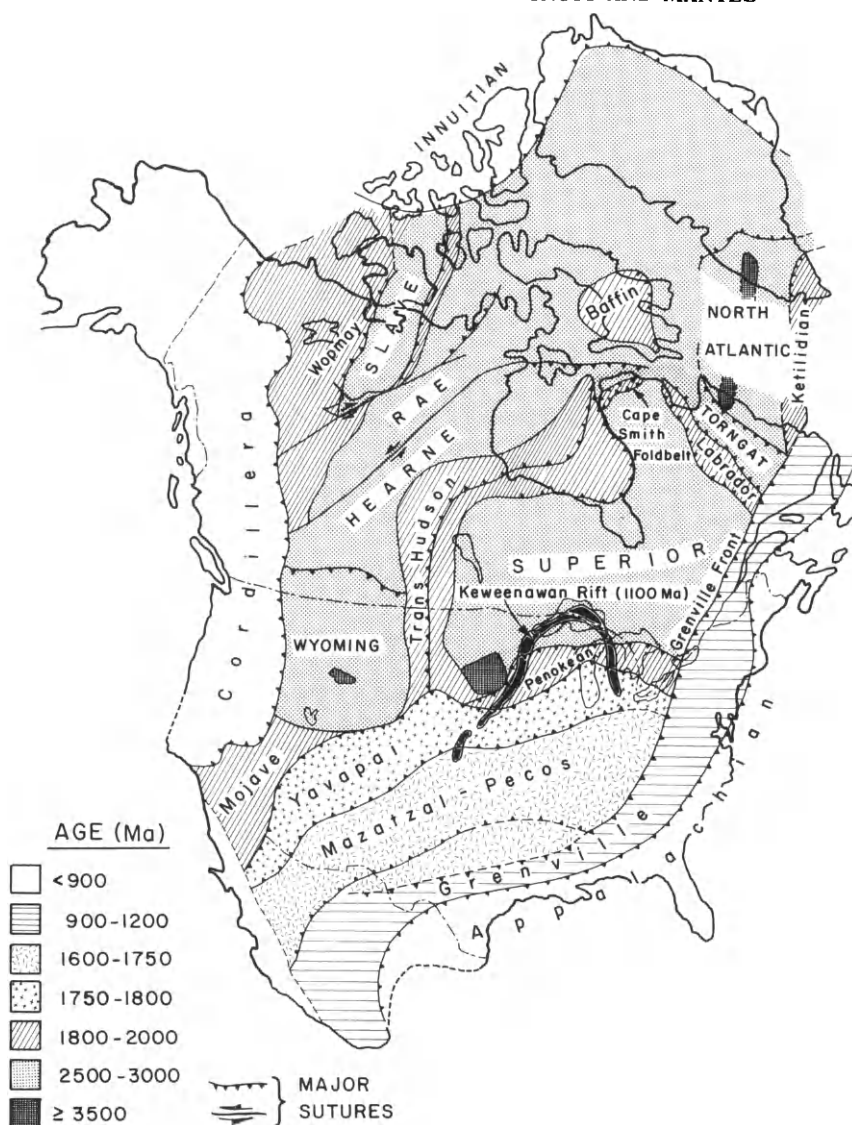


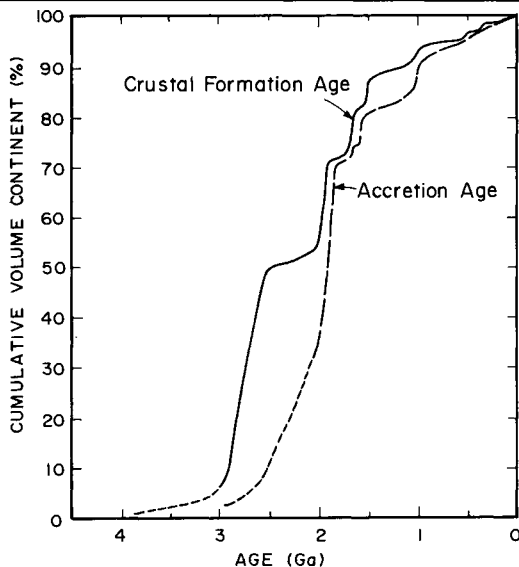
FIG 10.12 Distribution of major North American crustal provinces and sutures between provinces. Modified after Hoffman (1988). Movement restored on San Andreas fault and Late Mesozoic megashears in Mexico.

suturing of Archean crust, while the Trans Hudson, Penokean and Yavapai Provinces are accretionary zones up to 700 km in width between or adjacent to Archean cratons. The Phanerozoic Cordilleran and Appalachian Provinces appear to represent collages of accretionary terranes sutured along transform faults or, less commonly, along extinct subduction zones. Radiometric dates indicate that peak arc magmatism occurred during (or in a few cases following) rather than before the onset of collision, and thus provides a means of dating collisions.

The assembly of constituent Archean provinces took only about 100 Ma between 1850 and 1950 Ma. Estimated times of collision are summarized in table 10.2. The arc terranes of the Yavapai Province appear to have been accreted at about

TABLE 10.2 *Summary of Major Proterozoic Collisions in North America*

	Age (Ma)
Thelon	1950
Snowbird	1850–1920
Trans Hudson	1850
Penokean	1850
Makkovik	1850
Wopmay	1900
Labrador	1830–1880
Yavapai	1750–1780
Ouachita	1640 (?)

**FIG 10.13** Crustal formation and accretion growth rates for North America.

1750–1780 Ma. Other possible collisions are recorded at about 1710 Ma in Colorado and 1640 Ma in Oklahoma–West Texas. One or more collisions may have occurred in the Greenville Province at 900–1200 Ma.

A comparison of the rates of crustal formation and accretion for North America is given in figure 10.13. After the Early Proterozoic the shapes of the growth curves are similar. The accretion curve lags behind the crustal formation curves by generally <100 Ma. The greatest differences between these curves in the Late Archean and Early Proterozoic may not be real in that accretion rates are not well known prior to 2000 Ma. In particular, it is not clear if any or all of the Archean provinces were once part of the same supercontinent prior to possible fragmentation between 2000 and 2400 Ma.

Mantle Evolution

Mantle Reservoirs

As discussed in Chapter 7, Sr and Pb isotopic data seem to require a minimum of four, and perhaps several, ancient mantle sources for modern oceanic basalts. Scales

of these heterogeneities must range from a few meters to >10 km, and at least some heterogeneities must persist for billions of years (Davies, 1984). Several lines of geophysical evidence (see Chapt. 3) seem to indicate that plates penetrate well into the lower mantle, not favoring a convectively layered mantle. Both geochemical and geophysical constraints favor a “plum pudding” model for the mantle with the existence of heterogeneities of many scales and compositions at all depths, with perhaps a greater density of large heterogeneities in the lower mantle (fig. 7.19; Chapt. 7). Possible sources for mantle heterogeneities include recycled oceanic crust, subducted continental sediments, detached subcontinental lithosphere, and surviving primordial mantle.

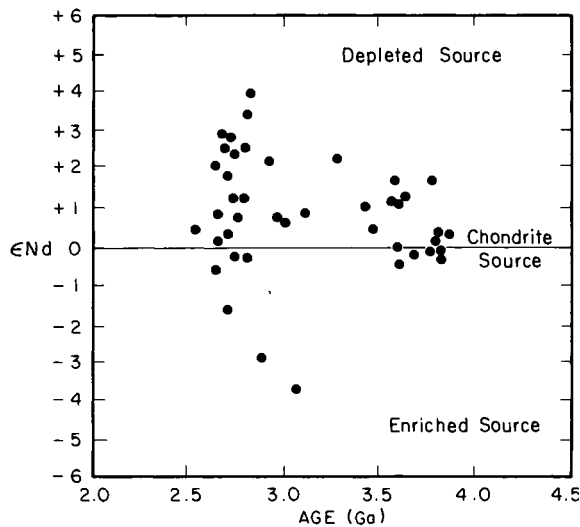
Survival of Mantle Heterogeneities

A major question in terms of mantle evolution is that of how mantle heterogeneities survive in a convecting Earth. Isotopic data discussed in Chapter 7 suggest that many heterogeneities have survived for over 10^9 a. Studies indicate that convection mixes efficiently on a scale of the size of convection cells, but that its efficiency on smaller scales depends on the number of overturns (Allegre et al., 1980). For instance, 10 overturns of the whole mantle, which requires 3 Ga at present rates, should mix down to a scale of hundreds of kilometers. Solid state diffusion, however, is very inefficient in mixing, and it will homogenize only over tens of centimeters to meters in billions of years. Mixing in the uppermost mantle due to the ascent of magmas may promote mixing over tens or a few hundred kilometers. Thus, mantle heterogeneities on the scales of meters to hundreds of kilometers should persist for billions of years in the absence of melting (Gurnis and Davies, 1986). With extensive melting, regions hundreds of kilometers across may be well mixed and yield homogeneous magmas, as for instance required by MORB sources.

Another factor important in the survival of heterogeneities is mantle strain rate. The average strain in the mantle over 3 Ga (10 overturns) appears to be about 100. Thus, a contaminant in the mantle would be reduced in its shortest dimension by a factor of 100 over this time. For mantle “plums” to retain coherency for billions of years they would have to reside at levels in the mantle where, on the average, strain is low. Alternately, the initial size of the plum must be large, perhaps involving a significant volume of the mantle, such as the Dupal anomaly (fig 7.16). Some relatively small heterogeneities may have survived for billions of years, but are strung out into long sheets or stringers. This is possible if parts of the mantle have very low flow velocities, such as near the centers of convection cells. The fact that some hotspots probably originate from regions with low horizontal flow velocities (since their velocities relative to each other are small) indicates that low-velocity flow regions exist in the mantle.

The Age of Depleted Mantle

It is commonly thought that the depleted mantle reservoir from which MORB are derived represents a residue left from extraction of the continents. Although other interpretations are possible, depleted mantle (DM) and continental crust are approximate geochemical complements of each other relative to primordial mantle if



Perhaps large quantities of LIL elements were enriched in the uppermost mantle at this time, leaving a voluminous DM reservoir behind (Condie, 1986). This early enriched mantle may have been the suboceanic lithosphere. If so, it would be rapidly recycled into the mantle introducing the earliest mantle "plums". Whether heterogeneities of this age exist in the Earth is an exciting question which can be tested only if some of these heterogeneities have served as magma sources for recent volcanics.

The Mantle Lithosphere

The base of the mantle lithosphere beneath oceans is probably a thermal boundary which deepens as a function of age (fig. 3.15). Suboceanic lithosphere like oceanic crust is transient in Earth history, continually recycled into the mantle. The subcontinental lithosphere, however, may be much older and capable of surviving for long periods of geologic time. The thickness of the continental lithosphere is poorly constrained, but it may extend to the 400-km discontinuity beneath Precambrian cratons (Chapt. 3). The age and thickness of the subcontinental lithosphere in southern Africa is constrained from diamonds in garnet lherzolite xenoliths formed in kimberlite pipes (Boyd and Gurney, 1986). Two lines of evidence suggest these xenoliths came from the lithosphere rather than the asthenosphere. First, the geographic distribution of pipes containing diamond-bearing xenoliths is limited to the Archean Kaapvaal craton, which appears to be underlain by a relatively thick mantle lithosphere. Second, and perhaps more important, is the fact that geothermometry studies indicate the xenoliths formed at temperatures $\leq 1100^{\circ}\text{C}$, which is less than the melting temperature of garnet lherzolite at asthenosphere pressures. Hence, they probably did not come from the partially melted asthenosphere. The most exciting aspects of these results are the Sm/Nd dates of minerals associated with the diamonds in the xenoliths, which indicate ages of over 3 Ga. From fossil geotherms reconstructed from xenolith geobarometry–geothermometry studies interpreted in conjunction with phase equilibria for diamonds, it appears that the subcontinental lithosphere beneath South Africa must have been 150–200 km thick by 3 Ga ago (Richter, 1985).

The idea of thick mantle lithosphere beneath the early continents is not new. With higher heat flow in the Archean it is difficult to avoid melting the lower parts of continents, and a thick lithosphere underplating has been called upon to act as a thermal buffer to prevent such melting (Davies, 1979). The moderate geotherms estimated in Archean continental crust (see Chapt. 9) are also consistent with such a thick lithosphere. The origin and survival of a thick subcontinental lithosphere presents another problem. One possibility is that this lithosphere is depleted mantle, produced at the same time as the overlying continent. Because depleted mantle is less dense than undepleted mantle, it should tend to ride high and resist mixing with the convecting asthenosphere. Just how this early lithosphere was produced in connection with continent formation is not clear, since early tonalites appear to have been derived from descending mafic crust as previously described. Perhaps it represents depleted portions of the mantle wedge that also may have contributed to continent formation. The final stages of lithosphere formation may have occurred during cratonization 50–200 Ma after major continental growth. During this time

large amounts of volatiles may have moved upwards into the crust, leaving a dry and relatively stable lithospheric root that persisted to the present time. The fact that continents have split and drifted apart at various times since the Archean, however, indicates that the lithospheric roots of continents do not resist fragmentation by convection in the mantle.

Secular Changes in Earth History

Introduction

Numerous geological and geochemical changes have been suggested to occur during the evolution of the Earth. Some are gradual changes, such as the recycling of sediments, while others may occur very abruptly and even result from catastrophic events, such as impact of asteroid-sized bodies on the Earth's surface. Major geochemical changes have been proposed to occur across the Archean–Proterozoic boundary reflecting rapid continental growth in the Late Archean (Taylor and McLennan, 1985). Excluding changes caused by extraterrestrial phenomena (which are discussed in Chapt. 11), most changes interpreted to occur in the geologic record reflect cooling of the Earth, recycling of rocks through the crust or mantle, or changes in composition of the Earth's atmosphere. Changes related to atmosphere composition will be discussed in Chapter 11.

To accurately identify geochemical or lithologic changes in Earth history it is necessary to compare rocks from similar tectonic settings to avoid ambiguities caused by inherent differences in tectonic setting (Condie, 1988). One possible approach to minimize the tectonic setting effect is to compare rocks from similar lithologic associations. For instance, compare pelites from the quartzite–pelite–carbonate association, or basalts from the greenstone association. Because many published compositional averages of specific rock types ignore the effect of tectonic setting, the significance of these analyses is uncertain, and interpretation of crustal or atmospheric evolution based on such data may be misleading.

Recycling

Many geological entities exhibit an exponential relationship with time (Veizer and Jansen, 1979). For instance, the distribution of igneous crystallization radiometric dates in the continental crust, the distribution of sediments (fig. 10.15) and the reserves of mineral deposits all show exponential increases in the time. Such exponentially increasing quantities may be caused by either: (1) an exponentially accelerating rate towards the present of a given process (such as magmatism, sedimentation, ore formation), or (2) recycling of a constant or growing geologic entity (such as continental crust or sediments). Although both factors may contribute to the exponential distributions, recycling seems to be the dominant factor in most instances. Especially not favoring accelerating rates with time is the fact that the Earth is cooling, which should slow down rates of magmatism, ore formation, and related phenomena. Also, radiometric dates do not favor accelerating processes in the last 1000 Ma. The exponential curves can be fit readily to recycling models, the faster the rate of recycling the steeper the slope of the age curves (Veizer and Jansen, 1979, 1985).

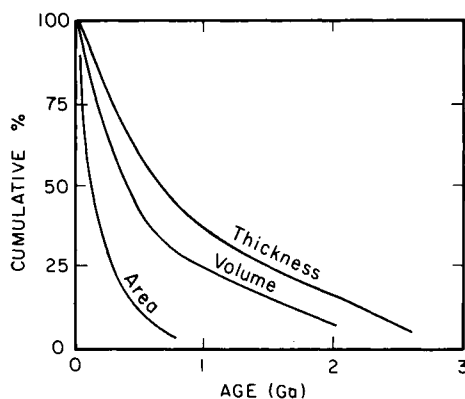


FIG 10.15 Present-day distribution of the preserved cumulative area, thickness, and volume of sedimentary rocks with age. Modified after Veizer and Jansen (1979).

From the distribution of radiometric igneous crystallization ages it is possible to calculate actual growth rates for the continental crust. Results indicate that very little continental crust existed before 3 Ga ago, and that the continents may have grown rapidly in the Late Archean (Veizer and Jensen, 1979), a conclusion consistent with other lines of evidence as previously discussed.

Early Proterozoic and Archean sediments crop out on a very small proportion ($<1\%$) of the Earth's surface and comprise $<10\%$ of the total volume of continental sediments (fig. 10.15). The increasing volume of sediments with time seems to reflect increasing rates of recycling, which vary with the growth of sediment area, thickness, and volume. To what extent sediment recycling is closed to additions and losses is difficult to evaluate. Nd isotopic results from sediments, however, suggest that 70–90% of the sediment distribution with time reflects closed-system recycling.

Komatiites

One of the striking differences between the Archean and post-Archean is the relative importance of komatiites in the Archean (Chapt. 9). Although most petrologists agree that this distribution reflects higher temperatures in the Archean mantle, where and how komatiites are produced remain unsolved questions. The high MgO contents of Archean komatiites (up to 32%) imply eruptive temperatures up to 1650°C (Campbell and Jarvis, 1984). This necessitates mantle temperatures $200\text{--}300^{\circ}\text{C}$ higher than at present. While ocean-ridge basalts appear to be produced at depths of 50–85 km, reflect 20–30% melting of garnet lherzolite and source temperatures of about 1400°C , komatiites may be produced at depths >200 km and reflect high degrees of melting ($>50\%$) and mantle temperatures $\geq 1800^{\circ}\text{C}$ (fig. 10.16). One of the problems with such high upper-mantle temperatures in the Archean comes from model studies of convection which suggest that Archean mantle temperatures $>200^{\circ}\text{C}$ more than at present produce chaotic convection that would completely disrupt and recycle Archean crust. Thus, little if any pre-2.5-Ga crust should be preserved. Because this is clearly not the case, temperatures $<200^{\circ}\text{C}$ hotter than at present for the Archean mantle are preferred. These temperatures are also consistent with the mounting evidence for a cool, relatively thick Archean

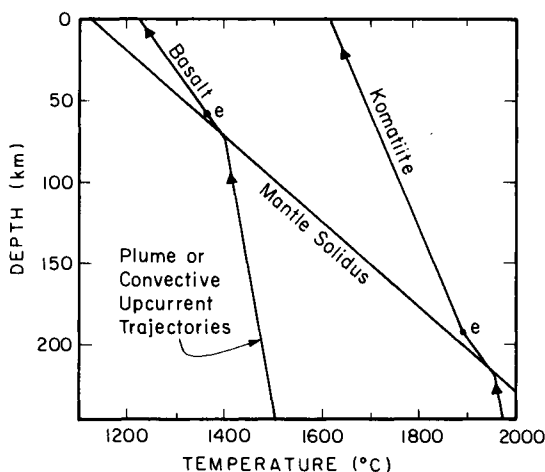


FIG 10.16 P-T relationships for the production of basalts and komatiites. Decrease in temperature at the mantle solidus reflects latent heat of melting; e, point at which magma escapes from sources.

subcontinental lithosphere as previously discussed. One possible way to have a somewhat cooler Archean mantle is if komatiites form by low rather than high degrees of melting. For lower degrees of melting (20–40%), the latent heat of melting is reduced and the temperature drop during magma production is minimized (100–200°C instead of 400°C) (fig. 10.16). The upper mantle temperature can be reduced even more if komatiites are produced in plumes rapidly rising from the lower mantle. Although komatiites must come from a hotter part of the Archean mantle than basalts, just where that is remains uncertain. They may be produced in hot “jets” or plumes that rapidly rise from the lower mantle. Alternately, or in addition to this source, komatiites may be produced in convective upcurrents beneath Archean ocean ridges. The komatiite trajectory in figure 10.16 is consistent with either source.

Sediments

Sr Isotopes in Marine Carbonates

It has been recognized for many years that the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in marine carbonates varies with age (fig. 10.17). Marine carbonates deposited in equilibrium with seawater inherit the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the seawater. Because this ratio is relatively constant in modern seawater, regardless of associated tectonic setting, it is not significantly affected by local sources. Hence, the Sr isotope composition of marine carbonates may monitor the composition of seawater with time. The chief factors controlling the Sr isotopic composition of the oceans are the composition of river waters and of hydrothermal springs on the seafloor. The current $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of seawater (0.709) represents about a 4:1 mixture of river water (0.711) and submarine volcanic water (0.703). Possible factors causing changes in the seawater isotopic ratio include: (1) variation in the composition of the continents exposed to weathering; (2) importance of volcanic activity; (3) rate of seafloor spreading which monitors the input of submarine volcanic Sr into the oceans; (4) extent of continent flooding by shallow seas, which is a measure of the amount of continent exposed to erosion; and

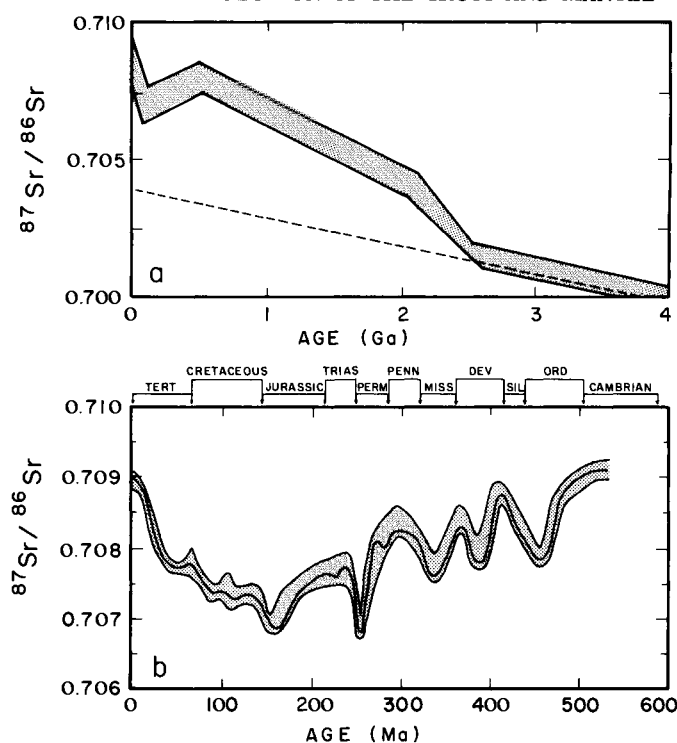


FIG 10.17 Variations in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of marine carbonates during the last 3.5 Ga (a) and during the last 600 Ma (b). After Veizer and Compston (1976) and Chaudhuri and Clauer (1986). Dashed line in (a) is mantle growth curve.

(5) variations in climate which affect the rate of Sr removal from the continents. Many of these factors, in turn, are related to plate tectonic processes.

Two scales of variation of Sr isotope ratios in marine carbonates are known: (1) long-term changes characterized by an increase in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio after 2.5 Ga, a decrease in the Paleozoic and an increase again in the Mesozoic–Cenozoic (fig. 10.17a); and (2) short-term ups and downs (50–100 Ma) resolved only in carbonates ≤ 600 Ma in age (fig. 10.17b). The gradual increase in $^{87}\text{Sr}/^{86}\text{Sr}$ ratio beginning in the Early Proterozoic appears to reflect rapid continental growth in Late Archean (Veizer and Compston, 1976) marking an increase in the input of continental Sr to the oceans. The decrease in Sr isotope ratios during the Paleozoic may be caused by the growth of Pangaea with a gradual reduction in the total coastline length as the continents aggregated. This would lead to a decrease in the flux of continent-derived Sr entering the oceans (Chaudhuri and Clauer, 1986). The opposite effect occurs during the fragmentation of Pangaea, beginning about 200 Ma ago, and may contribute to the rising $^{87}\text{Sr}/^{86}\text{Sr}$ curve after 150 Ma. Also contributing to this rise may be a decrease in both ocean ridge volumes and spreading rates during the Tertiary, both leading to regression of the seas.

The small-scale variations in the Sr isotope curve appear to be related to some combination of tectonic, climatic, and eustatic effects. The four Sr isotope minima in the Paleozoic correspond roughly to major orogenies. The reason for this correlation, however, is not currently understood.

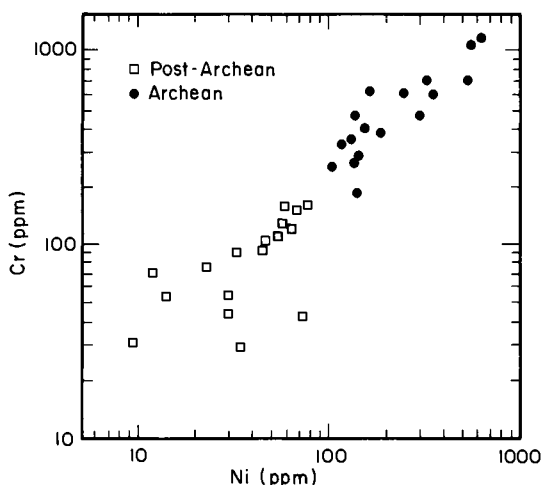


FIG 10.18 Cr–Ni distributions in average pelites of various ages. Data from both greenstone and quartzite–pelite–carbonate associations.

REE and Related Elements

Increases in the magnitude of the negative Eu anomaly and in La/Yb, Th/Sc, and Th/U ratios have been proposed to occur in pelites across the Archean–Proterozoic (A–P) boundary (Taylor and McLennan, 1985). These changes have been interpreted to reflect changes in the composition of the upper continental crust and in the growth rate of continents. However, when these geochemical changes are evaluated in terms of tectonic setting, some of them disappear and others decrease in magnitude (Condie, 1988). On the average, pelites from greenstone associations (GR pelites) have less pronounced Eu anomalies than pelites from quartzite–pelite–carbonate associations (QPC pelites). Increases in Eu anomaly and in La/Sc and Th/Sc ratios occur in both pelite groups across the boundary, although the changes in the QPC group may not be statistically real. The La/Yb ratio is virtually constant with time in both populations, and the Th/U ratio may actually decrease in the GR group across the A–P boundary. The large differences in element ratios reported to occur in pelites across the A–P boundary by Taylor and McLennan (1985) appear to reflect a tectonic setting bias. Their average Archean pelites are strongly weighted by GR pelites, whereas their post-Archean pelites are almost exclusively QP pelites. The changes in element ratios in GR pelites across the A–P boundary may reflect a greater importance of granites in post-Archean arcs, which in turn may be caused by a greater importance of fractional crystallization.

Ni and Cr in Pelites

An enrichment in Ni and Cr in Archean pelites relative to post-Archean pelites is well established (Condie, 1988) (fig. 10.18). A dramatic decrease in both of these elements occurs in crossing the A–P boundary. Several explanations have been proposed for this change: (1) the presence of komatiite in Archean sediment source areas; (2) intense chemical weathering of komatiite in source areas; and (3) enhanced adsorption of Ni and Cr on clay-sized particles in Archean seawater. Because it is

generally not possible to obtain mass balance with such elements as Ti, Zr, V, and Sc together with Ni, Cr, and Mg, it appears that the mere presence of komatiite in Archean sources is not a satisfactory explanation. Intense weathering of komatiites, however, may decouple these elements due to their different mobilities in weathering solutions. However, the fact that Ni–Cr enrichment occurs in greenstone pelites, as well as pelites from quartzite–pelite–carbonate (QP) associations presents a problem for the weathering model. This is because greenstone pelites are often integral parts of graywacke turbidite successions, and are derived in large part from active volcanic sources not subjected to intense weathering.

Another possibility that could concentrate Ni–Cr is the scavenging of these elements by clay-sized particles from seawater. This would require a greater concentration of these elements in the oceans than found today. A possible source for increased input of Ni and Cr into the sea is from hydrothermal leaching of oceanic crust in a manner similar to that proposed to occur at modern ocean ridges (Chapt. 7). The probable importance of submarine volcanism, and an oceanic crust composed largely of komatiite in the Archean, are consistent with such model. A komatiitic crust would provide a large, continuously replenished source for Ni and Cr. These elements could be leached from komatiites by hydrothermal waters at ocean ridges and enter the Archean oceans. If this mechanism is to explain the high Cr and Ni contents in Archean QP pelites, it also necessitates a relatively short ocean mixing time in the Archean compared to the residence times of Cr and Ni in seawater, in order for these elements to be transported into shallow cratonic basins where QP associations are deposited.

The Composition of Continents

Two approaches have been used to see if the composition of the continents has changed with time. The first makes use of widespread sampling of continental crust of various ages and levels of exposure (Chapt. 4) and the second employs the composition of pelites from cratonic associations (Taylor and McLennan, 1985). Elements that are relatively immobile yet have short residence times in seawater may be transferred almost in total from continental sources to derivative fine grained sediments. Such elements include Th, Zr, Sc, Co, and REE. The greatest uncertainty in estimating continental compositions is the composition of the lower crust.

Using both of these approaches, estimates of the composition of the upper continental crust of Archean and post-Archean age are summarized in table 10.3. Compared to Archean upper continental crust, post-Archean upper crust is enriched in alkali elements (K, Rb) and somewhat in other LIL elements (La, Th, U), and depleted in Ni, Co, Zr, and Nb. Concentrations of other elements, however, are very similar between the two ages of crust.

A possible explanation for these differences is that of progressively increasing vertical compositional zonation of the continental crust with time. This zonation is produced by metamorphism and melting in the lower crust, which results in upward migration of elements like K, U, and Th either in magmas or in an aqueous phase, leaving a depleted lower crust composed of granulites. This zonation, which is more pronounced in the Proterozoic than in the Archean, may reflect multiple reworking in mobile belts.

TABLE 10.3 *Average Chemical Composition of Archean and Post-Archean Continental Crust*

	Archean	Post-Archean
SiO ₂	65.5	66.0
Al ₂ O ₃	16.0	16.0
FeO	5.0	4.5
MgO	2.5	2.3
CaO	3.4	3.5
Na ₂ O	3.6	3.8
K ₂ O	2.7	3.3
TiO ₂	0.5	0.6
Rb	50	100
Ba	750	700
Sr	400	350
Ni	50	20
Co	25	10
Sc	13	11
Th	8	11
U	2	3
La	25	30
Yb	2	2
Zr	125	200
Nb	12	25

Oxides in wt%, trace elements in ppm. After Eade and Fahrig (1971) and Taylor and McLennan (1985).

TABLE 10.4 *Geochemical Classification of Basalts from Precambrian Greenstone Successions*

	≥ 3500 Ma		2500–3500 Ma		1500–2500 Ma		500–1500 Ma	
	<i>n</i>	%	<i>n</i>	%	<i>n</i>	%	<i>n</i>	%
IAB	24	46	100	68	86	31	9	20
CABI	17	33	30	20	90	33	25	54
CABC	8	15	8	5	90	33	10	22
NMORB	3	6	10	7	5	2	2	4
WPB	—	—	—	—	3	1	—	—
Total	52		148		274		46	

From Condie (1988). WPB, within-plate basalt; NMORB, normal ocean-ridge basalt; IAB, island arc basalt; CABI, calc-alkaline basalt from island arcs; CABC, calc-alkaline basalt from continental-margin arcs; *n*, number of samples.

Basalts

Differences in the composition of basalts from greenstone associations are apparent when compared as a function of age (Condie, 1988). When classified according to the ratios of relatively immobile incompatible elements (such as Th/Ta, Ti/Y, Th/Yb, Zr/Y), five tectonic categories can be defined in terms of modern basalts. Two major conclusions emerge when ancient greenstone basalts are considered in terms of these five categories (table 10.4): (1) basalts that have within-plate (WPB) or NMORB geochemical characteristics are rare to absent in all Precambrian categories; and (2)

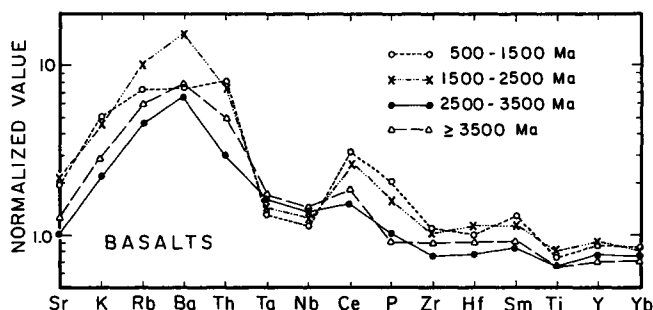


FIG 10.19 NMORB-normalized incompatible element distributions in basalts from Precambrian greenstone associations. From Condie (1988). Explanation of NMORB-normalized graphs given in Chapter 7.

in terms of incompatible element distributions, both Archean and Proterozoic greenstone basalts exhibit subduction zone geochemical signatures (fig. 10.19). Basalts with IAB affinities greatly dominate in the Late Archean. The absence or sparsity of basalts with WPB or NMORB geochemical characteristics from Precambrian greenstone successions of all ages supports inferences from lithologic associations (Chapt. 9), indicating that most greenstone successions do not represent remnants of oceanic crust or of continental rift assemblages, but that they are subduction-related.

Differences in absolute incompatible element concentrations between Archean and post-Archean greenstone basalts appear to reflect mantle source differences. The relatively enriched mantle sources implied by post-Archean basalts may be due to continental crust ($\pm$ lithosphere) being recycled into the mantle after 2.5 Ga. This explanation is consistent with rapid continental growth in the Late Archean. If differences between incompatible element distributions in Early and Late Archean basalts are statistically real, the lower LIL element contents and lower Th/Nb, Th/Ta, and La/Yb ratios of Late Archean basalts suggest that mantle sources were more depleted in the Late Archean than in the Early Archean (Condie, 1988). Greenstone basalts bracket the time interval (2.7–3.0 Ga) that a significant proportion of continental crust may have been extracted from the mantle, and the compositional differences between Early and Late Archean basalts may monitor this extraction.

Felsic Volcanics and Andesites

Archean felsic volcanics and andesites and shallow-level plutonic counterparts have similar trace element distributions to those of rocks from modern arc systems, except that they are strikingly depleted in Y and heavy REE (fig. 10.4) (Martin, 1986). These differences can be explained if Archean felsic and intermediate magmas (hereafter FI magmas) are produced by partial melting of descending wet, mafic crust with amphibole or/and garnet left in the melting residue, whereas most post-Archean FI magmas are produced by fractional crystallization. Garnet and amphibole retain Y and heavy REE in the source of the Archean magmas, but do not play a role in the formation of the post-Archean magmas.

The differences between Archean and post-Archean magma production probably

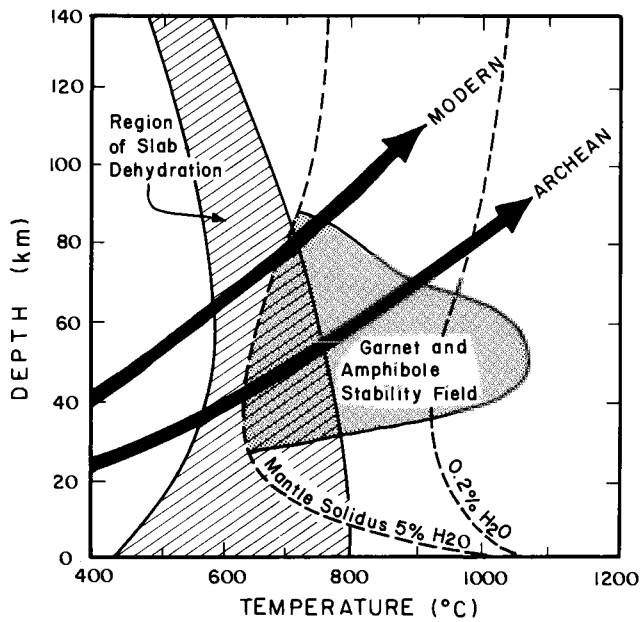


FIG 10.20 P-T diagram showing possible melting trajectories (bold arrows) in Archean and post-Archean subduction zones. Modified after Martin (1986). The 5% H₂O solidus applies to the Archean and the 0.2% H₂O solidus to the present time.

reflect cooling of the Earth (Martin, 1986). In the Archean, subducted oceanic crust may have been young (<30 Ma) and warm, and reached melting conditions during slab dehydration due to steeper subduction geotherms (fig. 10.20). Melting also occurred in the stability field of garnet and amphibole, and one or both of these phases were left in the residue. In contrast, modern subducted oceanic crust is old and cool and dehydrates before melting (see Chapt. 7). In this case, fluids released into the mantle wedge may promote melting, leading to the production of basalts that later undergo fractional crystallization to produce FI magmas. Thus, decreasing geotherms at convergent plate margins may have led to a change in the site of subduction-related magma production from the descending slab in the Archean to the mantle wedge thereafter.

Alkaline Igneous Rocks

Alkaline igneous rocks, such as trachytes, phonolites, basanites, kimberlites, and carbonatites, are characteristic of cratons as well as some continental rifts and oceanic islands. They do not, however, become important in the geologic record until after 200 Ma ago. Although alkaline igneous rocks are reported in Late Archean greenstones, they are extremely rare, and only a relatively small number of occurrences are known of Proterozoic and Paleozoic age. In part contributing to this distribution is the fact that continental alkaline igneous centers (volcanic and plutonic) are generally small and are removed readily during uplift and erosion. Hence, they have a low probability of survival. Most oceanic islands probably do not survive subduction. Another factor that is probably important in terms of the rarity

of Precambrian alkaline igneous rocks is cooling of the Earth. Production of alkaline mafic or ultramafic magmas in the upper mantle requires very small amounts of melting ($<10\%$), and prior to 2 Ga upper-mantle temperatures may have been too high for such small amounts of melting to occur.

Blueschists

Blueschists are formed in association with subduction and continental collision, and reflect burial to high pressures at relatively low temperatures. It has long been recognized that blueschists older than about 1000 Ma are apparently absent in the geologic record (Ernst, 1972). Those with aragonite and jadeitic clinopyroxene, which reflect the highest pressures, are confined to arc terranes <200 Ma in age. Three general ideas have been proposed for the absence of pre-1000 Ma blueschists: (1) steeper geotherms beneath pre-1000 Ma arcs prevented rocks from entering the blueschist P–T stability field (fig. 4.21a); (2) uplift of blueschists led to recrystallization to greenschist or amphibolite-facies mineral assemblages; and (3) erosion has removed old blueschists. It may be that all three of these factors contribute to the absence of pre-1000 Ma blueschists. Prior to 2000 Ma, steeper subduction geotherms may have prevented blueschist formation. After this time, however, when geotherms were not much steeper than at present, the second two factors may control blueschists preservation. Calculated P–T–time trajectories for blueschists suggest that they may increase in temperature prior to uplift (England and Richardson, 1977), resulting in recrystallization of blueschist-facies mineral assemblages to greenschist or amphibolite-facies mineral assemblages. After 500 Ma of uplift and erosion, only the latter two assemblages would be expected at the surface. Uplift and erosion after continental collisions may also remove blueschists. Even in young collisional mountain chains, such as the Himalayas, only few minor occurrences of blueschist have not been removed by erosion. Because almost all pre-1000 Ma orogenic belts are also collisional belts, blueschists may not be expected to have survived in these areas.

Mineral Deposits

As illustrated in figure 10.21, the distribution and importance of mineral deposits has varied through geologic time. Three factors seem to be chiefly responsible for this distribution: cooling of the Earth, the relative importance of various tectonic settings, and changes in the composition of the atmosphere (Hutchinson, 1981). Other factors, such as uplift and erosion and the amount of organic matter present, also contribute to the distribution of specific deposits.

Magmatic deposits (Cr, Pt, Ni), although found throughout the geologic record, are most important in the Late Archean and Early Proterozoic when large layered igneous intrusions were emplaced. Most of these were emplaced during or just after the period of rapid continental growth in the Late Archean. Cu–Zn massive sulfides and vein deposits occur throughout the geologic record associated with greenstones (arcs, collisional orogens). Porphyry Cu, Mo, and Sn deposits are also associated with greenstones, yet are rare very prior to 1000 Ma and are only important in the last 100 Ma; shallow vein deposits (Hg, Sb) have a similar time distribution. Erosion level may be responsible for the sparsity of these deposits prior to 1000 Ma. Most

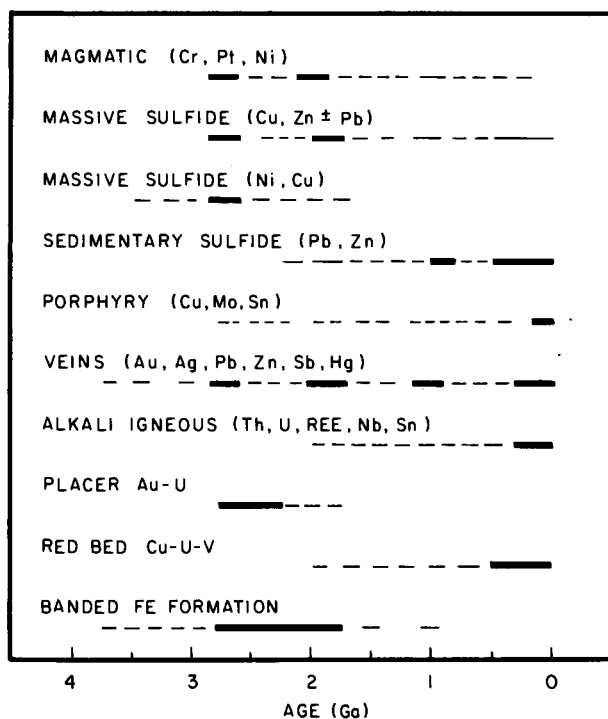


FIG 10.21 Distribution of major mineral deposits with geologic time.

Precambrian terranes are eroded to depths > 10 km, and porphyries and shallow veins form at depths less than this. Ni-Cu massive sulfides are limited to the Archean, as they were formed in association with komatiite magmas. The rather sudden appearances of sedimentary sulfides and alkaline igneous-related deposits (craton or rift-related) at 2 Ga reflects the availability of cratons following rapid growth of the continents in the Late Archean.

The narrow time span in the Early Proterozoic to which placer Au-U deposits are virtually confined has long puzzled geologists. It would appear that several factors contribute to this distribution. The availability of greenstone-granite sources for the Au and U, as well as widespread cratons after the end of the Archean, seem to be the two most important factors. Also, the preservation of detrital uraninite requires an oxygen-poor atmosphere (Chapt. 11). Redbed deposits and banded iron formation also require cratonic basins, which did not become widespread until in the Early Proterozoic. The abundant deposition of iron formation in the Late Archean and Early Proterozoic, and rapid decrease in importance thereafter, reflects an interplay of several factors including oxygen level of the atmosphere and ferrous iron content of seawater. The deposition of redbed Cu, U, and V is closely tied to decaying organic matter in buried terrestrial sediments, and the importance of these deposits in the Late Phanerozoic is related to the increasing abundance of terrestrial life forms.

The Archean-Proterozoic Boundary

Of the many changes that have been proposed to occur across the Archean-Proterozoic (A-P) boundary, only seven are well documented, as summarized in table 10.5.

TABLE 10.5 *Summary of Changes Across the Archean-Proterozoic Boundary*

Change	Immediate Cause
1. Increase in $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of marine carbonates	Increased input of continent-derived Sr into the oceans
2. Increase in range of incompatible element contents and of initial Nd, Pb, and Sr isotope ratios of greenstone basalts	Recycling of continental crust or lithosphere into mantle
3. Increase in sediment recycling rates	Increase in volume of continents
4. Apparent increase in the importance of rifts and cratons	Increase in volume of continents
5. Decrease in komatiites	Falling geotherms in the mantle
6. Decrease in Cr-Ni contents of pelites	Decrease in komatiite in continental or/and oceanic crust
7. Increase in heavy REE and Y contents of felsic and andesitic igneous rocks	Falling geotherms at subduction zones

Although immediate causes of these changes vary, the ultimate cause of all changes is cooling of the Earth. The fact that they coincide with the end of the Archean reflects, in some instances, the rapid growth of continental crust in the Late Archean (nos. 1–4). The last three can be related directly to falling geotherms. Note that none of the changes is abrupt, with most of them spread over several hundred million years, which is consistent with the Earth's cooling history (fig. 2.9). Whether the style of plate tectonics also changed during this time is not yet clear. Available data do not demand such a change.

Episodicity of Orogeny

Views differ as to whether orogeny is episodic, and if major peaks in orogeny are worldwide. Gastil (1960) compiled radiometric dates of various types from worldwide locations and suggested that orogeny is episodic with individual orogenies lasting 175–250 Ma and separated by gaps of 350–500 Ma. Based on a much larger number of radiometric dates, Dearnley (1965) suggested peaks in orogeny at 2750, 1950, 1075, 650, and 180 Ma. Possible correlations have been suggested between peaks in orogeny and seafloor spreading rates. Results for the Cenozoic suggest that major periods of collisional orogeny are either independent of rates of seafloor spreading or correlate with slow spreading (Schwan, 1985). Major, apparently worldwide orogenies at about 45 and 37 Ma seem to be bracketed by periods of rapid seafloor spreading. Several investigators have pointed out that loops in APW paths seem to correspond to times of major orogeny. Examples are loops in North American APW paths at 1850, 1750, and 1150 Ma (fig. 9.26). Such loops may be caused by changes in plate motions occurring in response to widespread continental collisions.

The problem of identifying episodicity of orogeny is a significant one. First of all, one is faced with resolution of dating methods. Until recently it was not possible to obtain a resolution better than 20–30 Ma in the Precambrian, yet Phanerozoic dates suggest that some orogenies are separated by time intervals <30 Ma. Only one method, the U–Pb zircon method, allows resolution of <10 Ma during the Precambrian, and hence to evaluate orogenic episodicity only zircon dates should be

TABLE 10.6 *Representative Durations of North American Orogenies*

Orogeny	Initiation (Ma)	Duration (Ma)
1. Laramide, W. USA	80	40
2. Nevadan, W. USA	160	30
3. Antler, W. USA	375	75
4. Taconic, E. USA	450	50
5. Green Mountain, Colorado	1750	50
6. Penokean, Wisconsin	1850	50
7. Wopmay, NW Canada	1900	<50
8. Grenville, NE Canada	1280	60
9. Batchawana greenstone belt, S. Ontario	2730	60
10. Shebandowan greenstone belt, SW Ontario	2730	50

employed. As discussed in Chapter 4, different isotopic systems have different blocking temperatures and record different thermal events. Hence, histograms of mixed Rb–Sr, U–Pb, and K–Ar dates are of little use in identifying orogenic episodicity. Still another problem is avoiding geographical bias. Often, many dates are available from a small geographic area and few or none from other areas. Thus, the amplitude of peaks on histograms of dates may not be representative of the lateral extent of a given orogeny. Also, orogeny may shift from one continent to another, and an orogenic gap on one continent may correspond to an orogenic maximum on another continent.

Figure 10.22 shows the distribution of published Precambrian U–Pb zircon dates. All dates shown are igneous crystallization dates, and they are grouped into three geographic areas (1, 2, and 3) reflecting, in part, probable continental connections during the Proterozoic (fig. 9.24). The largest number of dates is available from North America and the Baltic Shield (Scandinavia), and the smallest from South America and Antarctica. Several features are apparent from these distributions, as follows:

1. Although orogenic episodicity may characterize a given continent (or combination of continents), with two exceptions U–Pb zircon dates do not support worldwide episodicity.
2. When interpreted together with Nd isotopic results, two major worldwide periods of orogeny and crustal formation are recognized, at 2600–2800 Ma and 1700–1900 Ma.
3. When data from individual geographic areas are examined (table 10.6), it is clear that most orogenies have lasted about 50 Ma during the last 3 Ga. The broad peaks in zircon ages, for instance at 600–800 Ma in Africa–Australia and 1600–2000 in North America, represent several closely spaced orogenies, often separated geographically.
4. Orogenic gaps on one continent may be filled in by orogenies on another continent. For instance, the 2.0–2.6 Ga gap in North America is occupied by a 2.1–2.2 Ga event in South America–West Africa, and 2.5 Ga events in India and China. Also, the 1.0–1.6 and 0.5–0.9 Ga gaps in North America correspond to events in Africa, Australia, and Arabia. The so-called Lapalian

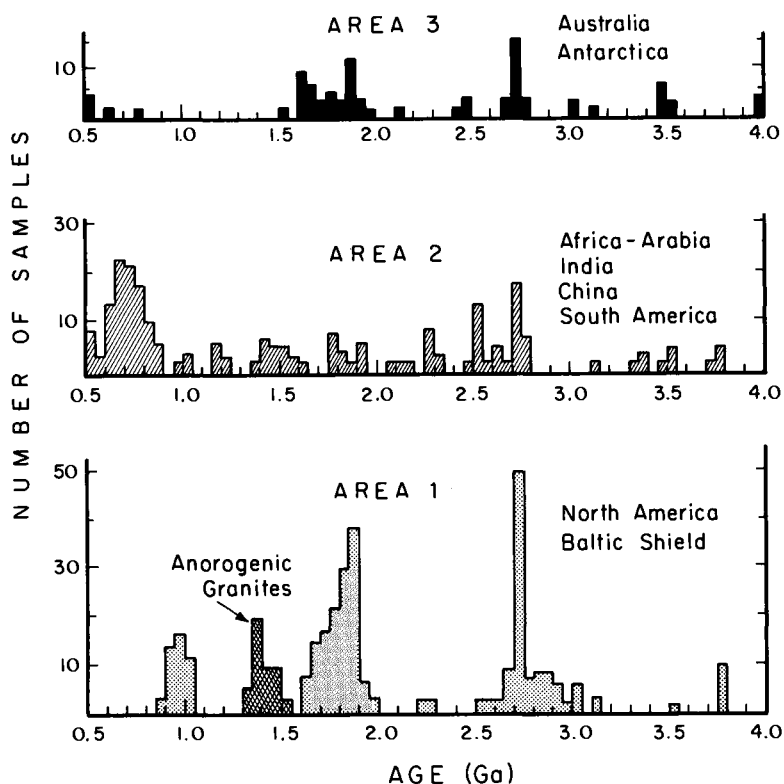


FIG 10.22 Frequency distribution of U-Pb zircon dates from Precambrian igneous rocks.

Interval (0.5–0.9 Ga) in North America corresponds to the widespread Pan-African orogeny in Africa and South America.

5. The period of widespread anorogenic granitic plutonism in North America (1.4–1.5 Ga) corresponds to major orogenies in Africa.

In terms of the Phanerozoic, it appears that major orogenies reflect continental collisions. Although there are some periods of time when widespread continental collisions coincide, there is no reason that this should commonly occur, and the distribution of zircon dates bears this out. With exception of the 2.8 and 1.9 Ga crustal formation peaks, worldwide synchronicity of orogeny is not favored by the distribution of U-Pb zircon dates.

Comparative Evolution of the Terrestrial Planets

It has long been known that the Earth is unique among the terrestrial planets. Not only is it a planet with oceans, an oxygen-bearing atmosphere, and living organisms, but it may be the only planet in which plate-tectonic processes are active. Why has the Earth undergone such a unique geologic history among the terrestrial planets? Although this is a complicated question which has no single answer, we can consider many of the variables involved.

TABLE 10.7 *Physical Properties of the Terrestrial Planets*

	Mass* (Earth = 1)	Radius (km)	Area/Mass*	Core/Mantle*	Density (gm/cm ³)	Distance from Sun (A.U.)
Mercury	0.0543	2440	2.5	~12	5.44	0.389
Venus	0.8137	6054	1.1	0.9	5.24	0.725
Earth	1	6370	1	1	5.52	1
Mars	0.1077	3390	2.5	0.8	3.94	1.53
Moon	0.0123	1740	6.1	0.12	3.34	1

*Relative to Earth.

The terrestrial planets, including the Moon, have similar densities (table 10.7) and probably similar bulk compositions. They are both evolving toward a stage of thermal and tectonic stability and quiescence as they cool. The rate at which a planet evolves is dependent upon a variety of factors which directly or indirectly control the loss of heat (Lowman, 1976; Schubert, 1979). First of all, the position of a planet in the solar system is important because it reflects the condensation sequence of elements from the cooling solar nebula (Chapt. 2). Of particular importance are the abundances of elements with short- and long-lived radionuclides (such as Al, U, Th, and K) which contribute to heating planetary bodies. The Moon contains considerably smaller amounts of U, Th, and K than the Earth, as is evidenced by geochemical studies of lunar samples. Analyses of fine-grained materials from the Viking landing sites suggest that Mars is significantly depleted in potassium relative to the Earth. Hence these bodies would have less radiogenic heating than the Earth. Planetary mass is important, in that the amount of accretional and gravitational energy is directly dependent upon mass. Planetary size is also important, in that greater area/mass ratios result in more rapid heat loss from planetary surfaces (table 10.7). Also important is the size of the iron core, in that much of initial planetary heat is provided during core formation (Chapt. 3). An exception may be Mercury, with the largest calculated core/mantle ratio. The implied high iron/silicate ratio for this planet may be due to loss of a large proportion of the silicate mantle during collision with an asteroid-sized body in the first 100 Ma of the planet's history, rather than to a large amount of iron in the inner solar nebula from which Mercury accreted. The volatile contents, and especially the water content of planetary mantles and the rate of volatile release, are important in controlling the amount of melting, fractional crystallization trends, and the viscosity of the planetary interiors which, in turn, affects the rate of convection. Convection appears to be the primary mechanism by which heat is lost during planetary evolution, and hence the rate of convection is important in terms of evolutionary state. It should be pointed out, however, that the only observational evidence that requires subsolidus convection today is from the Earth.

Planetary crusts are of three types. *Primary crust* forms during or immediately after planetary accretion by cooling at the surface. *Secondary crust* arises later in planetary history from partial melting of recycled primary crust, or from partial melting of planetary interiors. It is typically basaltic in composition. Examples include the lunar maria, the Earth's oceanic crust, and perhaps much of the crust on Mars and Venus. *Tertiary crust* is formed by partial melting and further differentia-

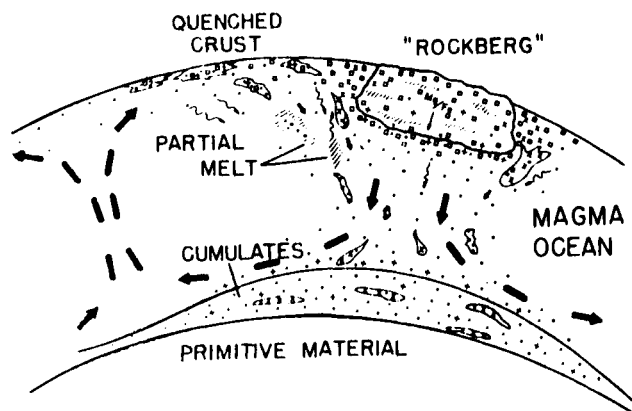


FIG 10.23 Schematic diagram illustrating the crystallization of a magma ocean on the Moon. After Longhi (1978). Arrows represent flow patterns and rockbergs are principally anorthosites.

tion of secondary crust, and the Earth's continents may be the only example of tertiary crust in the solar system.

Before considering comparative planetary evolution, it is important to review some of the important characteristics of each planet.

The Moon

A great deal has been learned about the geochemistry and geophysics of the Moon from the Apollo landings (Taylor, 1979, 1982). Models for the interior of the Moon indicate the presence of a primary crust 60–100 km thick composed chiefly of anorthosites and gabbroic anorthosites as represented by exposures in the lunar highlands. Mare basalts cover about 17% of the lunar surface but extend to depths of only up to 1 kilometer. Dated volcanics range from 3.2 to 3.9 Ga, and represent fluid flows, probably erupted chiefly from fissures. The lunar crust overlies a mantle composed of two layers. The upper layer or lithosphere extends to a depth of 400–500 km and is probably composed of cumulate ultramafic rocks. The second layer extends to about 1100 km, where a sharp break in S-wave velocities occurs. Evidence for a lunar core is inconclusive, and seismic and heat flow data limit convection to the deeper mantle and core regions. Radiometric dates indicate that the lunar highland crust is a primary crust that formed between 4.35 and 4.45 Ga. Impact continued on the lunar surface until about 3.9 Ga, when most of the large mare basins formed. The impacts that formed the mare basins did not initiate the melting that produced the basalts, which were erupted up to hundreds of millions of years later and represent secondary crust. The youngest eruptions appear to be 1 to 2 Ga in age.

The most popular model for lunar evolution involves the production of an ultramafic magma ocean that covers the entire Moon to depth of 500 km and crystallizes in ≤ 100 Ma beginning at about 4.45 Ga (fig. 10.23). Plagioclase floats, producing an anorthositic crust, and pyroxenes and olivine largely sink, producing an incompatible-element depleted upper mantle. Later partial melting of this mantle

produces the mare basalts. Detailed models for crystallization of the magma ocean indicate the process was complex, involving floating "rockbergs," cycles of assimilation, and mixing and trapping of residual liquids. The quenched surface and anorthositic rafts are also continually broken up by impact. It is possible, if not probable, that some form of plate tectonics was operative on the lunar surface at this time.

Mercury

The high density (5.45 gm/cm^3) of Mercury implies an Fe-Ni core that comprises about 66% of the planet's mass. Results from Mariner 10 show that the surface of Mercury is similar to that of the Moon in terms of crater distribution and probable age. Mercury differs from the Moon, however, in that it displays a global network of large lobate scarps. These scarps appear to represent high-angle reverse faults that developed during a period of global contraction related to continued cooling of the interior, and perhaps to core formation after crustal solidification. Although there is no unequivocal evidence for volcanism, the smooth plains are thought to represent widespread basaltic flows. The similarity of the highland-cratered terranes on Mercury to those of the Moon suggest a similar origin and age (Lowman, 1976). The faulting that produced the lobate scarps occurred both before and after the volcanism. In terms of timing, it is clear that the major magmatic events on Mercury occurred prior to the basin-forming events, which by analogy with the Moon, probably formed at about 3.9 Ga.

Mars

The Mariner and Viking missions to Mars indicate that it is quite different from the Moon and Mercury. The martian surface includes major shield volcanoes, fracture zones and rifts, as well as large canyons that appear to have been cut by running water. Also, much of the planet is covered with wind-blown deposits. More than half of the planet is covered by a variably cratered terrane similar to the surfaces of the Moon and Mercury. Large near-circular basins are similar to lunar mare basins and probably formed at about 3.9 Ga. The Tharsis region includes gigantic shield volcanoes and volcanic plains, and the Valles Marineris represents an enormous rift valley that spans one-quarter of the equator. The size of the volcanoes implies a very thick lithosphere. Studies of the surface morphology of lava flows on Mars indicate a total volume of lavas much greater than on the Earth and low viscosities of eruption. Major fracture systems on Mars and perhaps some volcanism appear to have remained active to between 1000 to 100 Ma ago. Chemical analyses from the Viking landing sites suggest the dominate volcanic rocks on Mars are Fe-rich basalts, and that weathering of the basalts occurs in a hydrous, oxidizing environment.

The geologic history of Mars was probably similar to the Moon and Mercury for the first few hundred million years, with major impact terminating with the basin-forming event at 3.9 Ga. Crystallization of a planet-wide magma ocean may have resulted in the formation of crust, mantle, and core prior to 4.4 Ga. Continued mantle convection resulted in extensive fracturing and rifting of the martian lithosphere and in widespread volcanism. The Tharsis uplift developed at this time,

perhaps in response to a large mantle plume (Mutch et al., 1976). The early atmosphere may have been very dense and rich in CO_2 and H_2O , and cooling led to extensive rainfall and erosion. With a drop in surface temperature permafrost formed, which was locally and perhaps catastrophically melted and caused massive floods that cut the large canyons (Squyres, 1984). The major erosional events probably occurred during or just before formation of the cratered plains of volcanic origin (i.e. prior to 3.9 Ga). Continued fracturing and volcanism on Mars extended to at least 1000 Ma (and perhaps 100 Ma), yet plate tectonics ceased early, due perhaps to the presence of a thick martian lithosphere.

Venus

Unlike the other terrestrial planets, Venus is similar to the Earth both in size (6054 and 6370 km, respectively) and mean density (5.24 and 5.52 gm/cm^3 , respectively). Both planets also have similar amounts of N_2 and CO_2 , although most of the Earth's CO_2 is not in the atmosphere. Venus differs from the Earth by the density and composition of its atmosphere and its high atmospheric and surface temperatures. These differences, together with the slightly lower bulk density of Venus, affect the nature and rates of surface processes (weathering, erosion, deposition) and of tectonic and volcanic processes. A planet's thermal and tectonic history is dependent upon its size and area/mass ratio, as described above. Hence, Venus and Earth are expected to have similar histories. However, the surface features of Venus are quite different from those of the Earth, raising questions about how Venus transfers heat to the surface and whether plate tectonics is active.

Much has been learned about the surface of Venus from spacecraft missions (Venera, Vega, and Pioneer-Venus) and from Earth-based radar observatories. Results suggest that the majority of the venusian surface is composed of blocky bedrock surfaces and less than one-fourth contains porous soil-like material (McGill et al., 1983). The distribution of the soil-like deposits does not support the presence of large areas of impact-produced regolith. The Russian Venera spacecrafts that landed on the venusian surface have revealed the presence of abundant volcanic features, complex tectonic deformation, unusual large ovoidal features of probable volcanic-tectonic origin, and an impact crater density, suggesting an age of the crust for the northern part of Venus of 0.5–1.0 Ga. Reflectance studies of the venusian surface suggest that iron oxides may be important. Partial chemical analyses made by Venera landers indicate that basalt is the most important rock type. The high K_2O recorded by Venera 8 (~6%), however, may reflect a granitic terrane.

Many volcanic craters have been described on the surface of Venus, with some volcanoes having up to 10 km relief. The question of whether any active volcanoes exist on Venus remains unanswered. The interpretation of electric field observations as lightning produced by volcanic eruptions is not definitive. These perturbations in the electric field may be caused by local plasma instabilities in the upper atmosphere. One of the most interesting topographic features on Venus is a large trough between Beta Regio and Phoebe Regio. This trough is interpreted as a rift system comparable in size and complexity to the East African rift system (fig. 10.24). It is possible that the Beta Regio rift system is part of a circum-global venusian rift system (Phillips and Malin, 1984). Areas interpreted as major rift systems on Venus are concentrated

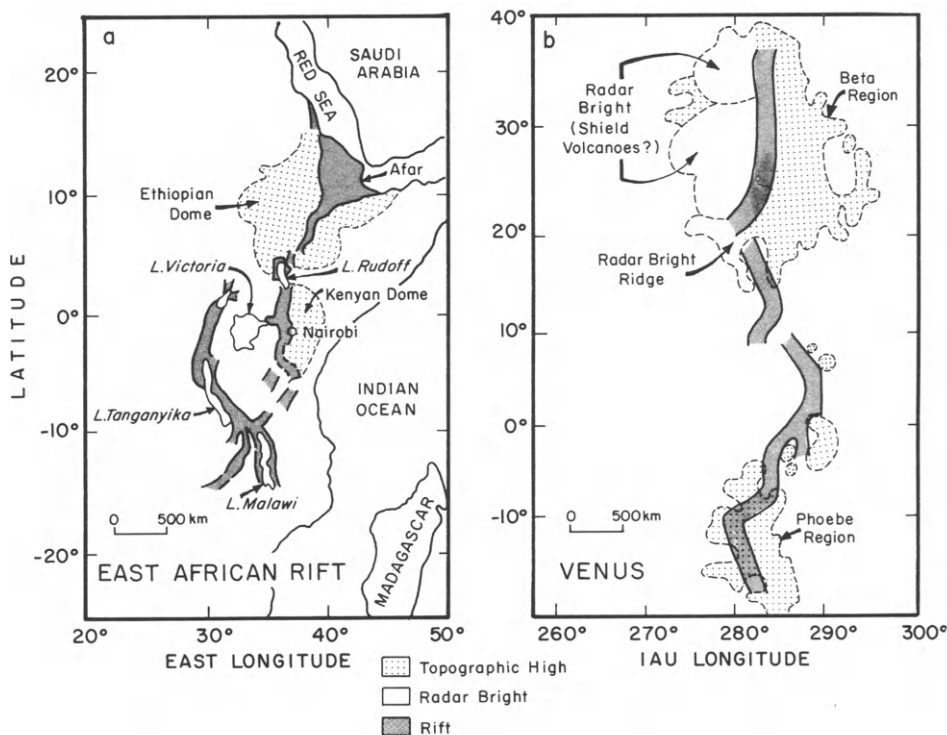


FIG 10.24 Comparison of the East African rift system with a proposed rift system on Venus. Modified after Phillips and Malin (1984).

in the equatorial highlands and also appear to represent the most thermally active areas on the planet. One of the exciting discoveries on Venus is the possibility of compressional mountain ranges as revealed by Earth-based radar profiles. All of these ranges are associated with Ishtar Terra, and range up to a few hundred kilometers in width and several thousand kilometers in length.

In terms of tectonic models for Venus, four sources of information are important: lithosphere thickness, topography, gravity anomalies, and the amount of ^{40}Ar in the atmosphere. If global heat loss on Venus is entirely by lithospheric conduction, then the average geotherm can be estimated from the average surface heat flow of 74 mW/m^2 . This heat flow is about twice the average ocean basin heat flow on the Earth and results in a steeper venusian geotherm (fig. 10.25). The base of thermal lithosphere in ocean basins is about 150 km (where the geotherm intersects the wet mantle solidus) compared to 40 km on Venus (with reference to the dry solidus). Assuming a wet upper-mantle on Venus makes little differences in the depth at which the geotherm intersects the mantle solidus. The base of the elastic lithosphere in ocean basins is about at the $500 \pm 150^\circ\text{C}$ isotherm or 50 km depth. This temperature corresponds to an elastic lithosphere thickness on Venus of $\leq 10 \text{ km}$, and hence both the thermal and elastic lithospheres of Venus are considerably thinner than those of the Earth's ocean basins.

Not counting the Earth's oceans, ranges in elevation on Venus are somewhat greater than on Earth. Of the total surface, 84% is composed of flat rolling plains

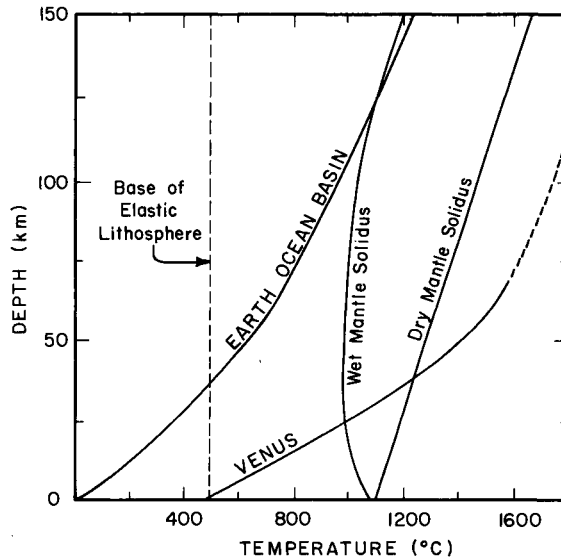


FIG 10.25 Comparison of average ocean basin geotherm on the Earth with calculated venusian geotherm. For Venus conduction is assumed the only mode of lithospheric heat transfer. Also shown are wet (0.1% H_2O) and dry mantle solidi.

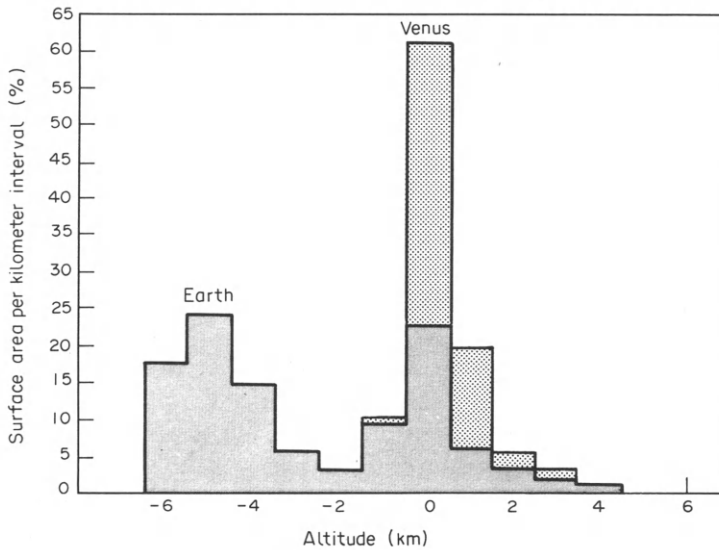


FIG 10.26 Comparison of relief on Venus and the Earth. After Pettengill et al. (1980). Surface height is plotted in 1 km intervals as a function of surface area. Height is measured from a sphere of average planetary radius for Venus and from sea level for the Earth. Copyright 1980 by Scientific American, Inc., all rights reserved.

some of which are over 1 km above the average plain elevation. Only 8% of the surface comprises true highlands, and the remainder (16%) lies below the average radius forming broad shallow basins. The large plateau-like areas might be similar to Earth's continents and the basins similar to ocean basins. The topographic distribution on the Earth, however, is bimodal due to plate tectonics which is responsible for ocean basins and the continents (fig. 10.26). Although the unimodal

distribution of elevation on Venus has been used as evidence against plate tectonics, two uncertainties weaken this argument: (1) if Venus has plate tectonics, it would exhibit a bimodal elevation distribution like the Earth only if it had comparable volumes of continental and oceanic crust; and (2) the presence of oceans on the Earth contributes to the bimodal elevation distribution. If the terrestrial oceans were to be removed, base level of the continents would be lowered and sediments would be carried into the ocean basins, leading perhaps to a unimodal elevation distribution. Major gravity anomalies on Venus correspond to major topographic features indicating that isostatic compensation depths are great (100–1000 km). The amount of ^{40}Ar in planetary atmospheres can be used as a rough index of past tectonic and volcanic activity because it is produced in planetary interiors by radioactive decay and requires tectonic–volcanic processes to escape. Venus has about one-third as much ^{40}Ar in its atmosphere as does the Earth, and implies less tectonic and volcanic activity for comparable ^{40}K contents.

Three thermal-tectonic models have been proposed for Venus: *conduction*, *hotspot*, and *plate tectonic models* (Phillips and Malin, 1984). In the conduction model, Venus loses heat by simple conduction through the thin lithosphere, and tectonics is a result of compression and tension in the lithosphere in response to the changing thermal state of the planet. It is likely that such a model describes the Moon, Mercury, and Mars at present. Not favoring such a model for Venus is the implication that the topography is very young (since it cannot be supported for long with warm, thin lithosphere) and the radar imagery which suggests rift valleys and compressional mountain chains. In the hotspot model, Venus is assumed to lose heat from large mantle plumes, which are also responsible for the large volcanoes observed on the surface. Thermal uplift associated with plumes can also lead to rifting. The deep levels of isostatic gravity compensation beneath some large topographic features are also consistent with large mantle plumes beneath these features.

The probable existence of major rift systems and possible existence of compressive mountain chains suggest that some form of plate tectonics may exist on Venus. The search for evidence of seafloor spreading on Venus has made use of Pioneer-Venus altimetry data to construct topographic profiles to compare with the shapes of spreading ridges on the Earth. In general, venusian ridges differ from ocean ridges in two ways: (1) they do not form a mode in crest height above the median plains, and (2) they do not form an extensive interconnected system as do ocean ridges on Earth. Thermal modeling suggests, however, that venusian spreading ridges should be topographically less pronounced than on the Earth, as well as being partially disguised by sediments. Recent studies of Aphrodite Terra suggest the existence of transform faults cutting symmetrical ridges. On the whole, data seem most consistent with a combined hotspot-plate tectonic model for Venus.

Another relevant question regarding venusian tectonics is whether continents exist. Ishtar Terra is the chief candidate for a continent because of its resemblance to terrestrial continents. It is plateau-like, and an arcuate ridge parallels the southwestern border much like an island arc. It also contains two linear mountain chains suggestive of collisional origins. Whether Ishtar Terra is a young or old feature is not yet known. The absence of water on Venus today does not necessarily argue against the formation of continents. A high D/H ratio in atmospheric water vapor on Venus

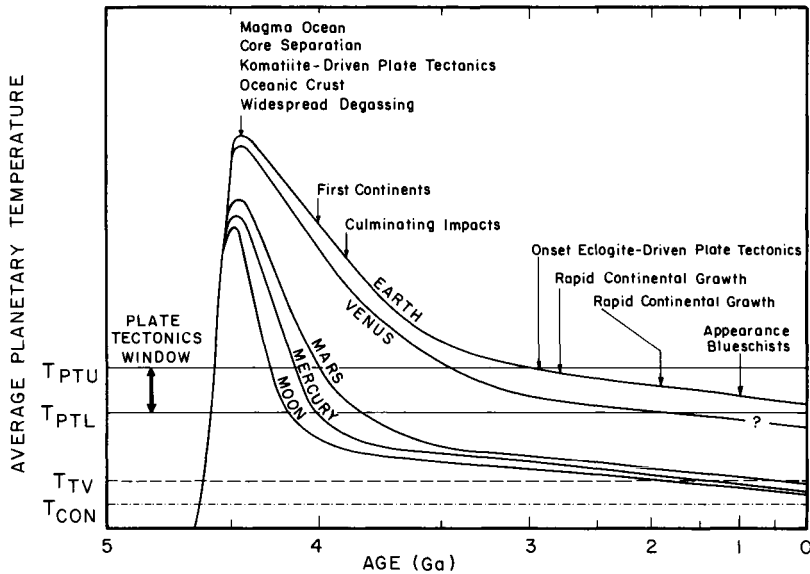


FIG 10.27 Schematic variation of average terrestrial planetary temperature with time. Threshold temperatures for major processes are indicated as follows: T_{PTU} and T_{PTL} , upper and lower temperature for eclogite-driven plate tectonics, respectively; T_{TV} , terminal volcanic temperature; T_{CON} , terminal subsolidus convection temperature.

implies that an early ocean may have existed before being boiled off by a runaway greenhouse effect (see Chapt. 11). During this boil-off, lighter hydrogen would be lost relative to deuterium, resulting in a high D/H ratio in the residual atmosphere.

The intriguing questions of whether plate tectonics is active on Venus and whether continents exist must await high-quality data from the venusian surface.

Comparative Planetary Histories

Models for the evolution of the terrestrial planets fall into two categories: (1) those that propose that all planets evolve through the same stages at different rates, and (2) models that require different histories for each planet. Although in the long run the second group of models may turn out to be more correct, data suggest that differences in the evolutionary histories of the terrestrial planets can be explained primarily by differences in heat productivities, volatile-element contents and cooling rates. One possible scheme of events is illustrated in figure 10.27 as a function of average planetary temperature. The temperature scale is schematic. All terrestrial planets undergo rapid heating during late stages of accretion, reaching maxima at about 4.5 Ga. Heat liberated primarily from core formation produces a widespread magma ocean and extensive degassing of the upper mantles. A thin komatiitic crust forms and is rapidly recycled into the magma ocean. As cooling progresses, the magma oceans crystallize in <100 Ma. In H_2O -rich planets, such as the Earth and probably Venus, komatiitic crust continues to form and is recycled by convection and surface impact. Each of the terrestrial planets may have had some form of komatiite-driven plate tectonics prior to about 4.3 Ga. Rapid cooling of all planets except Earth and Venus, which had large amounts of initial heat, led to rapidly

thickened lithospheres and to an end of plate tectonics by 4 Ga. Eclogite-driven plate tectonics may have begun on Earth between 3 and 3.5 Ga ago as it passed relatively slowly into the eclogite-driven plate tectonics thermal window (T_{PTU} , fig 10.27). Although Venus also should pass slowly through this window, higher temperatures prevented significant amounts of eclogite from forming, and argue against eclogite-driven plate tectonics on Venus. The other planets, with their thick lithospheres and rapid cooling rates, passed through this eclogite window relatively fast without initiation of plate tectonics. The model predicts that Earth should pass through the lower threshold for plate tectonics about 500 Ma in the future. The first continents appeared on Earth and perhaps on Venus at about 4 Ga and catastrophic eclogite subduction at 2.7–3.0 Ga led to rapid continental growth on the Earth. The temperature of terminal volcanism is reached first by the smallest body, the Moon, between 1 and 2 Ga ago, by Mars between 0.3 and 1.0 Ga and by Venus and the Earth some time in the future. Only the Moon, Mercury, and Mars are near or have passed the threshold of subsolidus convection.

Summary Statements

1. The earliest crust on the terrestrial planets was probably komatiitic in composition and formed during the period of 4.4–4.5 Ga, perhaps on the surface of magma oceans. As planets cooled, basalts became an important component of the crust.
2. Short-lived (<100 Ma) komatiite-driven plate tectonics may have existed on the terrestrial planets at about 4.5 Ga.
3. Preserved continental crust on Earth dates to 3.8–3.9 Ga and detrital zircons with ages up to 4.2 Ga suggest that very minor continental “islands” may have existed prior to 4 Ga.
4. Early terrestrial continental crust, chiefly tonalites and granodiorites, may have been produced by partial melting of descending hydrated mafic slabs. Continued growth of continents occurred by magma underplating and arc collisions, the latter of which may have locally thickened continents to >50 km.
5. Nd isotopic data from Early Archean sediments and melting relationships of tonalite do not favor the existence of voluminous continental crust prior to 3 Ga, and hence do not support models of early rapid continental growth. Most data suggest that continents grew rapidly in the Late Archean (2.7–3.0 Ga) with renewed growth at 1.9–2.0 Ga.
6. Rapid continental growth in the Late Archean may have occurred in response to catastrophic subduction (and partial melting) of basalt in response to subduction geotherms passing into the eclogite stability field.
7. After 2500 Ma, rates of continental growth and accretion for North America differ by <100 Ma. The North American continent was largely assembled by plate collisions between 1750 and 1950 Ma.
8. Mantle heterogeneities may survive for billions of years in regions of relatively low strain rate.
9. The positive ϵNd values that characterize most Archean rocks and reflect depleted mantle sources reflect formation of an enriched reservoir (crust or enriched mantle) prior to 4 Ga. A thick (150–200 km) mantle lithosphere

probably developed beneath Archean continents during or soon after their formation.

10. To identify meaningful lithologic or geochemical changes in Earth history, it is necessary to compare results from rocks formed in similar tectonic settings.
11. Increases in recycling rates of sediments and in $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios of marine carbonates after 2.7 Ga ago reflect rapid continental growth in the Late Archean.
12. Geochemical and geologic changes near the Archean-Proterozoic boundary are probably due to cooling of the Earth. Immediate causes of these changes are summarized in table 10.5.
13. Most orogenies last about 50 Ma. Although orogeny may be episodic on a given continent or portion thereof, with two exceptions (2.7 and 1.9 Ga) worldwide synchronicity of orogeny is not favored by the distribution of U–Pb igneous crystallization dates. Gaps in orogeny on one continent may correspond to major orogenies on other continents.
14. Differences in evolution of terrestrial planets can be explained chiefly by differences in cooling rates, volatile contents, and compositions and by distance from the sun.
15. Of the three thermal–tectonic models suggested for Venus (conduction, hot-spot, and plate tectonics), surface features tend to support some combination of hotspots and plate tectonics. Some plateau areas on Venus may represent ancient continental crust.
17. Of the terrestrial planets, only the Earth and Venus have cooled slowly enough to support plate tectonics. Whether plate tectonics is operative on Venus awaits higher-quality data.

Suggestions for Further Reading

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CHAPTER 11

Exogenic Earth Systems

Introduction

Exogenic Earth systems include the atmosphere, hydrosphere, and biosphere and their interactions with the crust and mantle. Also considered in this category are the effects of extraterrestrial impact and the history of ancient climates. In terms of exogenic systems the Earth is a unique body in the solar system. It is the only planet with oceans and with an oxygen-bearing atmosphere capable of sustaining higher forms of life. How did such an atmosphere–ocean system form and why only on the Earth? When and how was life created? What is the role of plate tectonics and extraterrestrial impact in the evolution of exogenic systems? It is these and related questions that will be addressed in the final chapter.

General Features of the Atmosphere

Atmospheres are the gaseous carapaces that surround some planets, and because of gravitational forces they increase in density toward planetary surfaces. The Earth's atmosphere is divided into six regions as a function of height (fig. 11.1). The *magnetosphere*, the outermost region, is composed of high-energy nuclear particles trapped in the Earth's magnetic field. This is underlain by the *exosphere* in which light-weight molecules (such as H_2) occur in extremely low concentrations and escape from the Earth's gravitational field. Temperature decreases rapidly in the *ionosphere* (to about $-90^{\circ}C$) and then increases again to near $0^{\circ}C$ at the base of the *mesosphere*. It drops again in the *stratosphere* and then rises gradually in the *troposphere* toward the Earth's surface. Since warm air overlies cool air in the stratosphere, this layer is relatively stable and undergoes very little vertical mixing. The temperature maximum at the top of the stratosphere is caused by absorption of ultraviolet (UV) radiation in the ozone layer. The troposphere is a turbulent region that contains about 80% of the mass of the atmosphere and most of its water vapor. Tropospheric temperature decreases toward the poles, which together with vertical temperature changes causes continual convective overturn in the troposphere.

The Earth's atmosphere is composed chiefly of nitrogen (78%) and oxygen (21%) with very small amounts of other gases such as argon and carbon dioxide. In this respect the atmosphere is unique among planetary atmospheres (table 11.1). Venus and Mars have atmospheres composed largely of CO_2 , and the surface pressure on Venus ranges up to 100 times that on Earth while the surface pressure of Mars is

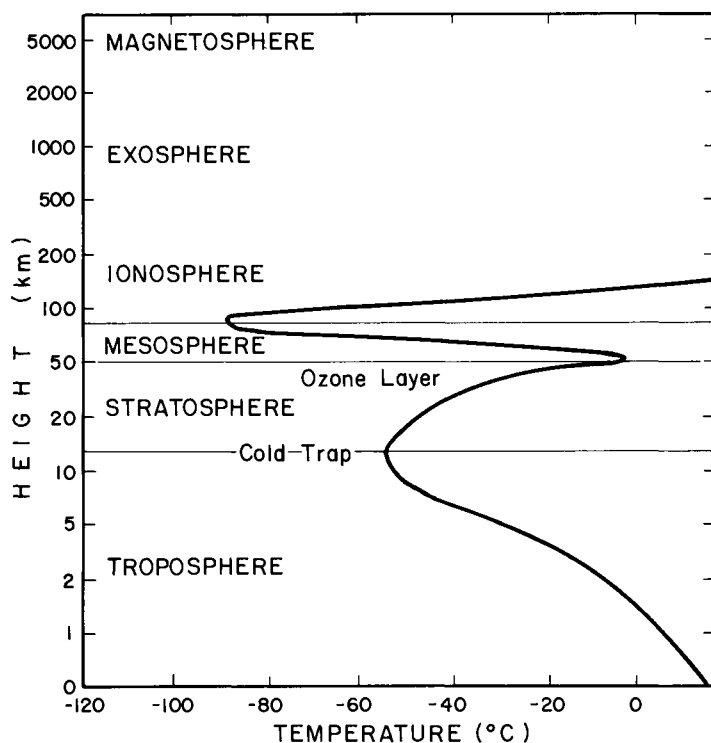


FIG 11.1 Layers of the Earth's atmosphere showing average temperature distribution.

TABLE 11.1 Composition of Planetary Atmospheres

	Surface Temperature (°C)	Surface Pressure (bars)	Principal Gases
Earth (Ancient)	-100 to 40	1-1.3	CO ₂ , N ₂ , H ₂ O
Earth	-20 to 40	0.1-1	N ₂ , O ₂
Venus	500 to 600	10-100	CO ₂
Mars	-130 to 25	≤0.01	CO ₂
Jupiter	-160 to -90	≤2	H ₂ , He
Saturn	-180 to -120	≤2	H ₂ , He
Uranus	-220 to -120	≤5	H ₂ , CH ₄
Neptune	-220 to -120	≤10	H ₂
Pluto	-220 to -200	≤0.005	CH ₄

$<10^{-2}$ of that of the Earth. The surface temperatures of Earth, Venus, and Mars are also very different (table 11.1). The outer planets are composed largely of low-density ices of hydrogen and helium, and their atmospheres consist chiefly of hydrogen and, in some cases, helium and methane.

Concentrations of gases in the Earth's atmosphere are controlled by several different processes. The distributions of minor gases such as CO, H₂, and ozone (O₃) are controlled primarily by reactions in the stratosphere caused by solar radiation known as *photochemical reactions*. Solar photons may fragment gaseous molecules

(such as O_2 , H_2 , CO_2) producing free radicals (C, H, O), a process called *photolysis*. One important reaction produces free oxygen atoms that are unstable, and recombine to form *ozone*. This reaction occurs at heights of 30–60 km with most ozone collecting in a relatively narrow band at about 25 km (fig. 11.1). Ozone, however, is unstable and continually breaks down to form molecular oxygen. The production rate of ozone is approximately equal to the rate of loss, and thus the ozone layer maintains a relatively constant thickness in the stratosphere. Ozone is an important constituent in the atmosphere because it absorbs UV radiation from the Sun, which is lethal to most forms of life. Hence, the ozone layer provides an effective shield that permits a large diversity of living organisms to survive on the Earth. The distributions of N_2 , O_2 , and CO_2 in the atmosphere are controlled by volcanic eruptions and by interactions between these gases and the solid Earth, the oceans, and living organisms.

The Primitive Atmosphere

Three possible sources are considered for the Earth's atmosphere: residual gases remaining after Earth accretion, extraterrestrial sources, and degassing of the Earth. Of these, only degassing accommodates a wide variety of geochemical and isotopic constraints. Degassing of the Earth may be direct by volcanism, or indirect by weathering at the surface. One line of evidence supporting a degassing origin for the atmosphere is the large amount of ^{40}Ar in the atmosphere (99.6%) compared to the Sun or carbonaceous chondrites (<0.1%). ^{40}Ar is produced by the radioactive decay of ^{40}K in the solid Earth and escapes into the atmosphere chiefly by volcanism. The relatively large amount of this isotope in the terrestrial atmosphere indicates that the Earth is extensively degassed of Ar and, because of a similar behavior, also of other gases.

Although most investigators agree the present atmosphere (except for O_2) is the product of degassing, whether a primitive atmosphere existed and was lost before extensive degassing began is a subject of controversy. One line of evidence supporting the existence of an early atmosphere is the fact that volatile elements should collect around accreting planets during their late stages of accretion. This follows from the very low temperatures at which volatile elements condense from the solar nebula (Chapt. 2). A significant depletion in rare gases in the Earth compared to carbonaceous chondrites and the Sun indicates that if a primitive atmosphere collected during accretion, it must have been lost. The reason for this is that gases with low atomic weights (CO_2 , CH_4 , NH_3 , H_2 , etc.) that may have comprised this early atmosphere should be lost even more readily than rare gases with high atomic weights (Ar, Ne, Kr, Xe). Just how such a primitive atmosphere may have been lost is not clear. One possibility is by a T-Tauri solar wind (see Chapt. 2). If the Sun evolved through a T-Tauri stage during or soon after (<100 Ma) planetary accretion, this wind of high-energy particles liberated could readily blow volatile elements out of the inner solar system. An alternate loss mechanism is by impact with one or more large bodies during the late stages of accretion.

Two models have been proposed for the composition of a primitive atmosphere. The Oparin–Urey model (Oparin, 1953) suggests that the atmosphere was reducing and composed dominantly of CH_4 with smaller amounts of NH_3 , H_2 , He, and H_2O ,

TABLE 11.2 *Excess Volatiles in Volcanic Gases and Near-surface Terrestrial Reservoirs*

	Volcanic Gases (%)	Near-surface Reservoirs* (%)
H ₂ O	83	87
CO ₂	12	12
Cl, N ₂ , S	5	1

After Rubey (1951).

*Includes atmosphere, hydrosphere, biosphere, and sediments.

while the Abelson model (Abelson, 1966) is based on an early atmosphere composed of CO₂, CO, H₂O, and N₂. Neither atmosphere allows significant amounts of free oxygen, and experimental studies indicate that reactions may occur in either atmosphere that could produce the first living organisms.

By analogy with the composition of the Sun and the compositions of the atmospheres of the outer planets and of volatile-rich meteorites, an early terrestrial atmosphere may have been rich in such gases as CH₄ and NH₃, and thus have been a reducing atmosphere. One of the major problems with an atmosphere in which CH₄ and NH₃ are important constituents is that both of these species are destroyed directly or indirectly by photolysis (Cogley and Henderson-Sellers, 1984). Ammonia is destroyed by UV radiation in as little as 1–10 a. In addition, NH₃ is highly soluble in water and should be removed rapidly from the atmosphere by rain and solution at the ocean surface. Although CH₄ is more stable against photolysis, OH, which forms as an intermediary in the methane oxidation chain, is destroyed by photolysis at the Earth's surface in <50 a. In that H₂ rapidly escapes from the top of the atmosphere, it also is an unlikely major constituent in an early atmosphere. Thus, a reducing primitive atmosphere seems unlikely, and only a non-oxidizing Abelson-type atmosphere seems capable of surviving photolysis.

The Degassed Atmosphere

Excess Volatiles

Most of the Earth's atmosphere appears to have formed by degassing of the mantle and crust. This degassing occurs directly during volcanism, or indirectly by weathering of igneous rocks at the Earth's surface. Of the two, volcanism appears to be most important both in terms of current degassing rates and calculated past rates. The volatiles in the atmosphere, hydrosphere, biosphere, and sediments that cannot be explained by weathering of the crust are known as *excess volatiles* (Rubey, 1951). These include most of the H₂O, CO₂, and N₂ in these near-surface reservoirs. The similarity in the distribution of excess volatiles in volcanic gases to those in near-surface reservoirs (table 11.2) strongly supports a volcanogenic origin for these gases, and thus also supports a degassing origin for the atmosphere.

Composition of the Early Atmosphere

Two models have been proposed for the composition of the early degassed atmosphere, depending on whether metallic iron existed in the mantle in the early

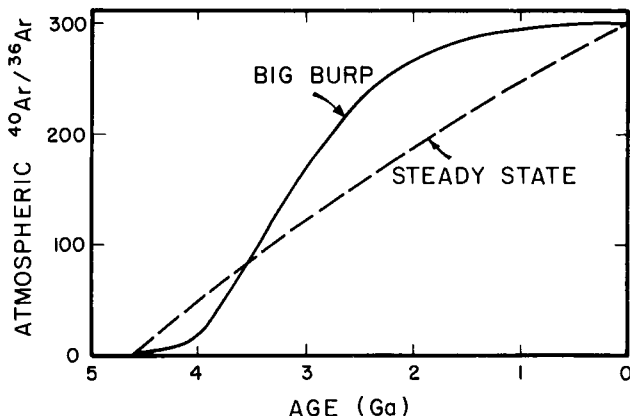


FIG 11.2 Idealized evolution of atmospheric $^{40}\text{Ar}/^{36}\text{Ar}$ in the steady-state and big burp models for atmospheric growth.

Archean. If metallic iron was present, equilibrium chemical reactions would liberate large amounts of H_2 , CO , and CH_4 and small amounts of CO_2 , H_2O , H_2S , and N_2 (Holland, 1984). If iron was not present, reactions would liberate large amounts of CO_2 , H_2O , and N_2 and minor amounts of H_2 , HCl , and SO_2 . Because evidence suggests the core formed during the late stages of planetary accretion (Chapt. 3), it is possible that little if any metallic iron remained in the mantle when degassing began. However, if degassing also began before the completion of accretion, metallic iron would have been present in the mantle and the first atmosphere would have been composed chiefly of H_2 , CO , and CH_4 . Because the relative timing of early degassing and core formation are not well constrained, the composition of the earliest degassed atmosphere is not well known. Core formation and extensive degassing (see below) were probably completed in <100 Ma after accretion, and the composition of the early atmosphere may have changed rapidly during this time interval in response to decreasing amounts of metallic iron in the mantle. Several lines of evidence, however, indicate that by 4.0 Ga the Earth's atmosphere was rich in CO_2 and H_2O (Holland et al., 1986; Kasting, 1987). Among these are evidence for running water, the necessity of an early greenhouse effect (see below) and the fact that the atmospheres of Venus and Mars are composed largely of CO_2 .

Growth Rate of the Atmosphere

Two extreme models are considered for the growth of the atmosphere with time: the *big burp model*, in which the atmosphere grows by rapid degassing soon after planetary accretion (Fanale, 1971), and the *steady-state model* where the atmosphere grows slowly over geologic time (Rubey, 1951). One way of evaluating atmospheric growth rate is to monitor the buildup of ^{40}Ar , ^{129}Xe , and ^4He in supracrustal rocks that have equilibrated with the atmosphere–ocean system. ^{40}Ar is produced by the radioactive decay of ^{40}K in the Earth, and as it escapes from the Earth's interior it collects in the atmosphere. Since ^{36}Ar is nonradiogenic, the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio should record distinct evolutionary paths for Earth degassing. The steady-state model is characterized by a gradual increase in the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio with time, while the big burp model is characterized by initially small changes in this ratio followed by rapid increases (fig. 11.2). This is because ^{40}Ar is virtually absent at the time of accretion,

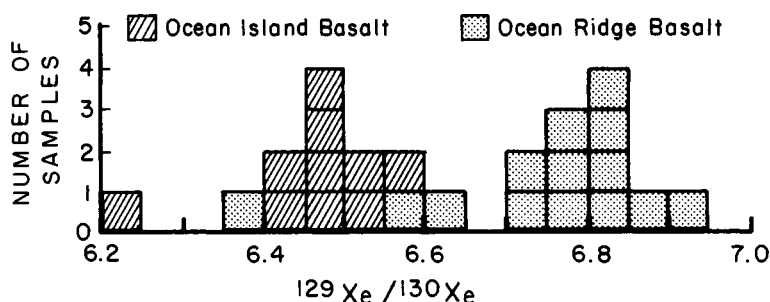


FIG 11.3 Distribution of $^{129}\text{Xe}/^{130}\text{Xe}$ ratios in oceanic basalts. The excess ^{129}Xe in MORB relative to the stable ^{130}Xe isotope implies a similar excess in MORB mantle sources.

and hence the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio is not sensitive to early atmospheric growth. Later in the big burp model, however, as ^{40}Ar begins to be liberated, the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio grows rapidly leveling off after about 2 Ga ago (Sarda et al., 1985). In order to test these two models it is necessary to determine $^{40}\text{Ar}/^{36}\text{Ar}$ ratios in rocks that equilibrated with the atmosphere-ocean system in the geologic past. Unfortunately, due to the mobility of argon, reliable samples to study are difficult to find. However, 2-Ga-old cherts are reported to have $^{40}\text{Ar}/^{36}\text{Ar}$ ratios similar to the present atmosphere (295) tending to favor the big burp model. Some argon degassing models suggest that the atmosphere grew very rapidly in the first 100 Ma after accretion, followed by a more continuous growth to the present (Sarda et al., 1985). These models indicate a mean age for the atmosphere of 4.4 Ga, suggesting very rapid early degassing of the Earth, perhaps beginning during the late stages of accretion. On the other hand, relatively young K-Ar ages of MORB sources (<1 Ga) show that the depleted upper mantle was not completely degassed and decoupled from the atmosphere early in Earth History (Fisher, 1985). These data, together with relatively young U-He and U-Xe ages of depleted mantle, suggest that rather extensive degassing has continued up to at least the last few hundred million years.

Another approach to studying the growth rate of the atmosphere is using rare gases that are trapped in ocean-ridge basalts (MORB). MORB glasses have proved to be particularly useful in that they appear to trap gases in the same proportions as they occur in mantle sources. The most informative gases in terms of atmosphere growth are those that have isotopes that are products of radioactive decay, such as ^4He , ^{40}Ar , and various Xe isotopes. The rate at which these isotopes accumulate in the mantle compared to stable isotopes of the same element can be related to the degassing rate of the mantle. One potentially important rare gas isotopic system to study mantle degassing is the ^{129}I - ^{129}Xe system (Staudacher et al., 1982). Large concentrations of ^{129}Xe known as ^{129}Xe anomalies are reported in some MORB glasses. These are produced by the radioactive decay of ^{129}I , which has a short half-life of 17 Ma. If the Earth was largely degassed soon after accretion, much of the Xe together with other gases would enter the atmosphere. Iodine, however, which, is not as volatile, should remain within the Earth, and ^{129}I should decay to ^{129}Xe producing an enrichment in ^{129}Xe in the mantle (fig. 11.3). It is also of interest in terms of planetary accretion rates, that for ^{129}I to be added to the Earth requires that accretion be completed within about five half-lives of ^{129}I , otherwise all of the ^{129}I

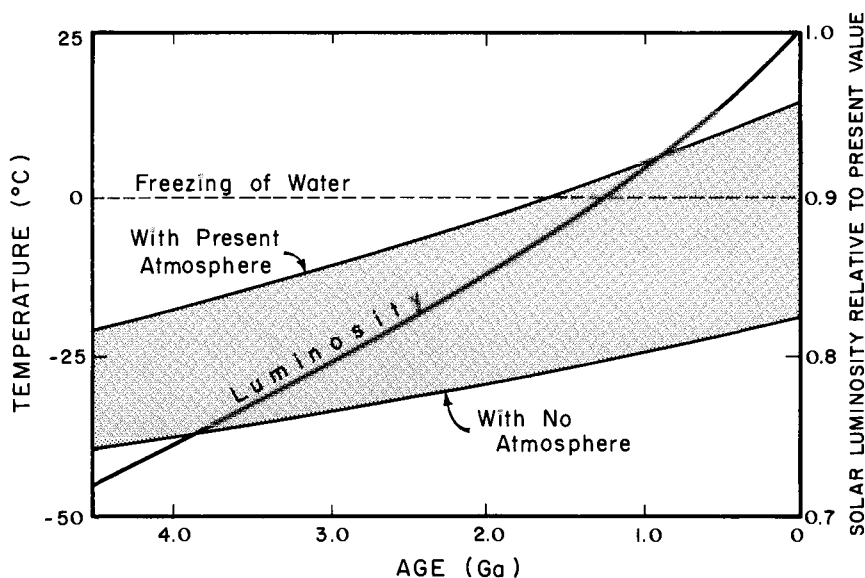


FIG 11.4 The estimated increase in solar luminosity with geologic time and its effect on the Earth's surface temperature.

would be lost in the solar nebula. Hence the presence of excess ^{129}Xe in the Earth implies a time interval between nucleosynthesis and accretion of ≤ 100 Ma. The fact that Xe anomalies are found only in MORB indicates that depleted upper mantle sources were extensively degassed within 20–30 Ma after accretion was completed (Staudacher et al., 1982; Allegre et al., 1987). The absence of ^{129}Xe anomalies in island basalts indicates that other parts of the mantle were not extensively degassed during this early event, and that they have degassed more gradually. There are, however, problems with the catastrophic early degassing model, not the least of which is that ^{129}Xe anomalies in MORB glasses are not found by some investigators (Fisher, 1985).

In conclusion, rare gas studies of terrestrial rocks suggest, but do not prove, that large portions of the mantle underwent degassing within 50 Ma following planetary accretion. Some degassing of the Earth has continued, however, at least up to the last few hundred million years. The fraction of the atmosphere released during the early degassing event is unknown, but may have been substantial.

The Faint Young Sun Paradox

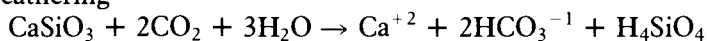
Models for the evolution of the Sun indicate that it was less luminous when it entered the main sequence 4.5 Ga ago. This is because with time the Sun's core becomes denser, and therefore hotter, as hydrogen is converted to helium. Calculations indicate that the early Sun was 25–30% less luminous than it is today, and that its luminosity has increased with time in an approximately linear manner (Kastin, 1987). The paradox associated with this luminosity change is that the Earth's average surface temperature would have remained below freezing until about 2 Ga ago for an atmosphere with the same characteristics as today (fig. 11.4). Yet the

presence of sedimentary rocks as old as 3.8 Ga indicates the existence of oceans and running water. The probable solution to the faint young Sun paradox is that the early atmosphere contained a much larger quantity of CO₂ than it does today. CO₂ levels of even a few tenths of a bar could prevent freezing temperatures at the Earth's surface due to an enhanced greenhouse effect. The *greenhouse effect* is caused by gases such as CO₂ and H₂O vapor that allow sunlight to reach a planetary surface but absorb infrared radiation reflected from the surface and reradiate some of this energy back to the surface. The net result is warming of the planetary surface in the same manner that greenhouses are warmed. An upper bound on the amount of CO₂ in the Early Archean atmosphere is provided by the carbonate–silicate cycle (see below) and the fact that evaporite gypsum was deposited by 3.5 Ga ago. For gypsum to be stable, the surface temperature must have been ≤58°C and the corresponding upper limit of the P_{CO₂} is a few bars.

The Carbonate–Silicate Cycle

One of the important exogenic systems controlling the P_{CO₂} in the atmosphere is the *carbonate–silicate cycle*. CO₂ enters the Earth's atmosphere during volcanic eruptions and hydrothermal activity along ocean ridges. It returns to the oceans by weathering of silicates and deposition of marine carbonates which can be summarized by the following reactions (Garrels and MacKenzie, 1971):

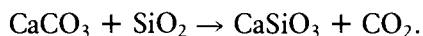
weathering



Deposition



The cycle is completed when pelagic carbonates are subducted, metamorphosed, and perhaps partially melted with CO₂ released and again entering the atmosphere by volcanism. The metamorphic reactions that liberate CO₂ can be summarized by the carbonate–silica reaction,



For equilibrium to be maintained in this cycle, increased input of CO₂ into the atmosphere results in more weathering and carbonate deposition, thus avoiding build-up of CO₂ in the atmosphere. Various feedback mechanisms in this cycle may have stabilized the Earth's surface temperature in the geologic past (Walker et al., 1983). As an example, if the solar luminosity were to suddenly drop, the surface temperature would fall, causing a decrease in the rate of silicate weathering due to a decrease in evaporation from the oceans (and hence a decrease in precipitation). This results in CO₂ accumulation in the atmosphere, which increases the greenhouse effect and restores higher surface temperatures. The converse of this feedback would occur if the surface temperature were to increase suddenly.

Although photosynthesis also plays a role in controlling CO₂ levels in the atmosphere, most data indicate that the carbonate–silicate cycle is the primary control, and probably has been the primary control in the geologic past.

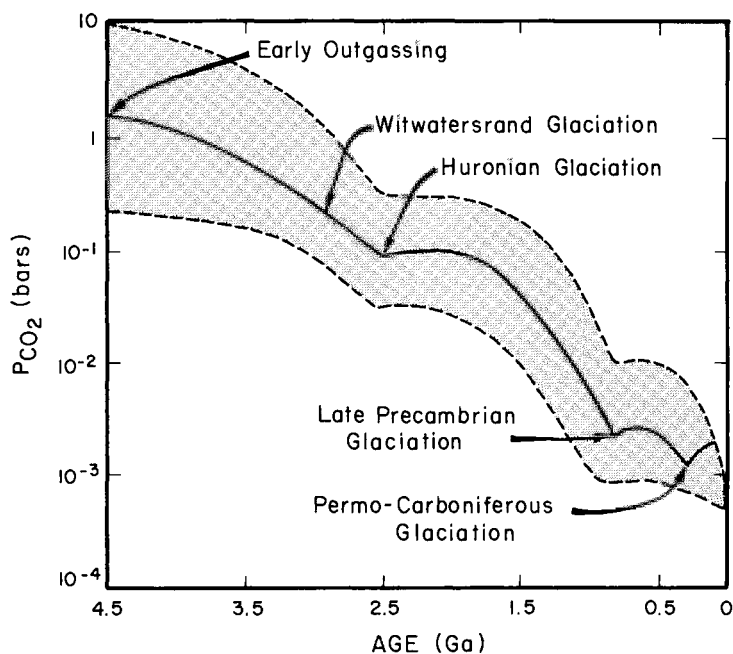


FIG 11.5 Model for the variation of atmospheric CO_2 level with geologic time. After Kastin (1987).

The History of Atmospheric CO_2

Although the history of CO_2 in the Earth's atmosphere is only beginning to be understood, it seems clear that the abundance of CO_2 has decreased with time. Constraints on atmospheric CO_2 content include paleoclimatic data and information from paleosols. Calculations indicate that a drop of only $2^\circ C$ in greenhouse temperature is enough to cause widespread glaciation (Kastin, 1987). Uncertainties in this result include the possible albedo effects of snow, ice, and clouds, which are ignored in the model. Decreases in P_{CO_2} corresponding to major glaciations in the geologic past are illustrated in figure 11.5. An upper limit of P_{CO_2} of 10 bar at 4.5 Ga assumes an ocean-covered Earth and a lower limit at this time (~ 0.2 bar) is the amount of CO_2 required to keep the surface temperature above freezing. The suggested increase in P_{CO_2} during the Mesozoic reflects the warm climates during this time.

Although the sudden decreases in P_{CO_2} accompanying glaciations may reflect decreases in surface temperature caused by changes in the Sun's luminosity or other astrophysical phenomena, the overall decrease in atmospheric CO_2 level with time appears to be related to some combination of changes in the carbonate-silicate cycle, the growth of ozone in the atmosphere, and the increase in photosynthetic organisms. Shallow marine carbonates deposited in cratonic basins are effectively removed from the carbonate-silicate cycle and preserved in the continental crust. The increasing importance of carbonate deposition and burial after widespread Late Archean continental growth is probably the major contributor to lowering of atmospheric CO_2 after 2.5 Ga. The growth of O_2 and O_3 in the atmosphere

beginning in the Early Proterozoic (see fig. 11.8) also reduced the need of CO₂ as a major greenhouse gas, and the increase in abundance of photosynthetic algae beginning perhaps in the Late Archean provided a new sink for CO₂. However, biogenic CO₂ is partly returned to the atmosphere as organic remains are exposed at the surface and weathered.

The Continuously Habitable Zone

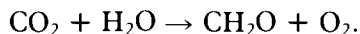
Living organisms require a very narrow range in surface temperature, and hence a narrow range in the content of greenhouse gases. From a knowledge of temperature and gas distributions in the inner solar system, the width of this zone, known as the *continuously habitable zone* or CHZ, can be estimated. If the Earth, for instance, had accreted in an orbit 5% closer to the Sun, the atmosphere would continue to rise in temperature and the oceans would evaporate, a condition known as a *runaway greenhouse* (Hart, 1978). Such a situation exists on Venus, and is clearly not habitable for even primitive unicellular organisms. The outer extreme of the CHZ is less certain, but appears to extend somewhat beyond the orbit of Mars. Thus, the CHZ in the solar system would appear to extend from about 0.95 to 1.5 AU.

This observation is important in that the CHZ is wide enough that habitable planets may exist in other planetary systems.

The Origin of Oxygen

Oxygen Controls in the Atmosphere

In the modern atmosphere, oxygen is produced almost entirely by photosynthesis by the well-known reaction,



A very small amount of O₂ is also produced in the upper atmosphere by photolysis of H₂O molecules. Oxygen is removed from the atmosphere by respiration and decay, which can be considered as the reverse of the photosynthesis reaction. Virtually all of the O₂ produced by photosynthesis in a given year is lost in <50 a by oxidation of organic matter. Oxygen is also liberated by the reduction of sulfates and carbonates in marine sediments. Without oxidation of organic matter and sulfide minerals during weathering, the O₂ content of the atmosphere would double in about 10⁴ a (Holland et al., 1986). It has also been suggested that forest fires may have contributed in the past in controlling the upper limit of O₂ in the atmosphere. It appears that the O₂ content of the modern atmosphere is maintained at a near-constant value by various negative feedback mechanisms involving primarily photosynthesis and decay. If, however, photosynthesis were to stop, respiration and decay would continue until all of the organic matter on Earth was transformed to CO₂ and H₂O. This would occur in about 20 a and involve only a minor decrease in the amount of O₂ in the atmosphere. Weathering would continue to consume O₂ and would take about 4 Ma to use up the current atmospheric supply.

Prior to the appearance of photosynthetic microorganisms in the geologic record, and probably for a considerable time thereafter, photosynthesis was not an



FIG 11.6 Early Proterozoic banded iron formation from northern Minnesota.

important process in controlling atmospheric oxygen levels. In the primitive atmosphere, O_2 content was controlled by the rate of photolysis of water, hydrogen loss from the top of the atmosphere, and weathering rates at the surface. The rate at which water is supplied by volcanic eruptions is also important because volcanism is the main source of water available for photolysis. If metallic iron was present in the mantle during the earliest stages of degassing of the Earth, H_2 would have been an important component of volcanic gases. Since H_2 rapidly reacts with O_2 to form water, significant amounts of volcanic H_2 would prevent O_2 from accumulating in the early atmosphere. As H_2O instead H_2 became more important in volcanic gases, in response to removal of iron from the mantle as the core grew, oxygen could begin to accumulate. Also, early photosynthesizing organisms may have contributed to the first oxygen in the atmosphere. As photosynthesis became more widespread, recombination of H_2 and O_2 to form water could not keep pace with O_2 input, and the oxygen level in the atmosphere must have increased. This assumes that the rate of weathering did not increase with time, an assumption which is supported by geologic data.

Geologic Indicators of Ancient Atmospheric Oxygen Levels

Banded Iron Formation

The distribution of banded iron formation with geologic time provides a constraint on O_2 level in the oceans and atmosphere. *Banded iron formations*, or *BIF*, are chemical sediments, typically thin-bedded or laminated with $\geq 15\%$ iron of sedimentary origin (fig. 11.6). They also commonly contain layers of chert and generally

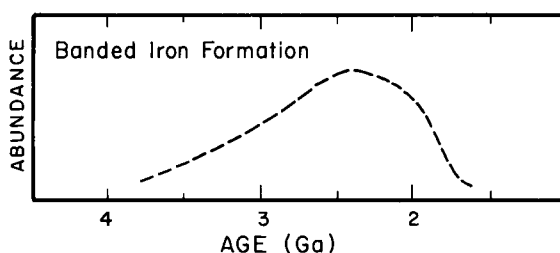


FIG 11.7 Relative abundance of banded iron formation during geologic time. After Walker et al. (1983).

have $\text{Fe}^{+3}/\text{Fe}^{+2} + \text{Fe}^{+3}$ ratios in the range of 0.3–0.6, reflecting an abundance of magnetite (Fe_3O_4). BIF are metamorphosed and major minerals include quartz, magnetite, hematite, siderite, and various other Fe-rich carbonates, amphiboles, and sulfides. Although most abundant in the Late Archean and Early Proterozoic (figs 10.21 and 11.7), BIF occur in rocks as old as 3.8 Ga (Isua, SW Greenland) and as young as 0.8 Ga (Rapitan Group, NW Canada). The Hamersley basin (2.5 Ga) in Western Australia is the largest known single BIF depository.

Most investigators consider the large basins of BIF to have formed on cratons in shallow marine or perhaps playa-like lake environments. Many BIF are characterized by thin wave-like laminations that can be correlated over hundreds of kilometers (Trendall, 1983). During the Late Archean and Early Proterozoic, enormous amounts of ferrous iron appear to have entered BIF basins with calculated Fe precipitation rates of 10^{12} – 10^{15} gm/a (Holland, 1984). Although some of the Fe^{+2} undoubtedly came from weathering of the continents, most appears to have entered the oceans by submarine volcanic activity. Deposition occurred as Fe^{+2} reacted with dissolved O_2 , probably at shallow depths forming flocculent, insoluble ferric and ferro-ferric compounds (Walker et al., 1983). Although Precambrian BIF clearly represents a large oxygen sink, the P_{O_2} in the coexisting atmosphere may have been quite low. For instance, hematite (Fe_2O_3) or $\text{Fe}(\text{OH})_3$ can be precipitated over a wide range of P_{O_2} values and Early Archean BIF may have been deposited in reducing marine waters. The large amount of oxygen in Late Archean and especially in Early Proterozoic BIF, however, appears to require the input of photosynthetic oxygen. This agrees with paleontologic data that indicate a rapid increase in numbers of photosynthetic algae during the same period of time. Many BIF contain well-preserved fossil algae remains. Thus, it appears that the increase in abundance of BIF in the Late Archean and Early Proterozoic (fig. 11.7) reflects an increase in oceanic O_2 content in response to increasing numbers of photosynthetic organisms. Only after most of the BIF was deposited and the Fe^{+2} in solution was exhausted did O_2 begin to escape from the oceans and accumulate in the atmosphere in appreciable quantities (Cloud, 1973). The drop in abundance of BIF at about 1.7 Ga reflects this exhaustion of reduced iron.

Redbeds and Sulfates

Redbeds are detrital sedimentary rocks with red ferric oxide cements. They generally form in fluvial or alluvial environments and the red cements are the result of

subaerial oxidation (Folk, 1976), and thus require the presence of O_2 in the atmosphere. The fact that redbeds do not appear in the geologic record until about 2.4 Ga ago suggests that oxygen levels were very low in the Archean atmosphere.

Sulfates, primarily gypsum and anhydrite, occur as evaporites. Although evidence of gypsum deposition is found in some of the oldest supracrustal rocks (~ 3.5 Ga), evaporitic sulfates do not become important in the geologic record until after about 2 Ga. Since their deposition requires free O_2 in the ocean and atmosphere, their distribution supports rapid growth of oxygen in the atmosphere, after the Early Proterozoic. Additional evidence favouring a switch to oxidizing conditions in the Early Proterozoic is provided by sulfur and carbon isotope data (see later discussion). An increase in the range of $\delta^{13}C$ values for Archean organic matter implies relatively large amounts of marine methane, which in turn requires very low atmospheric P_{O_2} prior to 2.5 Ga.

Detrital Uraninite Deposits

Several occurrences of Late Archean to Early Proterozoic detrital uraninite and pyrite are well documented, the best known of which are those in the Witwatersrand basin in South Africa (~ 2.8 Ga) and those in the Blind River–Elliot Lake area in Canada (~ 2.3 Ga). No significant occurrences, however, are known to be younger than Mid-Proterozoic (Walker et al., 1983). A few very minor occurrences of detrital uraninite are found in young sediments associated with rapidly rising mountain chains like the Himalayas. Both uraninite and pyrite are unstable under oxidizing conditions and are rapidly dissolved. The preservation of major Late Archean and Early Proterozoic deposits of detrital uraninite and pyrite in conglomerates and quartzites indicates that weathering did not lead to total oxidation and dissolution of uranium and iron. To preserve uraninite, P_{O_2} must have been $< 10^{-2}$ of the present atmospheric level. The few occurrences of young detrital uraninite are metastable, and the uraninite is preserved only because of extremely rapid sedimentation rates and will not survive for long.

The restriction of major detrital uraninite–pyrite deposits to ≥ 2.3 Ga again favors very low O_2 levels in the atmosphere prior to this time.

Paleosols

Paleosols are preserved ancient weathering profiles or soils that contain information about atmospheric composition (Retallack et al., 1984; Holland et al., 1986). In pre-2.0 Ga paleosols developed on granitic rocks, iron is preserved as ferric oxides, whereas a large amount of the iron appears to have been lost in paleosols developed on mafic rocks. Differences between the behavior of iron in these two types of paleosols can be used to constrain the P_{O_2}/P_{CO_2} ratio in the Precambrian atmosphere. This is because the O_2 level of the pre-2.0 Ga atmosphere was high enough to oxidize all Fe^{+2} released from granite weathering, but not all of that released during basalt weathering. Under present weathering conditions both granitic and basaltic soils are fully oxidized.

Biologic Indicators

The Precambrian fossil record also provides clues to the growth of atmospheric

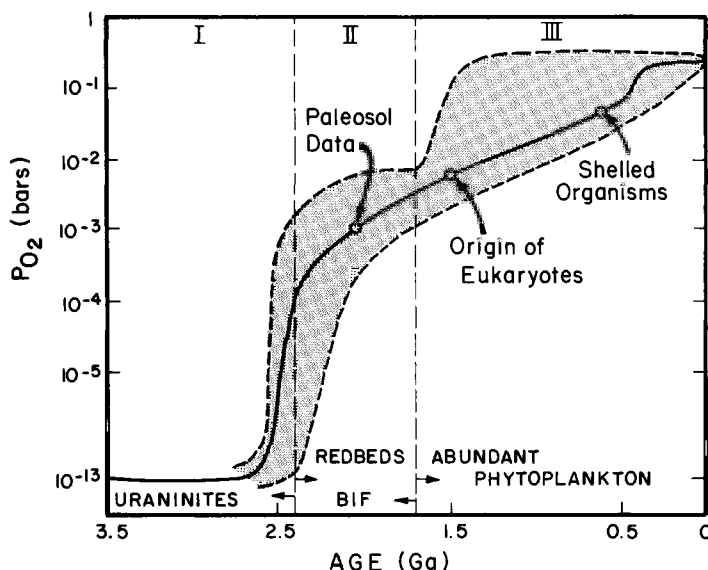


FIG 11.8 Model for the growth of atmospheric O_2 with geologic time. Modified after Kasting (1978). Major stages I, II, and III in atmospheric evolution are noted.

oxygen. Archean and Early Proterozoic life forms were entirely prokaryotic organisms, the earliest forms of which evolved in anaerobic (O_2 -free) environments. Prokaryotes (autotrophs) that produced free oxygen by photosynthesis appear to have evolved by 3.5 Ga. The timing of the transition from an anoxic to oxygenous atmosphere is not well constrained by microfossil remains, but appears to have begun approximately 2.5 Ga ago. Certainly by 2.0 Ga when heterocystous cyanobacteria appear, free O_2 was present in the atmosphere (Cloud, 1976). The first appearance of eukaryotes at about 1.5 Ga indicates atmospheric oxygen levels of at least 1% PAL (of the present atmospheric level), which is necessary for mitosis to occur. The appearance of simple metazoans at about 700 Ma requires O_2 levels high enough for oxygen to diffuse across membranes (~ 0.07 PAL).

The History of Atmospheric O_2

The development of O_2 in the Earth's atmosphere can be divided into three stages (Kasting, 1987): I, a *reducing stage* in which free O_2 occurs in neither the oceans or the atmosphere; II, an *oxidizing stage* in which small amounts of O_2 are present in the atmosphere and surface ocean water but not in deep ocean waters; and III, an *aerobic stage* in which free O_2 pervades the entire ocean-atmosphere system. These stages are diagrammatically illustrated in figure 11.8, with corresponding geologic indicators.

During stage I the ozone shield required to block out lethal UV radiation is absent. If such a shield existed on the primitive Earth it must have been produced by some gaseous compound other than ozone. Although marine cyanobacteria must have produced O_2 by photosynthesis during this stage, in order for large volumes of BIF to form, such O_2 must combine rapidly with Fe^{+2} and be deposited. The change

from stage I to II appears to have occurred during the Early Proterozoic, and is marked by the appearance of redbeds, an increase in abundance of evaporitic sulfates, a broad maximum in the deposition rate of BIF, and the end of major detrital uraninite-pyrite deposition. At this time, photosynthetic O_2 entered the atmosphere at a rate sufficient to clean the atmosphere of molecular H_2 . Evidence that the lower atmosphere and surface ocean waters during this stage were weakly oxidizing, yet deep marine waters were reducing, is provided by the simultaneous deposition of oxidized surface deposits (redbeds, evaporites) and BIF in a reducing (or at least non-oxidizing environment. Two models have been proposed for the control of atmospheric P_{O_2} during stage II. One proposes that low O_2 levels are maintained by mantle-derived Fe^{+2} in the oceans and BIF deposition, and the other relies on low photosynthetic productivity in the ocean. Which of these is most important is not yet clear.

The transition from Stage II to III appears to have occurred between 1.8 and 2.0 Ga, and is marked by the near-disappearance of BIF from the geologic record and the appearance of eukaryotic organisms. The onset of Stage III is defined by the exhaustion of Fe^{+2} from the oceans such that photosynthetic O_2 levels increase and the oceans become oxidizing. Further increases result in liberation of O_2 into the atmosphere. During this stage an effective ozone screen develops, which is probably responsible for the rapid diversification and increase in numbers of microorganisms in the Mid-Proterozoic. By 600 Ma the atmospheric P_{O_2} must have risen to at least 0.1 PAL to permit the appearance of carbonate shell-forming organisms.

Comparative Evolution of the Atmospheres of Earth, Venus, and Mars

The only three terrestrial planets that have retained atmospheres are Earth, Venus, and Mars, yet the compositions and densities of their atmospheres differ significantly from each other (table 11.1). The greenhouse effect, the distance from the Sun, and planetary mass all may have played a role in giving rise to such different atmospheres (Prinn and Fegley, 1987). Planetary surface temperature, as previously discussed, is controlled largely by the greenhouse effect and hence only when significant amounts of CO_2 and H_2O accumulate in the atmosphere will the surface temperature begin to rise. This can be illustrated by the effect of progressively increasing water pressure in a planetary atmosphere (Walker, 1977). Beginning with estimated surface temperatures as dictated by the distance from the Sun, the P_{H_2O} will increase in each planetary atmosphere until intersecting the water vapor saturation curve (fig. 11.9). For the Earth this occurs at about 0.01 atm, where H_2O begins to condense and precipitate on the surface. Further degassing does not increase the H_2O content of the atmosphere or the surface temperature, but rains out and begins to form the oceans. The growing oceans also stabilize the CO_2 content of the atmosphere via the carbonate-silicate cycle. Because Venus is closer to the Sun it had a higher initial surface temperature than Earth. This results in the greenhouse effect raising the surface temperature before the vapor saturation curve is intersected. As degassing continues, H_2O and CO_2 may rapidly accumulate in the atmosphere, producing a *runaway greenhouse* in which atmospheric temperature continues to rise and water never condenses. The reason H_2O is virtually absent in

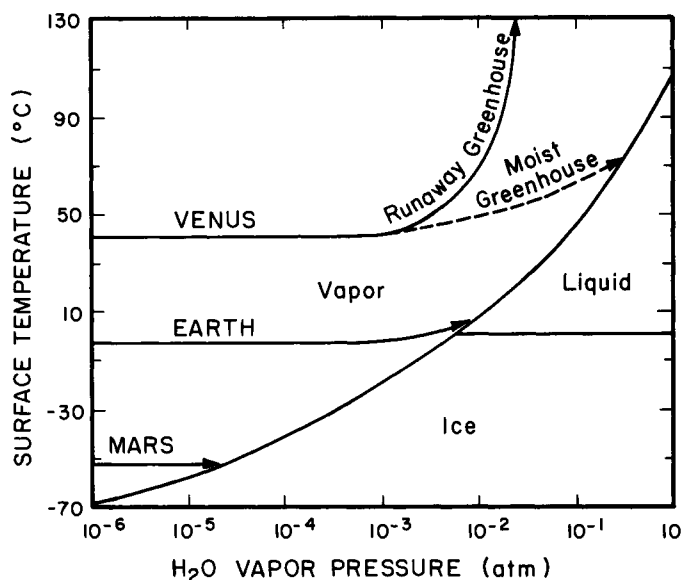


FIG 11.9 Idealized trajectories for increasing atmospheric P_{H_2O} on Venus, Earth, and Mars. Modified after Walker (1977).

TABLE 11.3 Comparison of the Present Atmosphere of Venus and Earth (values in volume percent)

	Earth		Venus
	Atmosphere	Near-surface Reservoirs*	
N ₂	78	2	2
O ₂	21	—	—
CO ₂	0.05	98	98
N ₂ O	≤1	≤1	≤1
Pressure (atm)	1		100

*Atmosphere, biosphere, hydrosphere, and sediments.

the venusian atmosphere today is due to photolysis in the upper atmosphere producing H₂, which readily escapes, and O₂ which may be incorporated in weathered rocks on the surface and perhaps returned to the venusian mantle by plate tectonics. It is noteworthy that the amount of CO₂ in the venusian atmosphere is approximately equivalent to the amount in near-surface reservoirs on Earth (most of it in marine carbonates) (table 11.3). Hence, if it were not for the oceans that provide a vital link in the carbonate-silicate cycle, Earth would probably have an uninhabitable atmosphere like Venus. Mars is farther from the Sun than Earth and has a colder initial surface temperature. Hence, as P_{H_2O} increases and intersects the vapor saturation curve, snow and ice rather than liquid water form (fig. 11.9). As with Earth, neither the surface temperature or atmospheric P_{H_2O} increases with further degassing. Upon further cooling (due to lack of a greenhouse effect), CO₂

would also freeze, forming the martian polar ice caps. The sparsity of light gases like H_2 and N_2 on Mars probably reflects the relatively small mass of the planet and its inability to retain light gases which escape from its gravitational field.

Although the above model accounts for most of the general features of the evolution of the atmospheres of Earth, Venus, and Mars, some aspects require modification. For instance, the model assumes that Mars was cold and frozen from the beginning, an assumption that seems unlikely in terms of the large canyons on the surface that were almost certainly cut by running water. These canyons and their tributary systems are remarkably similar to terrestrial canyons and seem to require surface temperatures above the freezing point of water during the time they formed. For this reason it is probable that an early CO_2 greenhouse effect existed on Mars, and this necessitates some kind of tectonic recycling of CO_2 . Because Mars is too small to retain oceans, and it is unlikely that eclogite-driven plate tectonics ever occurred (Chapt. 10), it is unlikely that an Earth-like carbonate-silicate cycle controlled CO_2 levels in the early atmosphere. Although the nature of the martian early recycling system is unknown, the planet cooled more rapidly than the Earth, eventually trapping most of the CO_2 in its interior. Due to insufficient greenhouse heating, the surface cooled and most of the remaining CO_2 condensed as ice on the surface.

The runaway greenhouse model for Venus assumes that Venus never had oceans, another assumption that is questionable in terms of the high D/H (deuterium/hydrogen) ratio in trace amounts of water in the venusian atmosphere. This ratio, which is 100 times higher than that in the Earth's water, may have developed during preferential loss of H_2 from the venusian atmosphere enriching the residual water in the heavier deuterium isotope, which has a much smaller escape probability than hydrogen. Although this observation does not prove the existence of early oceans on Venus, the planet at one time must have had at least 10^3 times more water than at present. For this reason it is of interest to consider an alternate model to the runaway greenhouse for the evolution of the venusian atmosphere known as the *moist greenhouse model*. This model differs from the runaway greenhouse in that an ocean condenses on Venus, preventing an early runaway greenhouse (fig. 11.9). On the Earth, water is blocked from entering the stratosphere by a cold trap (a temperature minimum) at 10–15 km above the surface (fig. 11.1), and most vapor condenses when it reaches the cold trap. However, if water vapor in the lower atmosphere exceeds 20%, as it must have when venusian surface water (oceans?) boiled away, it produces a moist greenhouse. At this time the cold trap moves to a high altitude (~ 100 km), and no longer is effective in preventing water vapor from rising into the upper atmosphere where it undergoes photolysis and hydrogen escape. Such escape could have eliminated a large venusian ocean in <30 Ma. Following this early loss of water, the venusian atmosphere would rapidly evolve into a runaway greenhouse caused by the remaining CO_2 .

The Oceans

Introduction

The oceans cover 71% of the Earth's surface and contain most of the hydrosphere. The composition of seawater varies geographically and with depth, and is related

principally to the composition of input river water and hydrothermal water along ocean ridges. Chemical precipitation, melting of ice, and evaporation rate also effect the salinity of seawater. Dissolved nutrients such as bicarbonate, nitrate, phosphate, and silica vary considerably in concentration as a result of removal by planktonic microorganisms such as foraminifera, diatoms, and algae. Seawater temperature is rather constant in the upper 300 m ($\sim 18^{\circ}\text{C}$), then drops dramatically to a depth of about 1 km where it remains constant ($3\text{--}5^{\circ}\text{C}$) to the ocean bottom. The cold bottom water in the ocean forms in polar regions, where it sinks to the bottom because of its high density, and travels toward the equator displacing warm surface waters that travel poleward. The geologic history of seawater is closely tied to the history of the atmosphere and oceanic growth, and the concentration of volatile elements in seawater is controlled, in part, by atmospheric growth (Holland et al., 1986). Most paleoclimatologists agree that the oceans formed early in Earth history by condensation of atmospheric water vapor as previously discussed. Some water, however may have been added also by early cometary impact.

Because of the relatively high CO_2 content of the early atmosphere, the first seawater to condense must have had a low pH (<7) due to dissolved CO_2 and other acidic components (H_2S , HCl , etc.). However, this situation was probably short-lived in that submarine volcanic eruptions and seawater recycling through ocean ridges would introduce large amounts of Na, Ca, and Fe. Thus a pH between 8 and 9, and a neutral or negative Eh were probably reached rapidly in the early oceans and maintained thereafter by silicate-seawater buffering reactions and recycling at ocean ridges.

Growth Rate of the Oceans

It is likely that the growth rate of the oceans paralleled that of the atmosphere, and thus a large fraction of the oceans may have formed in the first 100 Ma after planetary accretion. One way of monitoring the volume of seawater with time is by considering the *freeboard of continents*, which is the relative elevation of continents with respect to sea level. The presence of shallow marine sediments in some of the oldest preserved supracrustal rocks (3.6–3.8 Ga) strongly suggests that the freeboard of continents has not significantly changed since the early Archean (Wise, 1973). If, in addition, continents have been on the average 90% of their present thickness for 2.5 Ga, and the volume of seawater recycled back into the mantle has decreased with time, changes in the volumes of continents and oceans are limited to $\leq 20\%$ of their present values. If continents were 80% of their present thickness 2.5 Ga ago, their total area would be 60% of the present area and the corresponding volume of seawater would be $\geq 90\%$ of present.

Changes in the Composition of Seawater with Time

Most information on the composition of seawater in the geologic past relies on indirect data from marine sedimentary rocks. However, fluid inclusions in evaporite minerals and other phases precipitated from seawater may also be a potential source of valuable information. Only recently have fluid inclusions been studied with this problem in mind, and results are not yet conclusive. Some changes in seawater

TABLE 11.4 *Geological and Geochemical Indicators of Seawater Composition*

Observation	Interpretation
1. Mineral and stratigraphic sequence similar in evaporites of all ages	Major dissolved species have not varied by more than a factor of 2–3
2. Decreasing dolomite/calcite ratio in marine carbonates with time	Change in Mg^{+2} depository, decrease in P_{CO_2} with time, or increased loss of Mg^{+} by ocean-ridge recycling of seawater
3. K-rich illite in pre-350-Ma shales, Na-rich smectite thereafter	Higher K/Na in earlier seas or less leaching by land plants before 350 Ma
4. Gradual rise in Sr/Ca ratio of foraminifera during the Cenozoic	Change in Ca input at ocean-ridge hydrothermal systems or change in deposition rate of Sr-rich aragonite
5. Low Ce/La and high Th/U ratios found only after 300 Ma ago in marine sediments	Low O_2 content of seawater prior to 300 Ma maintains Ce chiefly as the reduced Ce^{+3} ion (which is soluble and remains in seawater) and U as U^{+4} (which is insoluble and precipitates)
6. Metals in Paleozoic black shales higher than afterwards	Greater input of metals into seawater from ocean-ridge hydrothermal systems in the Paleozoic
7. Archean marine carbonates enriched in Sr, Ba, Mn, and Fe compared to post-Archean carbonates	More rapid recycling of seawater at ocean ridges in the Archean

In part after Holser (1984).

composition can be monitored by mineralogical or chemical changes in mineral assemblages precipitated from the oceans (table 11.4). Although these changes seem to be well documented, their interpretation in terms of seawater or atmosphere composition is not well understood.

Another approach to constrain the composition of ancient seawater is to examine proportions of specific sediments in similar lithologic assemblages of different ages (Holser, 1984). The large mass of evaporites deposited during some intervals of geologic time may have been important in lowering the NaCl and $CaSO_4$ contents of world oceans. For instance, the vast volumes of salt (chiefly NaCl) deposited in the Late Permian may have caused the average ocean salinity to drop significantly during the Triassic and Jurassic (fig. 11.10). Evaporite deposition has been sharply episodic, and erosion of older evaporites has been relatively slow in returning salts to the oceans. The distribution of evaporites of all ages, coupled with a slow recycling rate into the oceans, suggests the mean salinity of seawater may have decreased since the Mid-Proterozoic due to evaporite deposition.

Oxygen Isotopes in Seawater

Oxygen-18 is fractionated from ^{16}O in the ocean–atmosphere system during evaporation, with the vapor preferentially enriched in ^{16}O . Fractionation of these isotopes also decreases with increasing temperature and salinity. The fractionation of the two isotopes is measured by the deviation of $^{18}O/^{16}O$ from standard mean ocean water (SMOW) in terms of $\delta^{18}O$ where,

$$\delta^{18}O(\text{‰}) = \left[\frac{^{18}O/^{16}O \text{ sample}}{(^{18}O/^{16}O) \text{ SMOW}} - 1 \right] \times 1000.$$

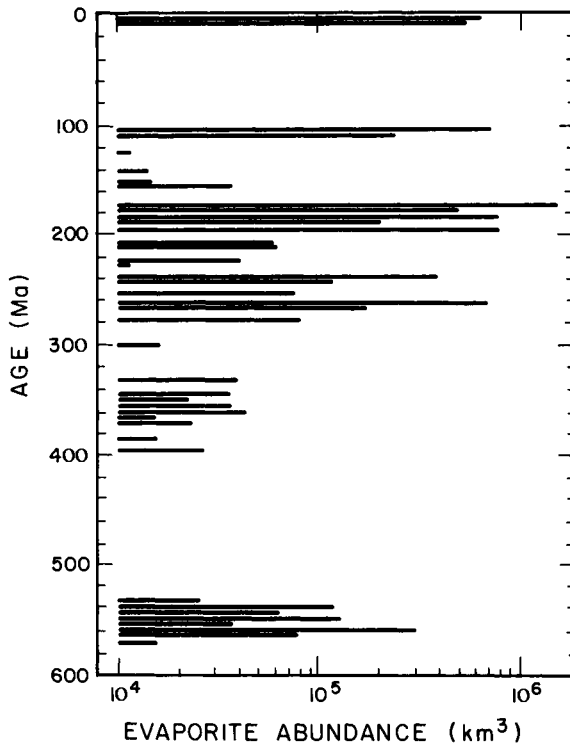


FIG 11.10 Estimated abundances of marine evaporites (NaCl+CaSO₄) during the Phanerozoic. Modified after Holser (1984).

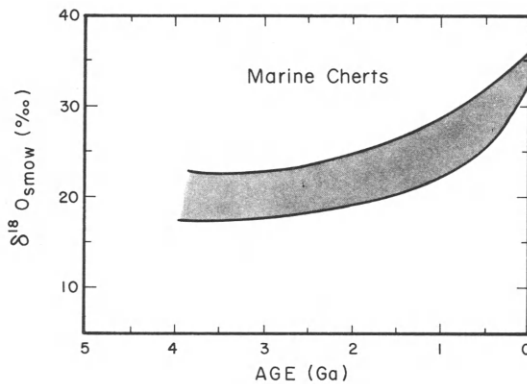


FIG 11.11 Variation in $\delta^{18}\text{O}$ in marine cherts with geologic time. Modified after Knauth and Lowe (1978).

Because marine cherts are resistant to secondary changes in the $\delta^{18}\text{O}$ values, they may be useful in the monitoring of the composition of seawater with geologic time (Knauth and Lowe, 1978). Data suggest that cherts increase in $\delta^{18}\text{O}$ from the Archean to the present (fig. 11.11). Modern cherts have δ values of 20–35‰, while

Archean cherts have values of 15–20‰. Although more scattered, similar differences exist in δ values for marine carbonates. Results suggest that $\delta^{18}\text{O}$ trends are not due to secondary processes such as diagenesis and alteration. Two models have been proposed to explain increasing $\delta^{18}\text{O}$ in cherts, and hence in seawater, with geologic time.

1. The Surface-cooling Model (Knauth and Lowe, 1978)

This trend may be explained by a gradual cooling of the Earth's surface temperature from about 80°C at 3.5 Ga ago to present temperatures. The $\delta^{18}\text{O}$ values of seawater may or may not be constant over this time; however, if not constant they must have changed in such a way so as not to effect the temperature control. A significant problem for the model is evidence for gypsum deposition in the Early Archean. Gypsum casts have been described in 3.5-Ga cherts from the Warrawoona Group in Western Australia, and indicate that seawater temperature at this time could not exceed 58°C or anhydrite rather than gypsum would have been deposited (Holland, 1984).

2. The Mantle–Crust Interaction Model (Perry et al., 1978)

If Archean oceans were about the same temperature as modern oceans they would have a $\delta^{18}\text{O}$ value 34‰ less than cherts precipitated from them or about –14‰. For seawater to evolve to a δ value of zero per mil with time it is necessary for it to react with substances with $\delta^{18}\text{O}$ values >0‰. Recycling seawater at ocean ridges and interactions of continent-derived sediments and seawater are processes that can raise seawater $\delta^{18}\text{O}$ values. Large volumes of seawater are recycled through the crust at ocean ridges and react with basalt that has a $\delta^{18}\text{O}$ of 5.5–6‰, and continental sediments have $\delta^{18}\text{O}$ values even higher (7–20‰). It is difficult to see how the oceans could have escaped these interactions, which would raise seawater δ values over geologic time. The possibility that seawater $\delta^{18}\text{O}$ increased more rapidly after the end of the Archean (fig. 11.11) is consistent with rapid continental growth in the Late Archean, which would result in more sediments delivered to the oceans after 2.5 Ga.

Carbon Isotopes in Seawater and Organic Matter

Most of the carbon in the Earth's exogenic systems is stored in sedimentary rocks with only about 0.1% in living organisms and the atmosphere–hydrosphere. Oxidized carbon occurs primarily as marine carbonates and reduced carbon as kerogen in sediments. In the *carbon cycle*, CO_2 from the oceans and atmosphere is transferred into sediments as carbonate carbon (C_{carb}) or organic carbon (C_{org}), the former of which monitors the composition of the oceans (fig. 11.12). The cycle is completed by weathering of sedimentary rocks which return CO_2 to the atmosphere, and input into the atmosphere–ocean system is largely from C_{carb} . However, ever since the origin of autotrophic life forms, this input has been augmented by C_{org} .

Because organic matter preferentially incorporates ^{12}C in comparison to ^{13}C , there should be an increase in the $^{13}\text{O}/^{12}\text{O}$ ratio (as measured by $\delta^{13}\text{C}$) in buried carbon with time. Secular trends in $\delta^{13}\text{C}$ with time since 3.8 Ga, however, are not apparent

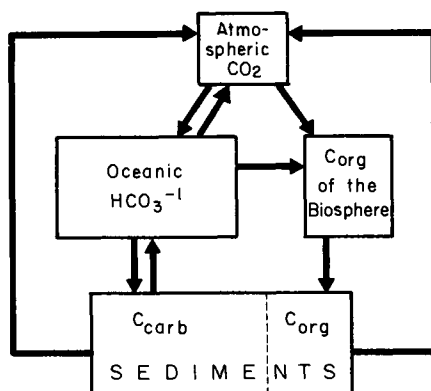


FIG 11.12 Schematic diagram of the carbon cycle. C_{carb} and C_{org} are carbonate and organic carbon, respectively.

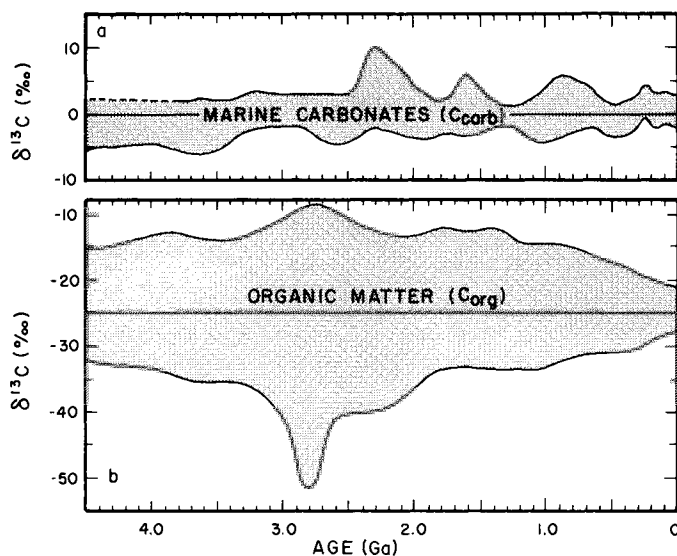


FIG 11.13 The isotopic composition of carbon in marine carbonate (C_{carb}) and organic matter (C_{org}) with geologic time. After Schidlowski et al. (1983).

in either C_{carb} or C_{org} data (Schidlowski et al., 1983) (fig. 11.13). The $\delta^{13}\text{C}_{\text{carb}}$ shows oscillations in the last 800 Ma with a minimum in the Ordovician and a maximum in the Carboniferous and Permian. Although fewer data are available from rocks older than 600 Ma, these oscillations appear to continue into the Precambrian. Highs on this curve may represent carbonates deposited in closed (or partly closed) basins, and lows reflect input of biogenic carbonate. The carbon-isotope record of organic carbon shows much greater scatter (fig. 11.13b), which probably reflects one or some combination of: (1) primary variations in the isotopic composition of parental organic matter; (2) metamorphism, which raises $\delta^{13}\text{C}_{\text{org}}$ values and probably accounts for those values greater than -20‰ ; and (3) methane-using bacteria which

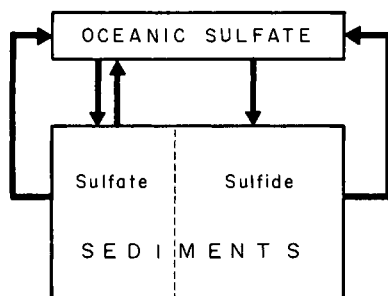


FIG 11.14 Schematic diagram of the sulfur cycle.

may explain extreme negative $\delta^{13}\text{C}_{\text{org}}$ values. Precambrian kerogens show a greater spread in $\delta^{13}\text{C}_{\text{org}}$ than Phanerozoic kerogens which lie between -21 and -32‰ , and the negative extremes (-35 to -50‰) found in some Late Archean kerogens are particularly striking. Such large negative $\delta^{13}\text{C}_{\text{org}}$ values may have been produced by bacteria involved with the CH_4 cycle in which ^{13}C is strongly preferred. The simultaneous occurrence of ^{13}C -rich and ^{13}C -poor kerogens in the Late Archean can be explained in terms of coexisting, but separate, aerobic and anaerobic basins. Because carbonates monitor exogenic carbon, $\delta^{13}\text{C}_{\text{carb}}$ values constrain $\delta^{13}\text{C}_{\text{org}}$ values of coevally produced organic carbon to an average close to -25‰ . The observed average $\delta^{13}\text{C}_{\text{org}}$ of $-27 \pm 7\text{‰}$ through time agrees well with this predicted value, indicating remarkably consistent biological activity on the Earth since 2.8 Ga.

Although several peaks in $\delta^{13}\text{C}_{\text{carb}}$ are apparent with time (fig. 11.13a), only the peak in the Carboniferous and Permian is well defined, and it appears to reflect an increase in burial rate of organic carbon. This is because organic matter selectively enriched in ^{12}C depletes seawater in this isotope, thus raising the $\delta^{13}\text{C}$ values of seawater. The maximum corresponds to rise and spread of vascular land plants in the Late Paleozoic, and such plants provided a new source of organic debris for burial (Berner, 1987). Also conducive to preservation of organic remains at this time were the vast lowlands on the supercontinent Pangaea, which appear to have been sites of widespread swamps where bacteria decay of organic matter is minimized. The drop in $\delta^{13}\text{C}$ at the end of the Permian is not understood. Pangaea did not begin to fragment until the Triassic, so the destruction of swampy environments does not seem to be the answer. Perhaps large amounts of photosynthetic O_2 generated by Carboniferous forests led to extensive forest fires which destroyed large numbers of land plants in the Late Permian.

Sulfur Isotopes in Seawater and Organic Matter

The geochemical cycle of sulfur resembles that of carbon (fig. 11.14). During this cycle ^{34}S is fractionated from ^{32}S , with the largest fractionation occurring during bacterial reduction of marine sulfate to sulfide. Sedimentary sulfates appear to record the sulfur isotopic composition of seawater. The mantle S^{34} value is near zero per mil, and bacteria reduction of sulfate strongly prefers ^{32}S , thus reducing $\delta^{34}\text{S}$ in organic sulfides to negative values (-18‰) and leaving oxidized sulfur species with approximately equivalent positive values ($+17\text{‰}$). Hence, the sulfur cycle is largely

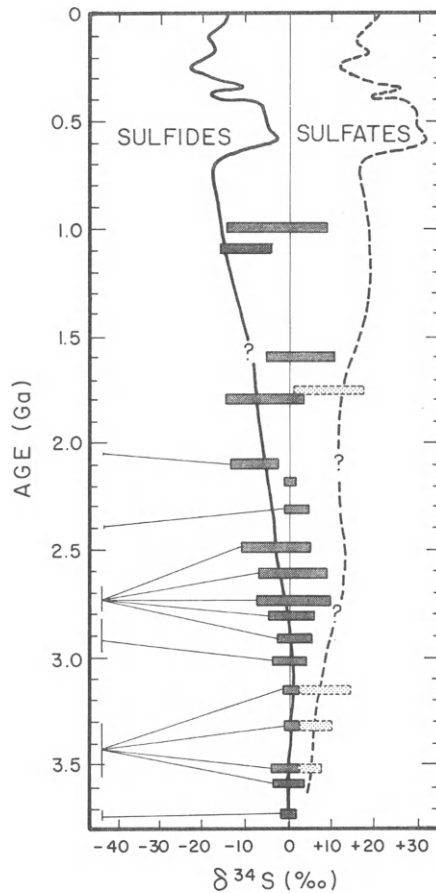


FIG 11.15 The isotopic composition of sulfur in marine sulfates and sulfides with geologic time. After Schidlowski et al. (1983).

controlled by the biosphere, and in particular by sulfate-reducing bacteria that inhabit shallow marine waters. Due to the mobility of dissolved SO_4^{-2} ions and rapid mixing of marine reservoirs, the $\delta^{34}\text{S}$ values of residual sulfate show limited variability ($+17 \pm 2\text{‰}$) compared to the wide spread of values in marine sulfides (-5 to -35‰).

The time variation of $\delta^{34}\text{S}$ in sulfates for the last 600 Ma is approximately the mirror image of the $\delta^{13}\text{C}_{\text{carb}}$ curve (figs 11.13 and 11.15). Thus a maximum ($+32\text{‰}$) is observed for $\delta^{34}\text{S}$ in the Cambrian and a minimum ($+11\text{‰}$) in the Carboniferous and Permian. This complementary relationship between the two δ values is not well understood, but undoubtedly results from interactions of the carbon and sulfur cycles (Holland et al., 1986). The overall sulfur isotope trends from 3.8 Ga to the present suggest a gradual increase in sulfate $\delta^{34}\text{S}$ and corresponding decrease in sulfide $\delta^{34}\text{S}$. Changes in $\delta^{34}\text{S}$ with time reflect: (1) the isotopic composition of sulfur input into the oceans from weathering and erosion, (2) changes in the relative proportions of sedimentary sulfide and sulfate receiving sulfur from the ocean-atmosphere system, and (3) the temperature of seawater (Schidlowski et al., 1983;

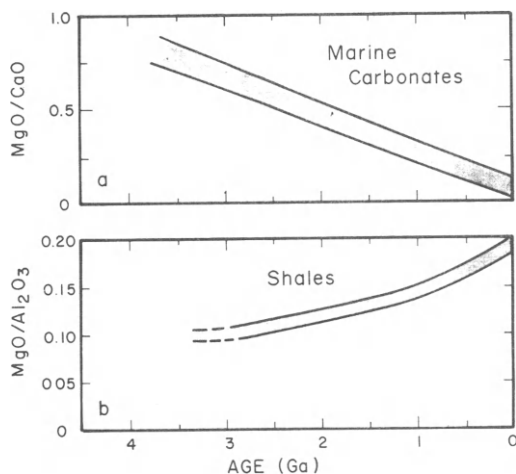


FIG 11.16 Secular variations in (a) MgO/CaO ratio in marine carbonates, and (b) MgO/Al₂O₃ ratio in shales. Data in part from Ronov and Migdisov (1971).

Ohmoto and Felder, 1987). Because weathering processes tend to average out sulfide and sulfate input, one or both of the latter two effects probably accounts for oscillations in the sulfate $\delta^{34}\text{S}$ curve with time (fig. 11.15). The isotopic composition of evaporitic sulfates should provide a measure of the intensity of bacterial sulfate in the oceans. Thus, the increasing $\delta^{34}\text{S}$ in marine sulfates since the Early Archean could be accounted for by increasing sulfide-sulfate ratios in the oceans reflecting either an increase in the numbers of sulfate-reducing bacteria or/and Archean oceans with much less sulfate than present oceans. There is uncertainty as to when the first sulfate-reducing bacteria appear in the geologic record, although most data suggest they were in existence by 2.8–3.0 Ga, and perhaps earlier. The enrichment in ^{34}S in 3.5 Ga barites compared to coeval sulfides from the same sequence of rocks is consistent with the existence of sulfate-reducing bacteria at this time. An alternate and intriguing explanation for the diverging $\delta^{34}\text{S}$ sulfide and sulfate curves with time is that the average surface temperature of the oceans has progressively cooled from 30–50°C in the Archean to present-day temperatures. This idea is based on results of experiments with living sulfate-reducing bacteria under increased water temperature (Ohmoto and Felder, 1987).

The Dolomite–Limestone Problem

It has long been recognized that the ratio of dolomite to calcite in sedimentary carbonates decreases with increasing age. This is also reflected by the MgO/CaO ratio of marine carbonates (fig. 11.16a). Most data suggest that these trends are primary and are not acquired during later diagenesis, metasomatism, or alteration. The molar MgO/MgO+CaO ratio in igneous rocks (~ 0.2) is close to that observed in pre-Mesozoic marine carbonates (0.2–0.3), which suggests that Mg and Ca were released during weathering in the same proportions they occur in source rocks prior to 100 Ma ago (Holland, 1984). The residence times of CO₂, Mg⁺² and Ca⁺² in seawater are short, and are delicately controlled by input and removal rates of diagenetic silicates and carbonates.

The origin of the secular change in the dolomite/calcite ratio with time is not fully understood and three general explanations have been proposed (Holland, 1976, 1984).

1. Change in Mg^{+2} Depository

Removal of Mg^{+2} from seawater may have changed from dominantly a carbonate depository in the Proterozoic to dominantly a silicate depository in the Late Phanerozoic. There is a tendency for the MgO/Al_2O_3 in shales to increase over the same time period (fig. 11.16b), providing some support for this mechanism, in that Mg appears to be transferred from carbonate to silicate reservoirs with time.

2. Decrease in Atmospheric CO_2

Silicate-carbonate equilibria are dependent upon the availability of CO_2 in the ocean-atmosphere system. When only small amounts of CO_2 are available in the ocean, silicate-carbonate reactions shift such that Ca^{+2} is precipitated chiefly as calcite rather than silicates. If the rate of CO_2 input into the ocean exceeds the release rate of Ca^{+2} during weathering, both Mg^{+2} and Ca^{+2} are precipitated as dolomite. Hence, a greater amount of dolomite in the Precambrian may reflect an atmosphere-ocean system with greater amounts of CO_2 , which is consistent with the atmosphere model previously discussed.

3. Mg^{+2} Loss during Seawater Recycling at Ocean Ridges

The overall decrease in the $^{87}Sr/^{86}Sr$ ratio of marine carbonates during the Paleozoic may reflect the growth of Pangaea and corresponding decreases in the amount of continental sediments delivered to the oceans (fig. 10.17b). As a consequence of this, seawater recycling at ocean ridges should have a greater influence on the composition of the oceans. Since Mg is extracted during this process, the Mg/Ca ratio in the Paleozoic oceans should decrease and calcite become more important than dolomite as a precipitating phase. This model breaks down, however, in the last 100 Ma when Pangaea fragments (and seawater $^{87}Sr/^{86}Sr$ increases), yet dolomite does not become important again.

Another attribute of carbonate deposition not well understood is the rarity of Archean carbonates. Among the causes proposed for this observation are the following: (1) the pH of Archean seawater was too low for carbonate deposition; (2) carbonates were deposited on stable cratonic shelves during the Archean and later eroded away; and (3) carbonates were deposited in deep ocean basins during the Archean and later destroyed by plate-tectonic processes. A probable higher P_{CO_2} in the Archean atmosphere has been appealed to by some to explain the sparsity of Archean carbonate (Cloud, 1968). The reasoning is that a high atmospheric P_{CO_2} results in more CO_2 dissolved in seawater, thus lowering seawater pH and allowing Ca^{+2} to accumulate because of its increased solubility. Although the P_{CO_2} in the Precambrian atmosphere was probably high, the CO_2 content is not the only factor controlling pH. Recent evidence suggests that although the CO_2 - CO_3^{-2} equilibria may have short-term control of pH, silicate equilibria have long-term control. In seawater held at a constant pH by silicate buffering reactions, the solubility of Ca^{+2} actually decreases with increasing P_{CO_2} and thus the CO_2 mechanism does not seem

capable of explaining the sparsity of Archean carbonates. As for the second explanation, carbonate rocks in Archean cratonic successions are rare, and erosion of such successions will not solve the missing Archean carbonate problem. The third possibility tentatively holds most promise. Because of the small number of stable cratons in the Archean, carbonate sedimentation may have been confined to deep ocean basins, controlled perhaps by the deposition of planktonic algae that played a role similar to that played by foraminifera today. If so, these pelagic carbonates would be largely destroyed by later plate-tectonic processes accounting for their absence in the geologic record.

Sea Level Changes

Changes in sea level leave an imprint in the geologic record in the distribution of sediments and biofacies, the areal extent of shallow seas as manifest by preserved cratonic sediments, and by calculated rates of sedimentation. A widely used approach to monitor sea level is to map successions of transgressive and regressive facies. During *transgression*, shallow seas advance on the continents, and during *regression* they retreat. Transgressions and regressions, however, are not always accompanied by rises or drops in sea level, and in order to estimate sea level changes it is necessary to correlate facies over large geographic regions involving different continents and preferably different tectonic settings (Steckler, 1984). Seismic reflection stratigraphy has also been used to detect sea level changes (Vail and Mitchum, 1979). Using this technique, onlap and offlap patterns are identified in reflection profiles and used to construct maps of the landward extent of coastal onlap from which changes in sea level are estimated.

Although a long-term component (200–300 Ma) in sea level change during the Phanerozoic seems to be agreed upon by most investigators, short-term components and amplitudes of variation remain uncertain. Major transgressions are recorded in the Early Paleozoic and in the Late Cretaceous (fig. 11.17). Estimates of the amplitude of the Cretaceous sea level rise range from 100 to 350 m above present sea level, while estimates of the Early Paleozoic rise generally fall in the range of tens of meters. Cyclicity in sea level change has been proposed by many investigators with periods in the range of 10–80 Ma. The Vail sea level curve (fig. 11.17) has been widely accepted, although the amplitudes are subject to uncertainty.

Several factors can cause rapid changes in sea level. Among the most important are tectonic controls which operate on timescales of 10^6 – 10^8 a, and glacial controls that operate on timescales of 10^4 – 10^8 a. Tectonic controls seem to produce sea level changes at rates <1 cm/ 10^3 a, while glacial controls can produce rate changes up to 10 m/ 10^3 a. With the present ice cover on the Earth, the potential amplitude for sea level rise is 200 m, of which a 70 m rise has already occurred. The short-term drops in sea level at 500, 400, and 300 Ma ago (fig. 11.17) probably reflect glaciations at these times. The rise of the Himalayas in the Tertiary produced a drop in sea level of about 50 m.

Long-term changes in sea level may be related to rates of seafloor spreading, and perhaps to changes in hotspot distribution. One of the most important causes of long-term sea level changes is the volume of ocean ridge systems. The cross-sectional area of an ocean ridge is dependent upon spreading rate, since its depth is a function

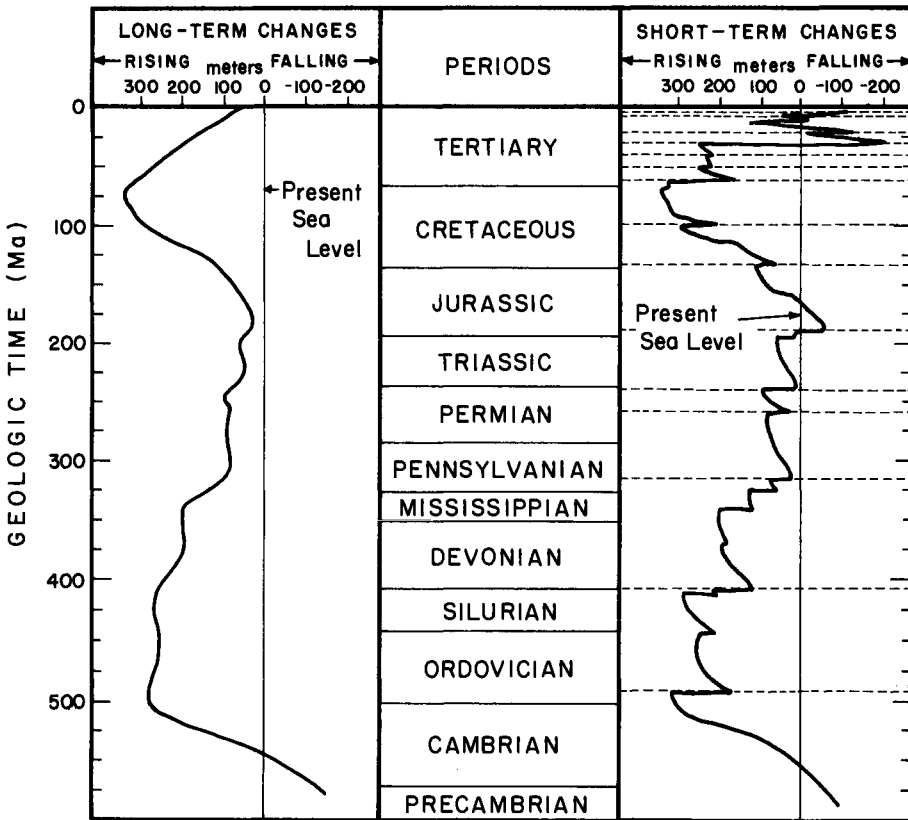


FIG 11.17 Long- and short-term changes in sea level during the Phanerozoic. Modified after Vail and Mitchum (1979).

of its age. For instance, a ridge that spreads 6 cm/a for 70 Ma has three times the cross-sectional area of a ridge that spreads at 2 cm/a for the same amount of time. The total ridge volume at any time is dependent upon spreading rate (which determines cross-sectional area) and total ridge length. Times of maximum seafloor spreading correspond well with the major transgression in the Late Cretaceous, thus supporting the ridge volume model. Calculations indicate that the total ridge volume in the Late Cretaceous could displace enough seawater to raise sea level by about 350 m. The two Phanerozoic long-term maxima in sea level, one in the Late Cambrian and the other in the late Cretaceous, follow break-up of supercontinents by about 70 Ma. This may reflect the time it takes new ocean ridge systems to propagate an increase in volume on the seafloor. Continental rifting decreases the volume of ocean basins and thus also contributes to the rise of sea level during fragmentation of supercontinents, as recorded for instance by the gradual rise in sea level during the Jurassic and Early Cretaceous.

Climatic Evolution

Introduction

To understand evolutionary changes in the Earth's climate it is necessary to review the major factors influencing modern climates. The broad climatic belts on the

Earth are latitude-dependent and reflect in part the intensity of solar radiation reaching the Earth's surface. The equatorial atmosphere is warmer than the polar atmosphere, and it is in continual motion in an attempt to homogenize temperature. Although latitudinal temperature gradients result in a strong north-south component for terrestrial winds, the Earth's rotation adds an east-west component. This results in winds blowing chiefly in a clockwise rotational pattern in the Northern Hemisphere and a counter-clockwise pattern in the Southern Hemisphere. These currents meet in the equatorial region to produce the trade winds. Wind currents, however, are modified by the amount of shallow flooding of the continents by seawater and by the distribution of mountains. The *albedo*, which is the percentage of heat reflected by the Earth's surface, is different for land and sea. Seawater has a low albedo (2–10%) and absorbs a great deal of solar heat. Land has a relatively high albedo (10–30%) and snow, ice, and clouds have very high albedos (50–85%). The oceans store heat and warm the atmosphere from below, modifying wind patterns as well as climatic patterns, and the greater the area of the oceans, the warmer the Earth's climate as a whole. Thus, times of transgression in the geologic past should also be times of warmer climates, a prediction which is supported by paleoclimatic studies.

The role of mountains in influencing climate is well known. Warm air blowing against mountains is cooled, resulting in large amounts of precipitation on the windward side, while the leeward side may receive very little precipitation. This effect, known as the rain shadow effect, can produce a humid climate on one side of a mountain range and a desert climate on the other. The highest rates of precipitation are found in the tropics where rapidly rising, warm air is cooled. Minima in precipitation occur at about 30° north and south latitude, where major deserts are found today. Increased precipitation between 40° and 50° north and south latitude reflects year-round convergence of warm subtropical air with cold polar air.

Seawater, by virtue of its low albedo and capacity to store heat, has a pronounced effect on climates. The oceans can store heat for long periods of time and act as a buffer to short-term fluctuations in climate. Oceanic surface currents are produced by winds, and their current patterns are similar to wind currents. As ocean currents move westward along the equator they absorb heat and surface evaporation rates increase. When these currents are deflected poleward by a continent, they precipitate their moisture producing humid climates along east coasts. Ocean currents may also introduce moisture into polar regions and be important in feeding glaciers. Currents that move from the poles toward the equator are cool, evaporation rates are low, and continents adjacent to these currents commonly have moderate climates. Melted polar ice also may be transported by ocean currents into low latitudes and cool oceanic temperatures.

Major worldwide changes in climate appear to be related to plate tectonics or to astronomical phenomena. Plate tectonic processes govern the size, shape, and distribution of continents which, in turn, influence the ocean-current patterns. The distribution of mountain ranges, volcanic activity, and changes in sea level are also controlled by plate tectonics, and each of these factors may affect climate. Sea level, in addition, is influenced by changes in the amount of polar ice residing in glaciers. Volcanism influences climate in two ways: (1) volcanic peaks may produce a rain

TABLE 11.5 Major Paleoclimatic Indicators

Indicator	Interpretation
1. Widespread glacial deposits (tillites, glacial pavement, dropstones, striated and faceted clasts)	Continental glaciation; Widespread cold climates
2. Coal	Abundant vegetation, moist climate, generally mid-latitudes
3. Aeolian sandstones	Arid or semi-arid; generally but not always hot or warm temperatures
4. Evaporites	Arid or semi-arid; generally but not always hot or warm temperatures
5. Redbeds	Arid to subtropical; commonly semi-arid or arid
6. Laterites and bauxites	Tropical, hot and humid climates
7. Corals and related invertebrate fauna	Subtropical to tropical, near-shore marine

shadow effect similar to other mountain ranges, and (2) volcanism can introduce large amounts of volcanic dust and CO_2 into the atmosphere, the former of which reduces solar radiation and the latter increases the greenhouse effect. Clearly, the distribution of continents and ocean basins is an important aspect of paleoclimatology, and reconstructions of past continental positions are critical to understanding ancient climatic regimes. Astronomical factors—such as changes in the Sun's luminosity, changes in shape of the Earth's orbit about the Sun, and changes in the angle between the Earth's equator and the ecliptic—may also influence climate.

The distribution of ancient climates can be constrained with *paleoclimatic indicators*, which are sediments, fossils, or other information that is sensitive to climate. A summary of major paleoclimatic indicators and their interpretation is given in table 11.5. Some sediments reflect primarily precipitation regimes rather than temperature regimes (Robinson, 1973). Coal, for instance, requires abundant vegetation and an adequate water supply, but can form at various temperature regimes, excepting arid hot or arid cold extremes. Aeolian sand dunes may form in cold or hot (or semi-arid) environments (as exemplified by modern dunes in Mongolia and the Sahara, respectively). Evaporites may form in both hot and cold arid environments, although they are far more extensive in the former. Laterites and bauxites seem to unambiguously reflect hot, humid climates. There have been false reports of continental glaciation based on erroneous identification of glacial deposits or on glacial deposits that reflect only localized glacial activity (Fairbridge, 1973). For instance, subaqueous slump, mudflow, and landslide deposits can be mistaken for tillites. No single criterion should be accepted in the identification of continental glacial deposits. Only a convergence of evidence from widespread locations such as tillites, glacial pavement, faceted and striated boulders, and glacial dropstones should be considered satisfactory. The distribution of various floral and faunal groups may also be useful in reconstructing ancient climatic belts. Modern hermatypic hexacorals, for instance, are limited to warm surface waters ($18\text{--}25^\circ\text{C}$) between latitude 38°N and 30°S (fig. 11.18). Late Jurassic hermatypic corals, plotted on the same reconstruction, are displaced northwards about 35 degrees. Because paleomagnetic and other data indicate the continents during the Late Jurassic were at about the same latitude as at present, the coral distributions indicate a warmer

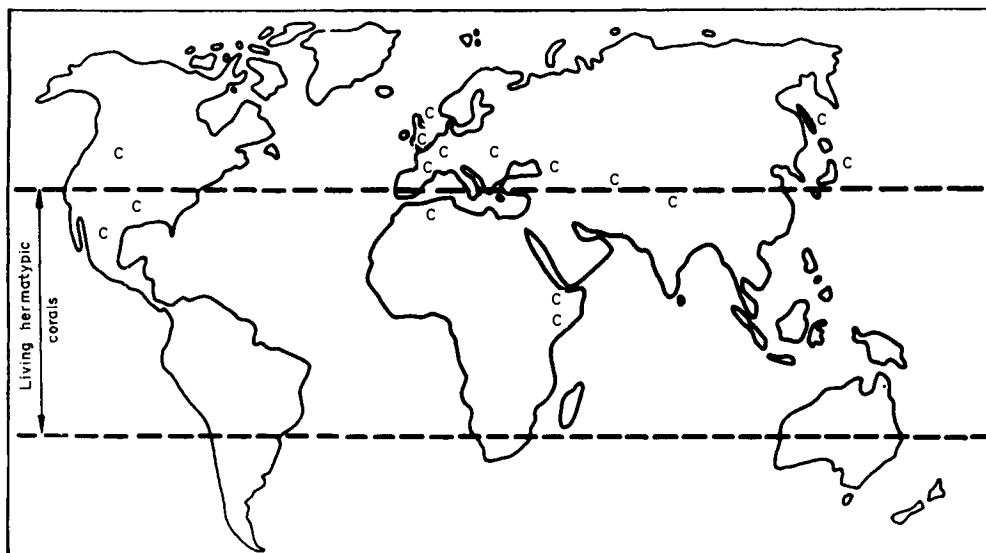


FIG 11.18 Distribution of modern and Jurassic hermatypic corals. After Frakes (1979).
C = Jurassic assemblages.

climate with tropical conditions extending to high latitudes. It is also possible to estimate paleoclimatic temperatures with oxygen isotopes in marine carbonates and cherts, provided the rocks have not undergone isotopic exchange since their deposition.

The surface temperature of the Earth appears to be related chiefly to three factors: (1) an increase of solar luminosity with time, (2) variation in albedo, and (3) variation in the greenhouse effect. Albedo and the greenhouse effect are partly controlled by other factors such as plate tectonics and global volcanism. All three factors play an important role in long-term (10^8 – 10^9 a) secular climatic changes, while plate tectonics and volcanism appear to be important in climatic changes occurring on time scales of 10^6 – 10^8 a (Crowley, 1983). Examples of tectonic changes that have occurred in the last 3 Ma that have had profound influences on climate include connection of North and South America at the Isthmus of Panama, and continued vertical uplift of the Andes and the Himalayas. On time scales $\leq 10^5$ a, climate is sensitive to astronomical phenomena (and in particular orbital perturbations of the Earth), solar variability, volcanism, and various feedback mechanisms within the land–sea–air–ice system.

Glaciations

Direct evidence for major terrestrial glaciations exists for perhaps 5–10% of Earth history. At least six major periods of continental glaciation are recognized in the geologic record (table 11.6). The earliest glaciation at 2800 Ma is represented only in South Africa in the Witwatersrand Supergroup (Harland, 1983), whereas the Huronian glaciation at 2300–2400 Ma is represented on several continents. The Huronian glaciation is not well dated, however, and may represent several glaciations of more local extent. Although the evidence for Late Proterozoic glaciations (600–900 Ma) is overwhelming, the geographic distribution and specific

TABLE 11.6 *Summary of Major Terrestrial Glaciations*

Age (Ma)	Major Locations
2800	Witwatersrand: South Africa
2300–2400	Huronian: North America, South Africa
900	Lower Congo: Africa
800	Sturtian: Australia, North America, South Africa
650	Varangian: Eurasia, South Africa, Australia
600	Late Sinian: China, North Europe, North America, South America
430–450	Late Ordovician } Gondwana
	Early Silurian }
250–380	Devonian } Gondwana
	Carboniferous }
	Permian }
0–15	Late Cenozoic, Antarctica–North America–Eurasia

In part after Harland (1983).

age of individual glaciations is not well known (Frakes, 1979; Harland, 1983). Major Late Proterozoic ice sheets were located in central Australia, west-central Africa, the Baltic Shield area, Greenland and Spitsbergen, western North America, and possibly in eastern Asia and South America. Puzzling features of the Late Proterozoic glacial deposits include their association with shallow marine carbonates suggestive of warm climatic regimes, and paleomagnetic data that indicate that Late Proterozoic continents (or supercontinent) (fig. 9.28) were at low latitudes. During the Paleozoic, Gondwana migrated over large latitudinal distances, crossing the South Pole several times. Major glaciations in Gondwana occurred in the Late Ordovician, Late Carboniferous, and Early Permian and are probably responsible for drops in sea level at these times (fig. 11.17). Minor Gondwana glaciations are also recorded in the Cambrian and Devonian. Although minor glaciations may have occurred in the Mesozoic, the next large glaciations were in the Late-Cenozoic, culminating with the multiple glaciations in the Quaternary.

Among the more important factors cited as responsible for glacial climates are decreases in solar heating due to volcanic eruptions, decreases in the input of CO₂ into the atmosphere–hydrosphere caused by decreased seafloor spreading rates (resulting in a decreased greenhouse effect), changes in continent positions and elevations caused by plate tectonics, and astronomical phenomena resulting in small changes in the Earth's orbital characteristics. Causes of the Late Archean and Early Proterozoic glaciations remain elusive. One suggested cause is rapid extraction of CO₂ from the atmosphere by photosynthetic microorganisms reducing the greenhouse effect. Calculations indicate that a drop of only 2°C is enough to precipitate an ice age (Kasting, 1987). However, such a model does not explain how glaciation should end, nor why large amounts of marine carbonate produced during the CO₂ withdrawal do not occur beneath Early Precambrian tillites. The absence of significant amounts of volcanic rock beneath Early Precambrian tillites, also does not favor volcanic dust injected into the atmosphere (reducing sunlight) as a cause of glaciation. Another possibility that has not been fully evaluated is the effect of rapid continental growth in the Late Archean on CO₂ content of the atmosphere (fig. 11.5). In particular, the Huronian glaciation(s), which are geographically widespread, follow the Late Archean rapid continental growth.

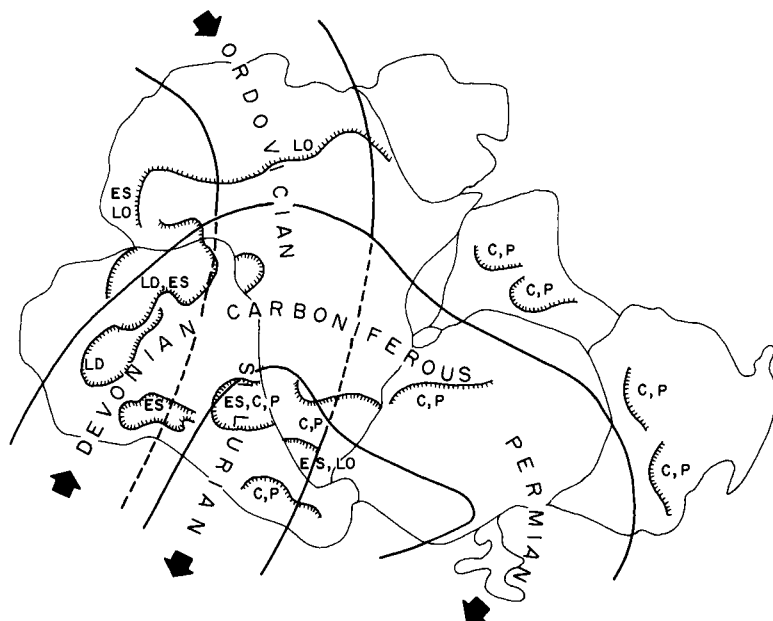


FIG 11.19 Apparent polar wandering paths and major glacial centers across western Gondwana during the Paleozoic. Modified after Caputo and Crowell (1985). Symbols: E, Early; L, Late; O, Ordovician; S, Silurian; D, Devonian; C, Carboniferous; P, Permian.

The Late Proterozoic glaciations that occurred at low latitudes are equally puzzling. Because of inadequate radiometric dates a given glaciation may not have been worldwide, but only of local or continental extent. Williams (1975) suggested that the obliquity of the Earth's orbit may have changed with time, and that during the Late Proterozoic the equator was tipped at a high angle to the ecliptic resulting in a decrease in mean annual temperature. Such a condition should favor widespread glaciation even at low latitudes. However, this model does not seem capable of explaining the angular momentum distribution in the Earth-Moon system. It is possible that a continued decrease in atmospheric CO_2 level (reducing the greenhouse effect) was responsible for initiation of the Late Proterozoic glaciations (fig. 11.5). The widespread shallow marine carbonates associated with Late Proterozoic glacial deposits are consistent with rapid CO_2 removal from the atmosphere. The problem still exists, however, as to how the glaciations end. Perhaps renewed ocean ridge activity accompanying the break-up of a Late Proterozoic supercontinent pumped sufficient CO_2 into the atmosphere to increase greenhouse warming.

Both Early and Late Paleozoic glaciations appear to be related to plate tectonics. During both of these times, Gondwana was glaciated as it moved over the South Pole (Caputo and Crowell, 1985). Major glaciations occurred in central and northern Africa in the Late Ordovician when Africa was centered over the South Pole (fig. 11.9). As Gondwana continued to move over the pole, major ice centers shifted into southern Africa and adjacent parts of South America in the Early Silurian. Gondwana again drifted over the South Pole beginning with widespread glaciation in the Late Devonian and Early Carboniferous in South America and Central Africa,

shifting to India, Antarctica, and Australia in the Late Carboniferous and Permian. Paleozoic glaciations can also be related to prevailing wind and ocean currents associated with the drift of Gondwana over the pole.

The next major glaciations began in the Mid-Tertiary and were episodic with interglacial periods. Between 30 and 40 Ma ago the Drake Passage between Antarctica and South America opened, and the Antarctic circumpolar current was established, thus thermally isolating Antarctica (Crowley, 1983). The final separation of Australia and Antarctica at about this time may also have contributed to cooling of Antarctica by inhibiting poleward transport of warm ocean currents. Glaciation appears to have begun in Antarctica 12–15 Ma ago, and by 3 Ma ago in the Northern Hemisphere. Initiation of Northern Hemisphere glaciation may have been in response to formation of Panama which caused the Gulf Stream, that had previously flowed westward about the equator, to flow north along the coasts of North America and Greenland. This shift in current brought warm, moist air into the Arctic, which precipitated and began to form glaciers.

Three factors have been suggested to be important in controlling the multiple glaciations that characterize the Pleistocene: (1) episodic variations in volcanic activity, (2) cooling and warming of the oceans, and (3) cyclic changes in the Earth's rotation and orbit. During the Pleistocene there is not a close correlation between volcanic activity and glaciation, not favoring the first idea. Cyclical changes in temperature of the oceans caused by changes in the balance between evaporation and precipitation may be in part responsible for multiple glaciations. A typical cycle might start as moist, warm air moves into the Arctic region precipitating large amounts of snow that result in glacial growth. As oceanic temperatures drop in response to glacial cooling, evaporation rates from the ocean decrease and the supply of moisture to glaciers also decreases. Hence, glaciers begin to retreat, climates warm, and an interglacial stage begins. The warming trend has a negative feedback and eventually increases evaporation rates, and moisture is again supplied to glaciers as the cycle starts over.

Most investigators agree that cyclical changes in the shape of the Earth's orbit about the Sun and in the inclination and wobble of the Earth's rotational axis can bring about cyclical cooling and warming of the Earth. Today, the Earth's equatorial plane is tilted at 23.5° to the ecliptic, but this angle varies from 21.5° to 24.5° with a period of about 41,000 years. Decreasing the inclination results in cooler summers in the Northern Hemisphere and favors ice accumulation. The Earth's orbit also changes in shape from a perfect circle to a slightly ellipsoidal shape with a period of 92,000 years, and the ellipsoidal orbit results in warmer than average winters in the Northern Hemisphere. The Earth also wobbles as it rotates with a period of about 20,000 years and the intensity of seasons varies with the wobble. When variations in oxygen isotopes or fauna are examined in deep-sea cores, all three of these periods, known as the *Milankovitch periods*, can be detected, with the 100,000-year period being prominent. These results suggest that, on time scales of 10,000–400,000 years, variations in the Earth's orbit are the fundamental cause of Pleistocene glaciations (Imbrie, 1984). Spectral analyses of $\delta^{18}\text{O}$ and orbital time series strongly support the model. On time scales $>400,000$ a, plate tectonic changes in continent-ocean geometry and associated volcanic eruptions appear to be the primary controls on climate.

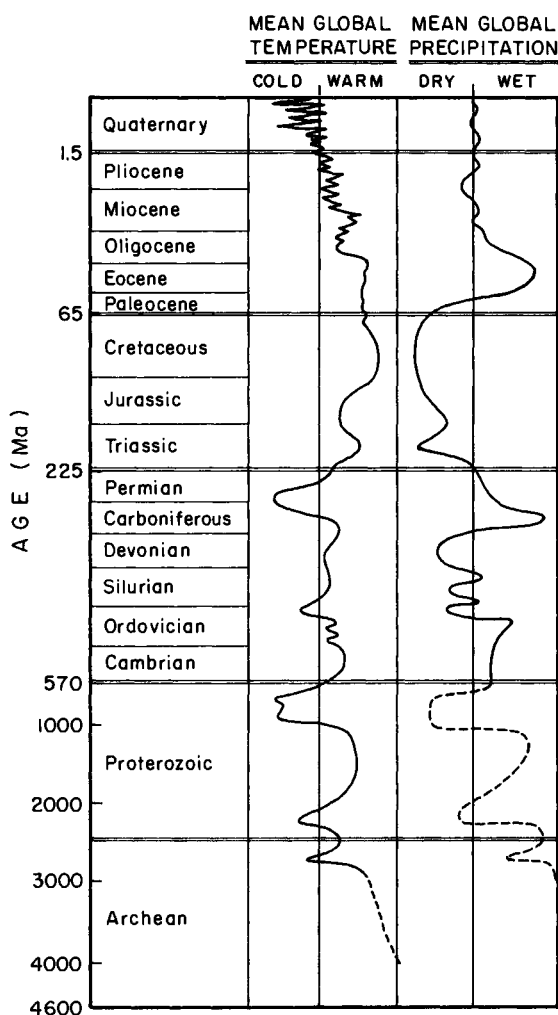


FIG 11.20 Generalized average surface temperature and precipitation history of the Earth. After Frakes (1979).

Paleoclimates

Using paleoclimatic indicators, together with results of models for atmospheric evolution, it is possible to characterize average surface temperatures and precipitation regimes with geologic time. The abundance of CO_2 in the Archean atmosphere resulted in greenhouse warming to average surface temperatures greater than at present (fig. 11.20). Few climatic indicators are preserved in Archean supracrustal successions. Archean banded-iron formation is commonly thought to have been deposited in shallow marine waters, and hence implies low wind velocities to preserve the remarkable planar stratification. This is consistent with a CO_2 -rich atmosphere, in which differences in temperature gradient with latitude, and hence wind velocities, are small (Frakes, 1979). The earliest evaporites and stromatolitic carbonates appear at about 3.5 Ga and also favor generally warm climates. The first

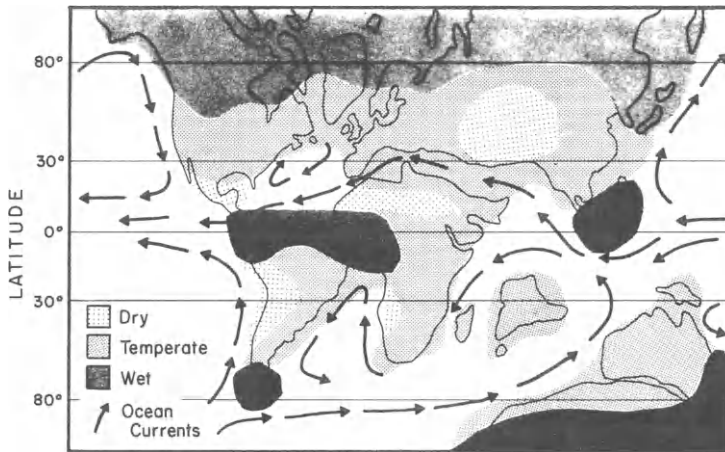


FIG 11.21 Distribution of major climatic regimes and ocean currents during the Late Cretaceous. In part from Hallam (1985).

glaciations at 2.8 and 2.3–2.4 Ga reflect falling atmospheric CO_2 levels in response to increasing numbers of photosynthetic microorganisms (chiefly cyanobacteria). The relative abundance of shallow marine carbonates and the preservation of lateritic paleosols in the Proterozoic suggest a continuation of warm, moist climates after the end of the Archean. The increasing proportion of evaporites and redbeds after 2 Ga ago appears to record the increasing importance of semi-arid to arid climatic regimes. Colder climates as recorded by widespread glaciations alternated with warm, moist climates as reflected by marine carbonates in the Late Proterozoic.

Most of the Middle and Lower Paleozoic are characterized by climates that, on the whole, were warmer than present climates, while Late Paleozoic climates show wide variations. Europe and North America underwent relatively small changes in climate during the Paleozoic, while Gondwana underwent several major changes. This is due to the fact that the continental fragments comprising Laurasia remained at low latitudes during the Paleozoic, while Gondwana migrated over large latitudinal distances. Major glaciations occurred in Gondwana in Late Ordovician, Late Carboniferous, and Early Permian.

The Mesozoic is characterized by warm, dry climates that extend far north and south of equatorial regions. Paleontologic and paleomagnetic data also support this interpretation. Initial Triassic climates were cool and humid like the Late Paleozoic, and they were followed by a warm, drying trend until Late Jurassic. The largest volumes of preserved evaporites are in the Late Triassic and Early Jurassic, and reflect widespread arid climates on Pangaea. Most of the Cretaceous is characterized by warm, commonly humid climates over the continents regardless of their latitude, with tropical climates extending from 45°N to perhaps 75°S , Hallam (1965) (fig. 11.21). Glacial climates are of local extent and only in polar regions. Mean annual surface temperatures during the Mid-Cretaceous, as determined from oxygen isotope data, are $10\text{--}15^\circ\text{C}$ higher than today. The Late Cretaceous is a time of maximum transgression of the continents, and the warm Cretaceous climates appear to reflect low albedo of the oceans (Crowley, 1983). The Early Tertiary was a time of declining temperatures, significant increased precipitation, and falling sea level.

Middle and Late Tertiary are characterized by lower and more variable temperature and precipitation regimes. These variations continued and became more pronounced in the Quaternary, leading to the alternating glacial and interglacial epochs. Four major Quaternary glaciations are recorded, with the intensity of each glaciation decreasing with time: the Nebraskan (1.3–1.46 Ma), Kansan (0.7–0.9 Ma), Illinoian (400,000–550,000 a), and Wisconsin (10,000–80,000 a). Each glaciation lasted 100 to 200 thousand years, with interglacial periods of 200 to 400 thousand years. Smaller glacial cycles with periods of the order of 20 to 40 thousand years are superimposed on the larger cycles.

Living Systems

General Features

Although the distinction between living and non-living matter is obvious for most objects, it is not so easy to draw this line between some unicellular organisms, on one hand, and large non-living molecules such as amino acids, on the other. It is generally agreed that *living matter* must be able to reproduce new individuals, it must be capable of growing by using nutrients and energy from its surroundings and it must respond in some manner to outside stimuli. Another feature of life is its chemical uniformity. Despite the great diversity of living organisms, all life is composed of a few elements (chiefly C, O, H, N, and P) that are grouped into nucleic acids, proteins, carbohydrates, and fats and a few other minor compounds. This suggests that living organisms are related, and that they probably had a common origin. Reproduction is accomplished in living matter at the cellular level by two complex nucleic acids—*RNA* and *DNA*. *Genes* are portions of DNA molecules that carry specific hereditary information. Three components are necessary for a living system to self-replicate: RNA and DNA molecules, which provide a list of instructions for replication; proteins that promote replication; and a host organ for the RNA–DNA molecules and proteins. The smallest entities capable of replication are amino acids.

Origin of Life

Perhaps no other subject in geology has been investigated more than the origin of life (Kvenvolden, 1974). It has been approached from many points of view. Geologists have searched painstakingly for fossil evidence of the earliest life, and biologists and biochemists have provided a variety of evidence from experiments and models that must be incorporated into any model for the origin of life. One major question that has not been completely answered to everyone's satisfaction is whether life began on the Earth or on some other body, to be later carried to the Earth.

Although numerous models have been proposed for the origin of life, two environmental conditions are a prerequisite to all models: (1) the elements and catalysts necessary for the production of organic molecules must be present, and (2) free oxygen, which would oxidize and destroy organic molecules, must not be present. In the past the most popular models for the origin of life involve a primordial “soup” rich in carbonaceous compounds produced by inorganic processes. Reactions in this soup, promoted by catalysts such as lightning or UV

radiation, result in production of organic molecules. Primordial soup models, however, seem unnecessary in terms of rapid degassing of the Earth prior to 4 Ga. Rapid recycling of the early oceans through ocean ridges would not allow concentrated "soups" to survive except perhaps locally in evaporite basins after 4 Ga ago. Because the chances of organic molecules being present in sufficient amounts, in the correct proportions, and in the proper arrangement are very remote it would seem that the environment in which life formed would have been very widespread in the Early Archean. Possibilities include volcanic environments and hydrothermal vents along ocean ridges.

Simple amino acids have been formed in the laboratory under a variety of conditions. The earliest experiments were those of Miller (1953), who sparked a hydrous mixture of H_2 , CH_4 , and NH_3 to form a variety of organic molecules including four of the 20 amino acids comprising proteins. Similar experiments, using both sparks and UV radiation in gaseous mixtures of H_2O , CO_2 , N_2 and CO (a composition more in line with that of the Earth's early degassed atmosphere) also resulted in production of amino acids, HCN , and formaldehydes, the latter of which can combine to form sugars. Heat also may promote similar reactions. Although it seems relatively easy to form amino acids and other simple organic molecules, just how these molecules combined to form the first complex molecules like RNA, and then evolved into living cells, remains largely unknown. Recent studies of RNA suggest that it may have played a major role in the origin of life. RNA molecules have the capability of splitting and producing an enzyme that can act as a catalyst for replication (Zaug and Cech, 1986). Necessary conditions for the production of RNA molecules in the Early Archean include a supply of organic molecules, a mechanism for molecules to react to form RNA, a container mineral to retain detached portions of RNA so they can aid in further replication, a mechanism by which some RNA can escape to colonize other populations, and some means of forming a membrane to surround a proto-cell wall (Nisbet, 1986). During the Archean, hydrothermal systems on the seafloor may have provided these conditions (Gilbert, 1986). In laboratory experiments, RNA splitting occurs at temperatures around $40^\circ C$ with a pH varying from 7.5 to 9 and with Mg^{+2} present in solution. These are similar to conditions that exist near vent exits of oceanic hydrothermal systems (fig. 11.22). Cairns-Smith (1982) has suggested that clays may play an even more important role, and actually provided a means for replication. The first genes may have formed in the interlayer positions of clays and the first replication may have been inorganic growth of the clays. Later organic compounds absorbed in the clays reacted to form RNA and, through natural selection, RNA molecules eventually disposed of their clay hosts. Another possible catalyst is zeolites which possess a wide variety of pores of different shapes and sizes that permit small organic molecules to pass through, while excluding or trapping larger molecules (Nisbet, 1986). They are also a characteristic secondary mineral around hydrothermal plumbing systems (fig. 11.22). The significance of variable-sized cavities in zeolites is that a split-off RNA molecule may be trapped in such a cavity where it can aid in replication of the parent molecule. Although the probability is small, it is possible that the first polynucleotide chain formed somewhere in an early hydrothermal vent of the seafloor.

The Early Archean RNA "world" may have existed in clay minerals, zeolites and the pore spaces of altered volcanic rocks. Because hydrothermal systems appear to

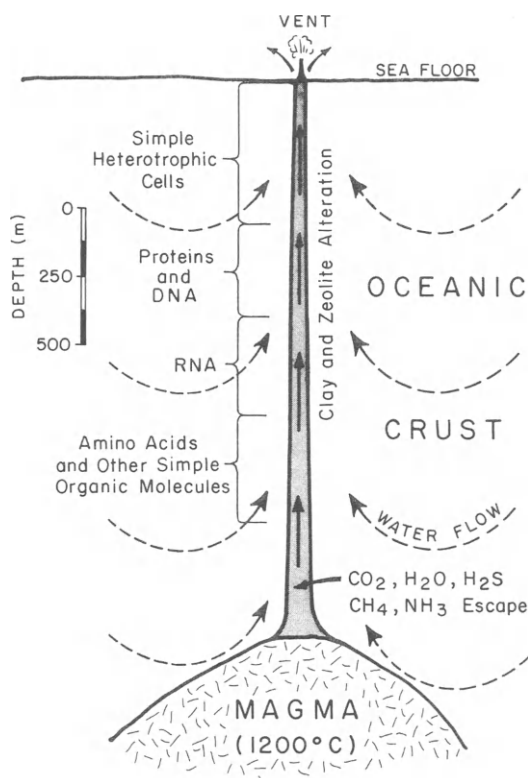


FIG 11.22 Idealized cross-section of Archean ocean-ridge hydrothermal vent showing possible conditions for the formation of life.

have lifetimes of the order of 10^4 – 10^5 a, at any given location, RNA populations must have evolved rapidly into cells, or more likely they were able to colonize new vent systems. The next stage in replication may have been the development of proteins from amino acids that were synthesized from CH_4 and NH_3 in hydrothermal vents. Later still, DNA must form and take over as the primary genetic library (Gilbert, 1986). The next stages of development, although poorly understood, seem to involve production of membranes which allow managing of energy supply and metabolism, both of which are essential for the development of a living cell.

The first cells were primitive in that they had poorly developed metabolic systems and survived by absorbing a variety of nutrients from their surroundings. They must have obtained nutrients and energy from other organic substances by fermentation, which occurs only in *anaerobic* (oxygen-free) environments. *Fermentation* involves the breakdown of complex organic compounds into simpler compounds that contain less energy, and the energy liberated during fermentation is used by organisms to grow and reproduce. Cells that obtain their energy and nutrients from their surroundings by fermentation or chemical reactions are known as *heterotrophs*, in contrast to *autotrophs* which are capable of manufacturing their own food. Two types of anaerobic cells evolved from DNA replication, probably by 4 Ga ago. The most primitive group, the *archaebacteria*, use RNA in the synthesis of proteins, whereas

the more advanced group, the *eubacteria*, have advanced replication processes and may have been the first photosynthesizing organisms.

Rapid increases in the numbers of early heterotrophs may have led to severe competition for food supplies. Selection pressures would tend to favor mutations that enabled heterotrophs to manufacture their own food, and thus, to become autotrophs. The first autotrophs probably appeared by 3.8 Ga as cyanobacteria (see fig. 11.27). These organisms produced their own food by photosynthesis, perhaps using H_2S rather than water since free oxygen, which is liberated during normal photosynthesis, is lethal to anaerobic cells. Just how photosynthesis evolved is unknown, but perhaps the supply of organic substances and chemical reactions became less plentiful as heterotrophs increased in numbers and selective pressures increased to develop alternate energy sources. Sunlight would be an obvious source to exploit. H_2S may have been plentiful from hydrothermal vents or decaying organic matter on the seafloor, and some cells may have developed the ability to use this gas in manufacturing food. As these cells increased in numbers the amount of H_2S would not be sufficient to meet their demands, and selective pressures would be directed toward alternate substances, of which water is the obvious candidate. Thus, mutant cells able to use water would out-compete forms only able to use H_2S .

The First Evidence of Life

Two lines of evidence are available for the recognition of the former existence of living organisms in Early Archean rocks: microfossils and organic geochemical evidence (Schopf et al., 1983). Many microstructures preserved in rocks can be mistaken for cell-like objects (inclusions, bubbles, microfolds, etc.), and progressive metamorphism can produce structures that look remarkably organic and at the same time can destroy real microfossils. For these reasons caution must be exercised in accepting microstructures as totally biologic.

Carbonaceous compounds in rocks are of two types—extractable (amino acids, hydrocarbons, sugars etc.) and nonextractable (kerogen). Although the extractable components from Archean sediments are similar (or indistinguishable) to those in modern organisms, it is not always clear when they were introduced into the rock system. Some may date to the time of sedimentation, while others were introduced by secondary processes in the recent geologic past. The insoluble carbonaceous compounds in kerogen, although very likely formed at the time a rock is deposited, may be of biologic or nonbiologic origin, or both. To date, analytical methods are unable to distinguish between these alternatives.

The oldest well-described assemblage of microfossil-like structures comes from cherts in the Barberton greenstone belt in South Africa (~ 3.4 Ga) and the oldest unambiguous structures of organic origin are 3.5-Ga-old stromatolites from the Pilbara region of Western Australia (Walter, 1983) (fig. 11.23). Three types of microstructures, ranging in size from $<1\ \mu\text{m}$ to $\sim 20\ \mu\text{m}$, have been reported from the Barberton sequence and from other Archean sediments. These are rod-shaped bodies, filamentous structures, and spheroidal bodies. The spheroidal bodies are similar to alga-like bodies from Proterozoic assemblages and are generally interpreted as such. Of the two known types of cells, prokaryotic and eukaryotic, only prokaryotic types appear to be represented among Archean microfossils. *Prokaryotes*

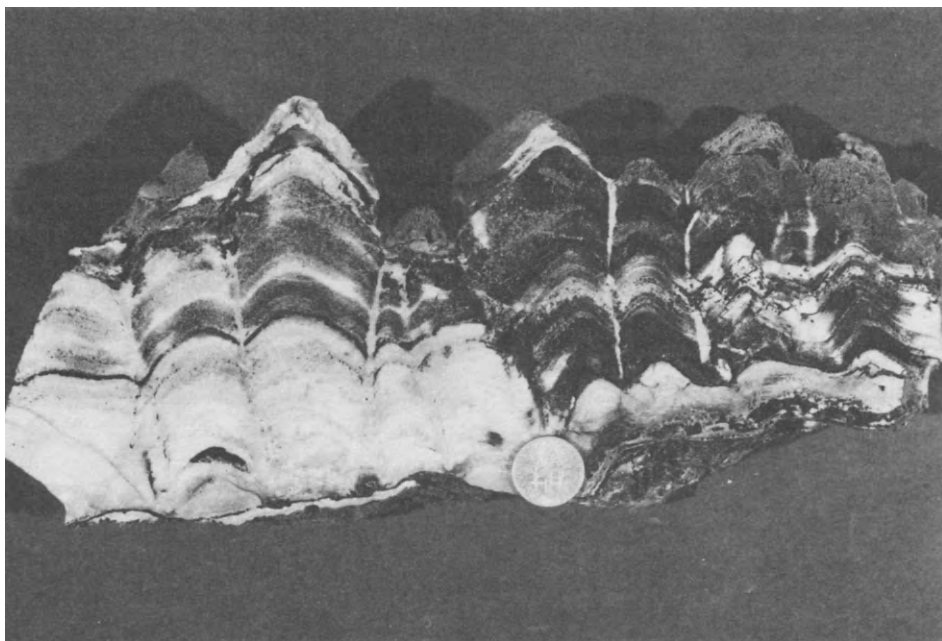


FIG 11.23 3.5-Ga-old stromatolites from the Pilbara region in Western Australia.
Courtesy of D. R. Lowe.

are primitive cells which lack a cell wall around the nucleus and are not capable of cell division; *eukaryotes* possess these features and hence are capable of transmitting genetic coding to various cells and to descendants.

Stromatolites

Stromatolites are finely laminated sediments composed chiefly of carbonate minerals that have formed by the accretion of both detrital and biochemical precipitates on successive layers of microorganisms (commonly cyanobacteria). They exhibit a variety of domical forms and range in age from about 3.5 Ga in Western Australia (fig. 11.23) to modern. Two parameters are especially important in stromatolite growth: water currents and sunlight. There are serious limitations to interpreting ancient stromatolites in terms of modern ones, however. First of all, modern stromatolites are not well understood and occur in a great variety of aqueous environments (Walter, 1983). The distribution in the past is also controlled by the availability of shallow stable shelf environments, the types of organisms producing the stromatolites, the composition of the atmosphere, and perhaps the importance of burrowing animals. It is possible to use stromatolites to distinguish deep-water from shallow-water deposition, since reef morphologies are different in these environments.

Early Archean stromatolites were probably built by photoautotrophs with mucus sheaths. These microbes were able to cope with high salinities, desiccation, and high sunlight intensities as indicated by their occurrence in evaporitic cherts. Late

Archean stromatolites are known from both lacustrine and marine environments, and some are very similar to modern stromatolites, suggesting they were constructed by cyanobacteria. Early Proterozoic stromatolites appear to have formed in peritidal and relatively deep subtidal environments.

Stromatolites increased in number and complexity from 1.6 Ga to about 800 Ma ago, when they decreased rapidly. It is likely that the abrupt decrease in numbers of stromatolites in the Late Precambrian reflects increased competition with metazoans. The extent to which stromatolites can be used to establish a worldwide Proterozoic biostratigraphy is a subject of controversy which revolves around the role of environment and diagenesis in determining stromatolite shape and the development of an acceptable taxonomy. Because the growth of stromatolites is at least partly controlled by organisms, it should in theory be possible to construct a worldwide biostratigraphic column. Another controversial subject is that of how stromatolite height is related to tidal range. Cloud (1986b) suggests that the height of intertidal stromatolites at maturity reflects the tidal range, whereas Walter (1977) suggests that the situation is much more complex. The distribution of laminations in stromatolites also has been suggested as a means of studying secular variation of the Earth-Moon system (Chapt.2).

Evolution and Biological Benchmarks

The Precambrian

Many benchmarks have been recognized in the evolution of microorganisms during the Precambrian (Schopf et al., 1983) (fig. 11.24). The first is that life on Earth formed prior to 3.5 Ga ago and, considering the evolutionary level of the earliest microfossils, probably by 4 Ga ago. Anaerobic fermenting heterotrophs and both oxygen-producing (photoautotrophs) and non-oxygen-producing autotrophs appear to have been present by 3.5 Ga. Aerobic photoautotrophs, which are forms capable of living in the presence of free O₂, and sulfate-reducing bacteria were probably in existence by 2.8 Ga. It was not until about 2 Ga, however, that microbes entirely dependent on the use of molecular O₂ appeared in the geologic record. These did not become widespread until 1.5–1.7 Ga. Eukaryotes probably appeared about 1.5 Ga ago, but did not become widespread and highly diversified until 1.0–1.2 Ga. Their origin and evolution was highly dependent upon increasing O₂ levels in the atmosphere.

By 1000 Ma the ozone screen in the upper atmosphere had developed to such an extent that most UV radiation did not reach the Earth's surface, and thus shallow-water environments were rapidly invaded by planktonic organisms. At about 700 Ma, *metazoans* (multicellular animals) evolved and rapidly diversified into a variety of ecological niches. Late Proterozoic soft-bodied metazoans known as the *Ediacaran assemblage* are known from most continents. Ediacarans become particularly widespread between 500 and 700 Ma and appear to be primarily planktonic or benthic forms. They represent at least 25 species, 15 genera and 3 phyla. Approximately two-thirds of the group appear to represent coelenterates, 25% annelids, 5% arthropods, and the remainder are of unknown affinity.

The Phanerozoic

Bioturbation or the churning of subaqueous sediments by burrowing organisms

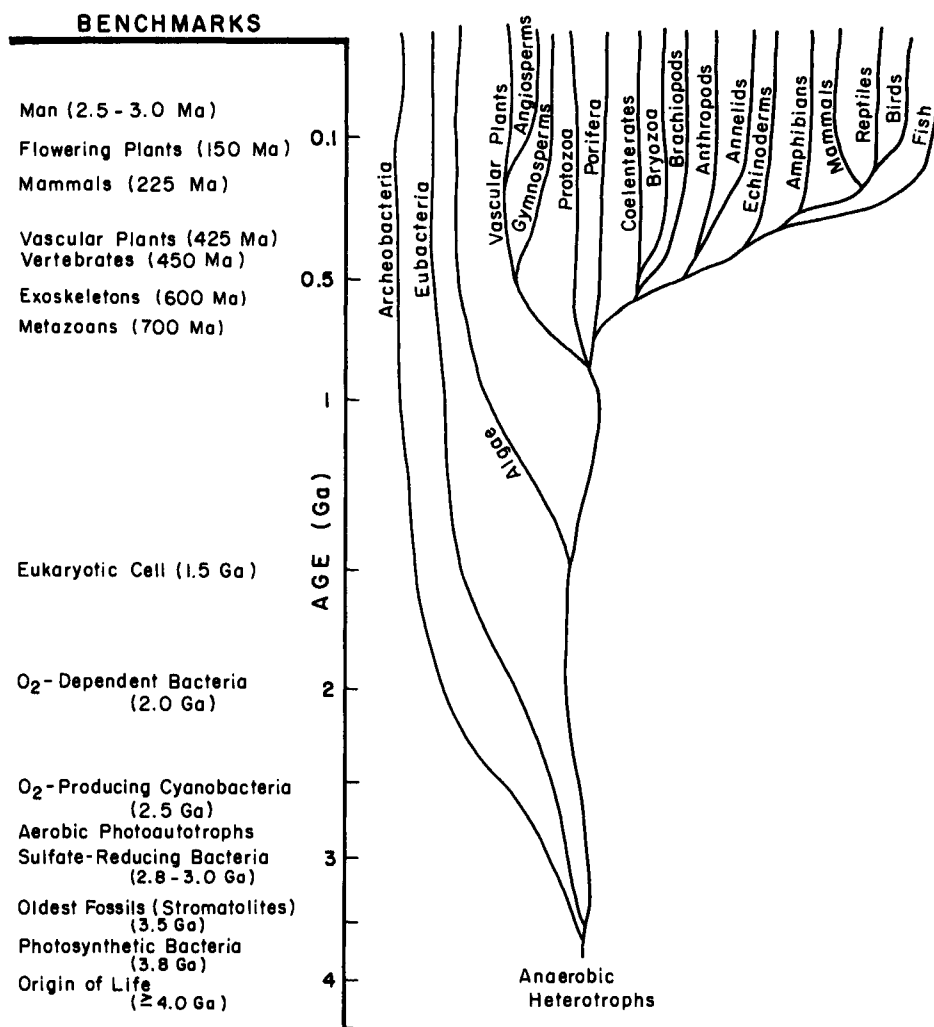


FIG 11.24 Evolutionary benchmarks and phylogeny of major groups of living organisms.

began in the Early Cambrian. Two biological inventions permitted organisms to invade sediments (Fischer, 1984). First, the development of exoskeletons allowed organisms (such as trilobites) to dig by using appendages and second, the appearance of coelomes permitted worm-like organisms to penetrate sediments. One of the immediate results of bioturbation was the return of buried organic matter to the carbon cycle, and hence a decrease in the loss of O₂ due to decay. The appearances of *calcareous* and *siliceous skeletons* in the Cambrian were also major evolutionary advances.

All of the major invertebrate phyla (except the Protozoa) made their appearances in the Early Paleozoic (fig. 11.24). This is primarily a time of primitive, shallow-marine life in which attached forms predominate. Cambrian faunas are dominated by trilobites (~60%) and brachiopods (~30%). By Late Ordovician, most of the

common invertebrate classes that occur in modern oceans were well established. Trilobites reached the peak of their development during the Ordovician, with a great variety of shapes, sizes, and shell ornamentation. Bryozoans, which represent the first attached communal organisms, first appeared in the Ordovician. Graptolites, cephalopods, crinoids, echinoderms, molluscs, and corals also began to increase in numbers at this time. Vertebrates first appeared during the Ordovician and Silurian as primitive fish-like forms without jaws. Marine algae and bacteria continued to be the important plant forms during the Early Paleozoic.

The Late Paleozoic is a time of increasing diversification of plants and vertebrates and of decline in many invertebrate groups. Brachiopods, coelenterates, and crinoids all increase in abundance in the Late Paleozoic, followed by a rapid decrease in numbers at the end of the Permian. The end of the Paleozoic was also a time of widespread extinction, with trilobites, eurypterids, fusulines, and many corals and bryozoa becoming extinct. Insects appeared in the Late Devonian. Fish greatly increased in abundance during the Devonian and Mississippian, amphibians appeared in the Mississippian and reptiles in the Pennsylvanian. Plants greatly increased in numbers during the Late Paleozoic as they moved into terrestrial environments. Psilopsids are most important during the Devonian, with lycopsids, ferns, and conifers becoming important thereafter. Perhaps the most important evolutionary event in the Paleozoic was the development of vascular tissue in plants, which made it possible for land plants to survive under extreme climatic conditions. Seed plants also began to become more important relative to spore-bearing plants in the Late Paleozoic and Early Mesozoic. The appearance and rapid evolution of amphibians in the Late Paleozoic was closely related to the development of forests, which provided protection for these animals. The appearance for the first time of shell-covered eggs and of scales (in the reptiles) allowed vertebrates to adapt to a greater variety of climatic regimes.

During the Mesozoic, gymnosperms rapidly increased in numbers, with cycads, ginkgoes, and conifers being most important. During the Lower Cretaceous, angiosperms (flowering plants) made their appearance and rapidly grew in numbers thereafter. The evolutionary success of flowering plants is due to the development of a flower and enclosed seeds. Flowers attract birds and insects that provide pollination, and seeds may develop fleshy fruits which, when eaten by animals, can serve to disperse the seeds. Marine invertebrates, which decrease in numbers at the end of the Permian, make a comeback in the Mesozoic (such as bryozoans, molluscs, echinoderms, and cephalopods). Gastropods, pelecypods, foraminifera, and coiled cephalopods are particularly important Mesozoic invertebrate groups. Arthropods in the form of insects, shrimp, crayfish, and crab also rapidly expand during the Mesozoic. Mesozoic reptiles are represented by a great variety of groups, including dinosaurs. Dinosaurs are of herbiferous and carnivorous types, as well as marine and flying forms. Birds and mammals evolve from reptilian ancestors in the Lower Jurassic. The development of mammals is a major evolutionary breakthrough, in that their warm-blooded nature allowed them to adapt to a great variety of natural environments (including marine), and their increased brain size allowed them to learn more rapidly than other vertebrates. During the Cenozoic, mammals evolved into large numbers of groups filling numerous ecological niches. Man evolved in the anthropoid group about 3 Ma ago. The vertebrate groups characteristic of the

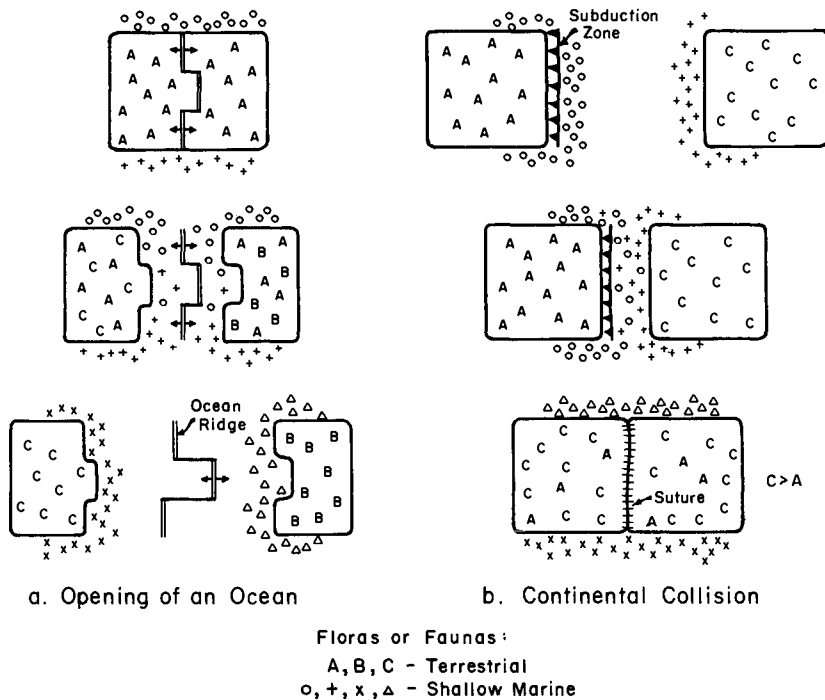


FIG 11.25 Diagrammatic illustration of evolutionary changes accompanying and following the opening of an ocean basin (a) and continental collision (b).

Mesozoic continued to increase in numbers and angiosperms expanded exponentially.

Plate Tectonics and Evolution

Plate tectonics has played a major role in the evolution and extinction of some groups of organisms. Modern ocean basins are effective barriers not only to terrestrial organisms (animal and plant) but also to most marine organisms. Larval forms of marine invertebrates, for instance, can only survive several weeks, which means they can travel only 2000–3000 km with modern ocean-current velocities. Hence, the geographic distribution of fossil organisms provides an important constraint on the sizes of oceans between continents in the geologic past. As illustrated in figure 11.25, when a continent fragments, organisms that cannot cross the growing ocean basin become isolated, which can lead to the evolution of diverse groups on each of the separating continents (Hallam, 1974). This affects both marine and terrestrial organisms. Examples are the diversification and specialization of mammals in South America and in Africa following the opening of the South Atlantic in the Late Cretaceous. Similar changes occurred in bivalves of East Africa and India after the separation of these continents in the Cretaceous and Tertiary. Continental collision, on the other hand, removes an oceanic barrier and formerly separate faunal and floral groups mix and compete for the same ecological niches (fig. 11.25b). This leads to extinction of groups that do not successfully compete. Collision also creates

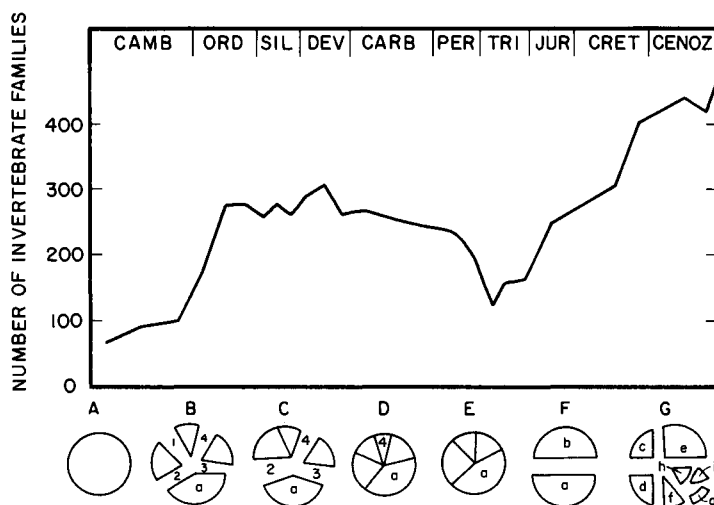


FIG 11.26 Correlation of diversity of invertebrates with time and with patterns of continental fragmentation and assembly. After Valentine and Moores (1970). Events: (A) Pangaea; (B) fragmentation of Pangaea to produce pre-Caledonian (1), pre-Appalachian (2), pre-Hercynian (3), and pre-Uralian (4) Oceans; (C) suturing of Caledonian–Acadian orogenic systems; (D) suturing of Appalachian and Hercynian systems; (E) suturing of Urals–reassembly of Pangaea; (F) opening of the Tethys Sea; (G) closing of Tethys Sea, reopening of Atlantic, fragmentation of Gondwana. Continents: (a) Gondwana; (b) Laurasia; (c) North America; (d) South America; (e) Eurasia; (f) Africa; (g) Antarctica; (h) India; (i) Australia.

a new land barrier between coastal marine populations. If these marine organisms cannot migrate around the continent (due perhaps to climatic barriers), diversification occurs and new populations evolve along opposite coastlines. During the Ordovician collisions in the North Atlantic (Chapt. 8), many groups of trilobites, graptolites, corals, and brachiopods became extinct. The creation of new arc systems linking continents has the same effect as a collision. For instance, when the Panama arc was completed in the Late Tertiary, mammals migrated between North and South America, which led to the extinction of large endemic mammals in South America. Panama also separated marine populations which led to divergence of Pacific and Caribbean marine organisms. It has been recognized for some time that the overall diversity patterns of marine invertebrates during the Phanerozoic can be related to the aggregation and fragmentation of Pangaea (Valentine, 1973) (fig. 11.26). The initial rise in number of marine invertebrate families in the Late Cambrian probably reflects filling of new ecological niches by the increasing metazoan populations. Successive collisions of continents in the Late Paleozoic (C,D) is accompanied by a decline in invertebrates. This is followed by a rapid rise in invertebrate numbers during the Mesozoic and Cenozoic as Pangaea fragmented into numerous continents (F,G).

The biogeographic distribution of Cambrian trilobites supports the existence of several continents separated by major ocean basins during the Early Paleozoic (fig. 11.27). Major faunal province boundaries commonly correlate with suture zones between continental blocks brought together by later collisions. Wide oceans are implied in the Cambrian between North America and Europe, Siberia and Europe,

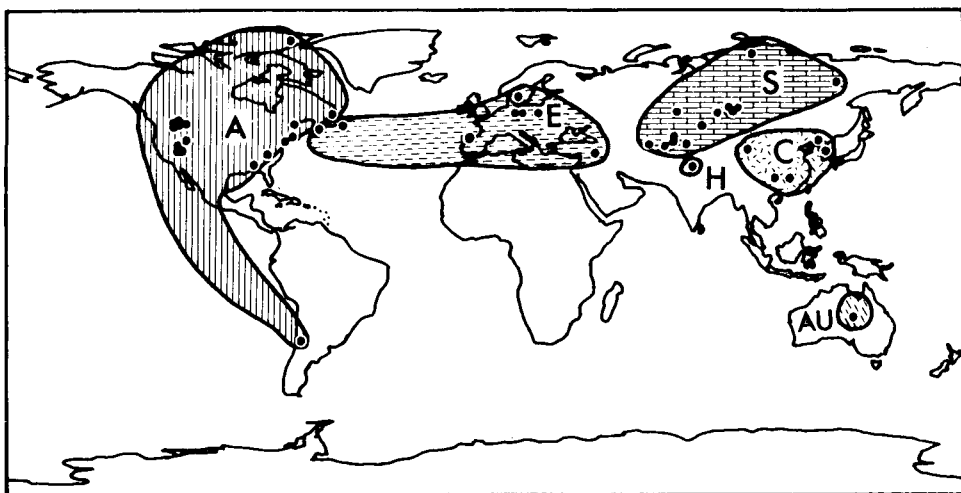


FIG 11.27 Trilobite biogeographic regions during the *Evinia-Elrathina* realms of the Middle Cambrian. After Burrett and Richardson (1980). A, American; E, European; S, Siberian; H, Himalayan; C, Chinese; AU, Australian.

China and Siberia. Minor faunal provincialism within individual blocks probably reflects climatic differences. Studies of Early Ordovician brachiopods also indicate they belonged to at least five distinct geographic provinces, which was reduced to three provinces during the Europe–North American collision in the Late Ordovician. The opening of the North Atlantic basin in the Cretaceous resulted in the development of American and Eurasian invertebrate groups from an originally homogeneous Tethyan group. Also similar ammonite populations from East Africa, Madagascar, and India indicate that only shallow seas existed between these areas during the Jurassic.

The similarity of mammals and reptiles in the Northern and Southern Hemispheres prior to 200 Ma ago demands land connections between the two hemispheres (Hallam, 1973). Early separation of Africa, India, and Australia (Early to Middle Mesozoic) led to the evolution of unique groups of mammals (e.g. the marsupials) in the Southern Hemisphere, while the fact that North America and Eurasia were not completely separated until Early Tertiary accounts for the overall similarity of Northern Hemisphere mammals today. When Africa, India, and Australia collided with Eurasia in the Mid-Tertiary, mammalian and reptilian orders spread both ways, and competition for the same ecological niches was keen. Such competition led to the extinction of 13 orders of mammals.

Plant distributions are also sensitive to continental drift. The most famous are the *Glossoptera* and *Gangemoptera* flora in the Southern Hemisphere. These groups range from Carboniferous to Triassic in age and occur on all continents in the Southern Hemisphere and in northeastern China, confirming that these continents (including at least South China) were connected during this period of time. The complex speciation of these groups could not have evolved independently on separate landmasses. The general coincidence of Late Carboniferous and Early Permian ice sheets and the *Glossopteris* flora appears to reflect adaption of *Glossopteris* to relatively temperate climates and its rapid spreading over high

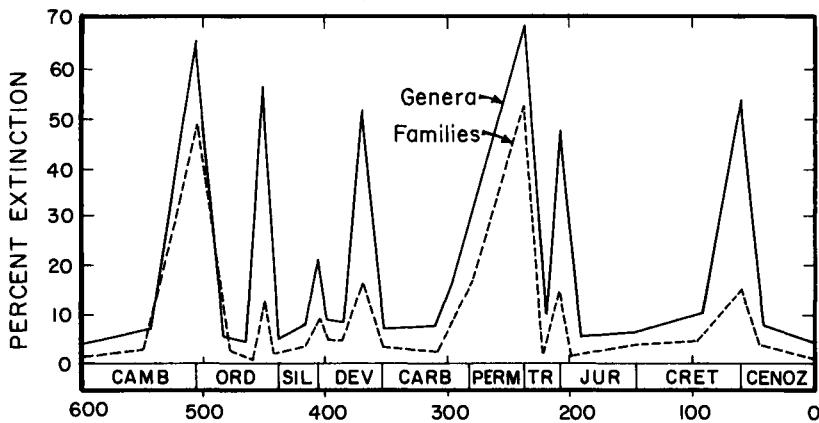


FIG 11.28 Major marine invertebrate extinctions during the Phanerozoic.

latitudes during the Permian. The break-up of Africa and South America is reflected by the present distributions of the rain forest tree *Symphonia globulifera* and a semi-arid leguminous herb *Teramnus uncinatus*, which occur at approximately the same latitudes on both sides of the South Atlantic (Melville, 1973). Conifer distribution in the Southern Hemisphere reflects continental break-up, as is evidenced by the evolution of specialized groups on the various continents after continent dispersal in the Early and Middle Mesozoic.

Although a correlation between the rate of taxonomic change of organisms and plate tectonics seem to be well established, the actual causes of such changes are less well agreed-upon. At least three different factors have been suggested to explain rapid increases in the diversity of organisms during the Phanerozoic (Hallam, 1973).

1. An increase in the areal distribution of particular environments increases the number of ecological niches in which organisms can become established. For instance, an increasing diversification of marine invertebrates in the upper Cretaceous may reflect increasing transgression of continents, resulting in an areal increase of the shallow-sea environment in which most marine invertebrates live. Such transgression appears to have resulted from progressive displacement of seawater onto the continents by active ocean-ridge systems.
2. Continental fragmentation leads to morphological divergence because of genetic isolation.
3. Some environments are more stable than others with regard to physical factors such as temperature, rainfall, and salinity. Studies of modern organisms indicate that stable environments lead to intensive partitioning of organisms into well-established niches and to corresponding high degrees of diversity. Environmental instability, decreases in environmental area, and competition of various groups of organisms for the same ecological niche can result in extinctions. Although continental collisions lead to these types of changes, other factors also are important.

Extinctions

Extinction and diversification of organisms have alternated with time, as if new organisms developed to fill vacated niches following major extinctions (fig. 11.28).

Important *mass extinctions*, which are extinctions of many diverse groups of organisms over short periods of time, are those at the end of the Permian, the end of the Cretaceous, and in upper Devonian (Frasnian). Other mass extinctions are recorded at the end of the Cambrian, Ordovician, and Triassic. Mass extinction episodes affect a great variety of organisms, marine and terrestrial, stationary and swimming forms, carnivores and herbivores, protozoans and metazoans. Hence the casual processes do not appear to be related to specific ecological, morphological, or taxonomic groups.

Proposed causes of extinction fall into three groups: extraterrestrial, physical, and biological. Extraterrestrial models include increased production of cosmic and X-radiation from nearby stars, increased radiation during reversals in the Earth's magnetic field, and climatic changes caused by supernova events or by impact on the Earth's surface. Of these, only two have been the subjects of detailed study. Correlation analysis between times of microfossil extinctions in deep-sea sediments and polarity reversals of the Earth's magnetic field does not support a relationship between the two (Plotnick, 1980). As discussed in the next section, a great deal of evidence seems to support impact on the Earth's surface as a cause for the extinctions at the Cretaceous–Tertiary boundary, and perhaps other extinctions.

Among the physical environmental changes that have been proposed to explain extinctions are rapid climatic changes, reduction in oceanic salinity caused by widespread evaporite deposition, fluctuations in atmospheric oxygen level, and changes in sea level. Collision of continents also leads to extinction of specialized groups of organisms. Rapid changes in environmental factors lead to widespread extinctions, while gradual changes permit organisms to adapt and may lead to diversification. It is clear that no single cause is responsible for all major extinctions, and it is probable that most extinctions have various contributing factors.

Extraterrestrial Impact Phenomena

Introduction

Perhaps no other topic has stimulated so much discussion and controversy in recent years as the impact model for explaining mass extinctions, and especially the extinctions at the Cretaceous–Tertiary (K–T) boundary (Alvarez et al., 1982). Evidence supporting the idea comes from Ir and Os anomalies in clay layers at the K–T boundary, the chemical and mineralogical composition of the boundary clays, spherules of possible impact-melt origin at many locations in boundary clays, and soot particles and shocked mineral grains in some boundary clays. Most of these features have been reported from widely separated worldwide locations. Asteroid or cometary impact is an appealing mechanism not only to account for these features, but to explain their worldwide distribution. Although voluminous, short-lived volcanic eruptions (such as occurred in the Deccan Traps at 60–65 Ma) also have been suggested as a mechanism for the K–T mass extinctions (Courillot et al., 1986), they do not seem capable of explaining all of the above features characteristic of the K–T boundary as well as the impact model. Although dating and correlation of the K–T boundary is not very precise, data suggest that mass extinctions took place gradually over a few million years, and involved various marine and terrestrial organisms. Some fossils, such as dinosaurs that became extinct at the K–T

boundary, are so rare that it is not known if their extinction was abrupt or gradual. Some invertebrates (such as ammonites and rudists) dropped significantly in numbers 1–2 Ma before the end of the Cretaceous whereas others (bivalves, brachiopods, foraminifera) became extinct precisely at the K–T boundary.

Data suggest that mass extinctions are periodic with approximately a 26 Ma extinction period (Hut et al., 1987). Periodic phenomena that may be responsible for these extinctions are cometary showers and asteroid impacts, both of which have periods of about 30 Ma. Comets reside in the Oort cloud in the outer part of the solar system ($\geq 2 \times 10^4$ AU), and this cloud may undergo periodic disturbances which send cometary showers into the inner solar system. Possible causes of Oort cloud disturbance include interactions with a companion star of the Sun or oscillations of the solar system in the galactic plane. Periodic asteroid impacts have also been suggested as a cause of mass extinction periodicity.

Extinction Mechanisms

It is important to distinguish between the ultimate cause of mass extinctions, such as impact, and the immediate cause(s), i.e. rapid changes in environment that kill plants and animals in large numbers worldwide. Impact on the Earth's surface of a cometary shower or an asteroid (only 10 km in diameter) may give rise to many changes in the environment that result in extinctions of various groups of organisms. The major consequences of collision of 10-km-diameter asteroid on the Earth's surface are summarized as follows.

1. *Darkness.* Fine dust particles would spread worldwide in the upper atmosphere, completely cutting out all sunlight for a few months after the impact (Toon et al., 1982). This would suppress photosynthesis and initiate a collapse of food chains, causing death to many groups of organisms by starvation. For instance major groups that became extinct at the end of the Cretaceous, such as dinosaurs, marine planktonic and nektonic organisms and benthic filter feeders were in food chains tied directly with living plants. Organisms less affected by extinction, including marine benthic scavengers and deposit feeders and small insectivorous mammals, were in food chains dependent upon dead plant material.
2. *Cold.* The dust-produced darkness would be accompanied by extreme cold, especially in continental interiors that are far from the moderating influence of oceans. Within 2 or 3 months continental surface temperatures would fall to -20°C (fig. 11.29).
3. *Increased greenhouse effect.* If the asteroid collided in the ocean, which seems likely in that geological evidence of a terrestrial impact has not been found, both dust and water vapor would be spread into the atmosphere. Calculations indicate that, after the dust settles, water vapor will remain in the upper atmosphere producing an enhanced greenhouse effect that would raise surface temperatures well in excess of the tolerance limits of many terrestrial organisms.
4. *Acid rain.* Energy liberated during the impact may cause N_2 , H_2O , and O_2 in the atmosphere to react producing HNO_3 and various nitrogen oxides.

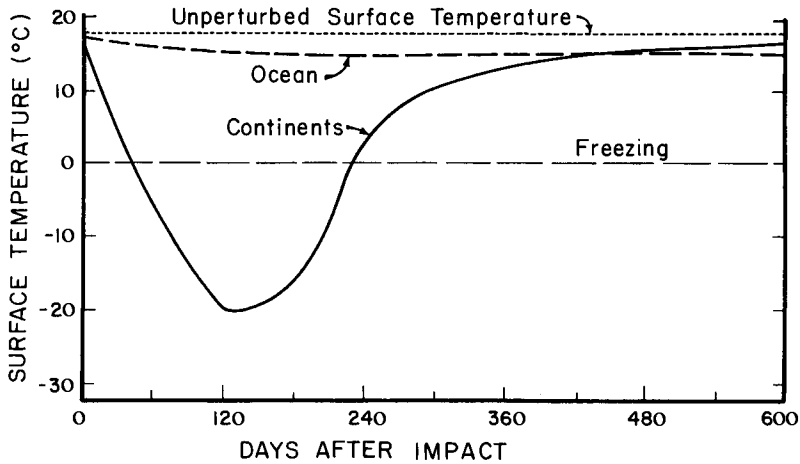


FIG 11.29 Change in terrestrial surface temperature of the oceans and continents following impact with a body about 10 km in diameter. Atmospheric dust density following impact is 1 gm/cm^2 . Modified after Toon (1984).

Hence, another side-effect of the impact bearing on extinctions is the possibility of a nitric acid rain.

5. *Wildfires.* The soot particles reported in clays at the K-T boundary may be the result of widespread wildfires ignited by infrared radiation from the initial impact.
6. *Toxic seawater.* Asteroid collision should also result in the introduction of a wide variety of trace elements, many of which are toxic, into the oceans (such as Zn, Ni, Ge, Co, Cd, Pb, Hg, Se). Organisms living in surface marine waters would be exposed to prolonged concentrations of toxic elements, thus contributing to the widespread extinctions of marine organisms.

Earth-crossing Asteroids

Earth-crossing asteroids are asteroids that are capable of colliding with the Earth with only small perturbations of their orbits, which presently cross the Earth's orbit. Approximately 50 Earth-crossing asteroids (ECA) are known, of which 40 have diameters $\geq 1 \text{ km}$. Estimates of the total population of ECA are in the range of 1000–2300 (Shoemaker, 1984). Those with diameters of $\geq 10 \text{ km}$ probably do not exceed eight in number. Probabilities of collisions with the Earth are $\sim 6 \times 10^{-6}/\text{a}$ for asteroids $\geq 1 \text{ km}$ in diameter and $2 \times 10^{-8}/\text{a}$ for those with diameters $\geq 10 \text{ km}$. This amounts to about 10 asteroids with diameters $\geq 10 \text{ km}$ colliding with the Earth since the end of the Precambrian.

Comet Showers

About three long-period and 10 short-period ($< 20 \text{ a}$) comets pass inside the Earth's orbit each year. Because long-period comets have velocities greater than asteroids, collisions with the Earth should liberate approximately an order of magnitude more

energy (Hut et al., 1987). There is very little evidence that allows a distinction to be made between a comet and an asteroid impact at the K–T boundary. Perturbations of the Oort cloud will send cometary showers into the inner solar system. Monte Carlo calculations indicate that Earth-crossing comets make about nine returns to the inner solar system before being ejected into a different orbit. This corresponds to a mean lifetime of about 0.5 Ma. Estimates suggest that about 50% of the preserved cratering of the Earth is due to cometary showers rather than asteroids or meteors. A major shower involving $\sim 10^9$ comets of >3 km in diameter would result in about 20 impacts on the Earth's surface and should occur every 300–500 Ma. Smaller showers ($\sim 10^8$ comets) involving about two impacts should occur every 30–50 Ma.

The probable ages of known impact structures (and impact glasses) on the Earth's surface <100 Ma in age suggest an episodicity of impacting with peaks at 0, 35, 65, and 99 Ma (Hut et al., 1987). Recognized mass extinctions that annihilate 50–95% of ecologically diverse lower taxa within 1–3.5 Ma are recognized at 93, 66 (K–T), and 36 Ma ago. These extinctions, which correlate well with the three impact peaks, are graded in that each extinction progresses from the least tolerant to the most-tolerant ecological groups. The duration and periodicity of the extinctions also match predicted duration and periodicity of cometary showers, and favor the idea that cometary showers are not only responsible for most of the terrestrial impacts in the last 100 Ma, but also are responsible for at least three mass extinctions, including the K–T extinctions.

Summary Statements

1. Depletion in rare gases in the Earth indicates that if a primitive atmosphere existed it was lost during the late stages of accretion. Such an early atmosphere would have been composed chiefly of CO_2 , CO , H_2O , and H_2 .
2. The terrestrial atmosphere has grown by degassing of the Earth. The most important constituents in the Early Archean atmosphere were probably CO_2 and H_2O . Isotopic studies of rare gases (especially $^{40}\text{Ar}/^{39}\text{Ar}$ and $^{129}\text{Xe}/^{130}\text{Xe}$) in ocean-ridge basalts suggest, but do not prove, extensive degassing of the Earth within 50 Ma after accretion followed by more gradual degassing to the present time.
3. In order to avoid freezing surface temperatures on Earth prior to 2 Ga ago caused by low solar luminosity, an atmosphere rich in CO_2 is likely in order to promote a greenhouse warming effect.
4. The CO_2 level of the terrestrial atmosphere has decreased with time in response to some combination of changes in the carbonate–silicate cycle, growth of the ozone layer, and increases in the numbers of photosynthetic organisms.
5. The continuously habitable zone for life in the solar system probably extends from about 0.95 to 1.5 AU.
6. Oxygen in the atmosphere is produced by photosynthesis and reduction of marine sulfates and carbonates, and it is lost by weathering, respiration, and decay. Prior to the appearance of photosynthetic organisms, minor amounts of O_2 were produced by H_2O photolysis in the upper atmosphere.
7. The increase in abundance of banded iron formation in the Late Archean and

Early Proterozoic probably reflects increasing numbers of oxygen-producing microbes. The drop in abundance of banded iron formation after 1.7 Ga reflects exhaustion of ferrous iron in seawater.

8. The growth of O_2 in the atmosphere progressed in three stages: I, free O_2 occurs in neither the oceans nor atmosphere; II, small amounts of O_2 occur in the atmosphere and surface ocean water; and III, O_2 invades the entire atmosphere-ocean system. Stage II began at approximately 2.4 Ga and is marked by the appearance of redbeds, an increase in the abundance of sulfates, and near-disappearance of detrital uraninite-pyrite deposits. Stage III began about 1.8–2.0 Ga and is characterized by the near-disappearance of banded iron formation and appearance of eukaryotic organisms.
9. Overall differences in the atmospheres of Earth, Venus, and Mars reflect distance from the Sun (controlling initial surface temperature), planetary mass, and the role of a greenhouse effect. Light gases (N_2 , H_2) were lost from Mars and rapid planetary cooling resulted in trapping most of its CO_2 in the interior. Venus may have had early oceans that evaporated due to a moist greenhouse effect. H_2O moved into the upper venusian atmosphere where photolysis produced H_2 (which escaped) and O_2 (which probably oxidized surface rocks) leaving a dense CO_2 atmosphere.
10. The growth of the Earth's oceans probably paralleled atmospheric growth and reached $\geq 90\%$ of present ocean volume by the end of the Archean. On the whole, changes in seawater composition probably have been small, and reflect variations in the CO_2/O_2 ratio of the atmosphere, changes in the relative roles of continental sediment input into seawater, recycling at ocean ridges, short-lived periods of widespread evaporite deposition, and variations in abundances of land plants.
11. Increasing $\delta^{18}O$ values in marine cherts with time reflect either cooling of the oceans or, more likely, interactions of seawater with continent-derived sediments and seawater recycling at ocean ridges.
12. Increases in $\delta^{13}C_{carb}$ in the Carboniferous and Permian probably reflect an increase in land plants. The $\delta^{13}C_{org}$ data scatter with geologic time reflecting primary variations in $\delta^{13}C_{org}$ of original organisms, metamorphism, and the effects of methane-using bacteria which may account for extreme negative $\delta^{13}C_{org}$ in some Early Precambrian kerogens.
13. The $\delta^{34}S$ values of marine sulfates appear to increase with time from 3.8 Ga while $\delta^{34}S$ values of marine sulfides decrease. These changes may be caused by an increase in numbers of sulfate-reducing bacteria or by cooling of the oceans with time.
14. Increases in the Ca/Mg and calcite/dolomite ratios in marine carbonates with time may reflect one or some combination of changes in the Mg^{+2} depository in seawater, a decrease in atmospheric CO_2 level, or losses of Mg due to recycling at ocean ridges.
15. The most important factors controlling changes in sea level are changes in the volume of the ocean-ridge system (long-term changes) and glaciation (short-term changes). Sea level highs during the Early Paleozoic and Late Cretaceous appear to reflect the fragmentation of supercontinents in the Late Proterozoic and Triassic, respectively.

16. Important paleoclimatic indicators include glacial deposits, coal, aeolian sandstones, evaporites, redbeds, laterites and bauxites, various floral and faunal populations, and $\delta^{18}\text{O}$ values of marine carbonates and cherts. The Earth's surface temperature is related to solar luminosity, albedo, and the greenhouse effect, and the latter two factors are partly controlled by plate tectonics and volcanism.
17. Six major periods of continental glaciation are recognized at 2800, 2300–2400, 600–900, 450, 250–350 and 0–15 Ma. Important causes of glaciation include decreases in solar heating due to volcanic eruptions, decreased CO_2 degassing of the Earth, changes in continent–ocean geometry caused by plate tectonics, and changes in the Earth's orbital characteristics. The Precambrian, Early Paleozoic, and Late Mesozoic are times of widespread warm climates.
18. Life appears to have formed on Earth by about 4 Ga ago. Steps in the formation of a living cell include reactions of gaseous components to produce amino acids, formation of polypeptides including RNA and DNA, and the appearance of membranes. Suitable Early Archean sites for a supply of organic compounds, an energy source, and mineral catalysts include volcanic centers and submarine hydrothermal vent systems. RNA may have played a key role in the development of self-replication.
19. The first microbes were probably anaerobic, fermenting heterotrophs. Autotrophs appeared by 3.5 Ga. Aerobic photoautotrophs and sulfate-reducing bacteria appeared by 2.8 Ga and eukaryotes by about 1.5 Ga. Metazoans appeared at about 700 Ma ago. Major Phanerozoic biologic events include the appearance of calcareous and siliceous skeletons (500 Ma), the origin of vertebrates (450 Ma), vascular plants (425 Ma), mammals (225 Ma), flowering plants (150 Ma), and man (3 Ma).
20. Opening of ocean basins leads to diversification and specialization of organisms, while collision of continents leads to competition for survival and extinctions. Other possible causes of mass extinctions include asteroid or comet impacts, rapid climatic changes, reduction of seawater salinity, changes in atmospheric O_2 level, and changes in sea level.
21. Processes leading to mass extinctions following cometary or asteroidal impact include darkness, cold temperatures, increased greenhouse effect, acid rain, wildfires, and toxic seawater.
22. Mass extinctions at the Cretaceous–Tertiary boundary and at 93 and 35 Ma ago correlate with, and have similar durations and periodicities to, cometary showers on the Earth, and are consistent with a cometary impact model for these extinctions.

Suggestions for Further Reading

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Study Questions

CHAPTER 2

1. The rate of accretion of a planet in a gaseous nebula can be approximated by,

$$dM/dt = \pi r^2 v \rho_c \quad (2a)$$

where M and t refer to the mass of the planet and time, respectively. r is the radius, v the velocity of the planet in the nebula, and ρ_c is the density of particles in nebula in the vicinity of the accreting planet.

- a) show that the time needed for a planet to grow to a radius r is,

$$t = ([4/3]r \rho_p)/v \rho_c \quad (2b)$$

where ρ_p is the density of the planet. Assume that v and ρ_c remain constant during growth.

b) Assuming a density of the solar nebula of 10^{-10} gm/cm^3 , how long would it take to accrete an asteroid 1000km in diameter? How long for the Earth with a radius of 6378km? Use an asteroid density of 3.3 gm/cm^3 and the Earth's present density of 5.52 gm/cm^3 . Assume average velocities of 5 km/sec for the asteroid and 15 km/sec for the Earth for the first 50 My of accretion.

- c) Make a plot of t versus r for the first 50 My after accretion began.

d) Some models suggest the Earth reached its present size in about 50 My. Discuss the significance of your calculations and plot to this model including possible reasons for any discrepancies.

2. a) Assuming the Moon was captured by the Earth, calculate the age of a prograde orbital capture at the Roche limit ($2.9 R_E$) for a constant lunar retreat rate of 5 cm/y. Discuss the significance of your results in terms of the age of the Earth (4.6 Ga). If the rotational period of the Moon matched that of the Earth (*i.e.*, a synchronous orbit) soon after capture, calculate this period (expressed in days) from

the following relationship (in cgs units),

$$a/E_R = 1.93 \times 10^{-3} (P^2 \rho)^{1/3} \quad (2c)$$

where a is the orbit radius (3.84×10^5 km), E_R is the Earth's radius (6378 km), P is the period of rotation, and ρ is the Earth's density (5.5 gm/cm³).

b) Make a plot showing the relationship of angular velocity of the Earth versus days/year, assuming conservation of angular momentum and a decreasing rate of angular velocity of the Moon of 5.3×10^{-22} rad/sec. The current angular velocity of Earth is 7.29×10^{-5} rad/sec.

3. Although studies of asteroids show they have irregular shapes, as size increases their shape tends to approach that of a sphere. As long as the strength of planetary rock exceeds the pressure at a given depth, an asteroid retains irregular shape, which may have been acquired by collision. As the interior pressure approaches and exceeds the rock strength (S), the central part of the asteroid is crushed and the surface begins to approach a spherical shape as given by the following relationship,

$$S = 1.34 \times 10^{-10} (\rho^2) (r_{\max}^2) \quad (\text{SI units}) \quad (2d)$$

where ρ is the average planetary density and $r_{\max}$ is the maximum radius that will sustain an irregular shape.

Chondrites have an average density of 3.5 gm/cm³ and strength of 6×10^8 N/m², whereas iron meteorites have corresponding values of 7.9 gm/cm³ and 4×10^8 N/m². Calculate $r_{\max}$ (in km) for a chondritic asteroid and for an asteroid with the ratio of chondrite/iron equal to the Earth's mantle/core ratio of 68/32. Compare your results to the observed sizes and shapes of asteroids and to rocky satellites around the outer planets.

CHAPTER 3

1. Seismic wave velocity data for the core indicate that the outer core must contain 5–15% of low-atomic number elements. If sulfur is the only low-atomic number element, 8% is required to explain the density of the outer core. Experimental data in the system Fe-FeS indicate that the eutectic composition in this system probably satisfies the following equation,

$$(1-x)^{1-\beta} (4x)^{-\beta} = 2^{(1-\beta)} \quad (3a)$$

where x is the fraction of FeS and $\beta = T_{\text{Fe}}/T_{\text{FeS}}$ the ratio of the melting points of pure Fe and pure FeS. Extrapolation of thermodynamic data to high pressures suggests that $\beta = 1$ at the outer core boundary at 5200 km.

a) Calculate the eutectic composition at 5200 km and compare the weight percent sulfur to that required by seismic data (8%). Discuss the significance of the difference. Show this composition on a simplified binary phase diagram for Fe-FeS, assuming a temperature at the inner core boundary of 6300°C and a eutectic temperature at this depth of 5800°C (Fig. 3.18).

b) As cooling at the inner core boundary progresses from 6300°C to 6000°C, what will happen? Show this diagrammatically on your binary phase diagram. Interpolate the melting point of pure Fe at the outer core boundary from relations in Fig. 3.18.

2. An important quantity in modelling mantle convection is T_M , the mean time would take the entire mantle to be “cycled” through ocean ridges, which is related to M_M , the mass of the mantle (4.1×10^{24} kg) as follows,

$$T_M = M_M/f, \quad (3b)$$

where f is rate of oceanic lithosphere formation at ridges. f can be further expressed as,

$$f = D \rho (da/dt), \quad (3c)$$

where D is the thickness and ρ is the density of the lithosphere produced at ocean ridges, and da/dt is the aerial rate of plate creation at ridges.

a) Calculate T_M based on present-day parameters ($da/dt = 3 \text{ km}^2/\text{y}$; $\rho = 3.4 \text{ gm/cm}^3$; and $D = 50 \text{ km}$) and compare your result to the mean age of the mantle of 2 Ga estimated from Nd and Pb isotopic data from basalts. Discuss the significance of your T_M value.

b) Construct a plot of T_M versus D . How thick is the lithosphere at ocean ridges if all of the mantle had been cycled through ocean ridges in the last 2 Ga? How does this thickness compare with the average thickness of oceanic lithosphere allowed by seismic and experimental petrologic data of about 50 km?

c) The mean age of the mantle τ can be estimated from the following expression for whole-mantle convection,

$$\tau = T_M(1 - e^{-t/T_M}) \quad (3d)$$

where t is any particular time in the geologic past. Calculate τ for the case when $t = t_E$, the age of the Earth (4.6 Ga) and compare your result to the 2 Ga value determined from Nd and Pb isotope studies. It is probable that the difference in the values is due to an error in estimating T_M . What factor in equation 3b or 3c may have been different in the Archean to account for this discrepancy? Calculate the mass of the mantle involved in convection to reduce T_M to 2 Ga, assuming an average mantle density of 4.4 gm/cm^3 . If only the upper mantle convects, calculate how deep the convection cells should be for this mass of the mantle. The volume of the mantle is $9.06 \times 10^{20} \text{ m}^3$ and the Earth's radius is 6378 km.

3. The pressure P at any depth d in a planet can be estimated from the following expression,

$$P(\text{pascals}) = [2\pi G \rho^2 / 3] \times (R^2 - r^2) \quad (\text{SI units}) \quad (3e)$$

where G is the gravitational constant ($6.673 \times 10^{-11} \text{ Nm}^2\text{kg}^{-2}$), ρ the mean planetary density, and R the planetary radius, where $r = R - d$.]

a) Determine if the spinel $\rightarrow$ perovskite + periclase reaction will occur in the Moon, Mars and Mercury, and if so, at what depths. Assume it occurs in the Earth at 670km depth. Assume the following planetary quantities:

	Mean Density (gm/cm ³)	Radius (km)
Earth	5.52	6378
Mars	3.94	3398
Mercury	5.42	2439
Moon	3.34	1738

b) Generate a plot of depth versus pressure to 500km depth in Mars, Earth, Mercury, and the Moon. Discuss the results in terms of possible seismic discontinuities in the upper mantle of each planet.

CHAPTER 4

1. Seismic and xenolith data suggest that the subcontinental lithosphere (SCL) is thicker beneath Archean cratons than beneath younger continental crust. Evaluate and discuss this possibility using heat flow and heat generation data given in Table 4.6 assuming upper and lower continental crustal layers that are 10 and 30km thick, respectively.

a) For a surface heat flow (q_s) of 50 mW/m^2 and a SCL 110km thick (crust + SCL = 150km) typical of the post-Archean continental crust, calculate the heat flow coming from the sublithospheric mantle + core. Assume the step model (Fig. 4.14A) for all three layers. What fraction of the total heat flow comes from the upper crust, lower crust, SCL, and from greater depths?

b) Construct a plot of h , the thickness of the SCL, versus surface heat flow for zero heat contribution from the lower crust, and for the heat production calculated using the radiogenic element contents given in Table 4.6. What surface heat flow would be expected from an Archean craton with a SCL 300km thick and the same contents of radiogenic isotopes? Is this value similar to those observed for Archean cratons (see Table 4.2) and if not, offer one or more possible reasons why. How much heat comes from below the lithosphere? Discuss the significance of your calculations.

TABLE 4.6 *Data for Heat Flow Problem*

	U(ppm)	Th(ppm)	K(%)	ρ (gm/cm ³)
Upper crust	2.5	10	2.7	2.7
Lower crust	0.30	1.8	0.50	2.9
SCL	0.1	0.5	0.04	3.4

Use the following heat production relations:

$$A(\mu\text{Wm}^{-3}) = \rho[0.097 \text{ U(ppm)} + 0.036 \text{ K(\%)} + 0.026 \text{ Th(ppm)}].$$

where ρ = the average density, and U, Th, and K are element concentrations.

2. Ion-probe U/Pb isotopic data from hand-picked zircons from a 3.4-Ga-old metasediment from South Africa are given in Table 4.7. Microscopic study of the zircons indicates three populations: A) dark, subrounded zircons with cores; B) pink, subrounded zircons without cores; and C) small, clear euhedral zircons. On a graph of concordia, plot the zircon data using different symbols for each population. If you can fit a chord to the data, calculate and show graphically the discordia ages. Discuss possible interpretations of the data.

TABLE 4.7 *Zircon U-Pb Isotopic Data*

Population	²⁰⁷ Pb/ ²³⁵ U	²⁰⁶ Pb/ ²³⁸ U
A	30.44	0.712
	26.74	0.647
	30.30	0.721
	30.25	0.707
	17.5	0.450
	15.60	0.350
	20.05	0.475
B	22.30	0.582
	24.51	0.621
	20.1	0.53
	19.2	0.52
C	7.54	0.401
	7.49	0.408
	8.16	0.419
	7.41	0.402
	7.15	0.395

Concordia can be calculated from the following relations:

$$^{207}\text{Pb}/^{235}\text{U} = e^{(\lambda_1 t)} - 1$$

$$^{206}\text{Pb}/^{238}\text{U} = e^{(\lambda_2 t)} - 1$$

where the decay constants are $\lambda_1 = 9.85 \times 10^{-10} \text{ y}^{-1}$ and $\lambda_2 = 1.55 \times 10^{-10} \text{ y}^{-1}$.

3. a) Using the results in Table 4.8 from prograde metamorphic minerals, calculate both the average rate of burial and uplift during a continent-continent collision.

TABLE 4.8 *P-T Data*

P-T	Age
4 kbar, 450°C	362 ± 8 Ma Rb/Sr (biotite, K-feldspar, plagioclase isochron)
6 kbar, 600°C	340 ± 10 Ma ⁴⁰ Ar/ ³⁹ Ar hornblende
2 kbar, 400°C	302 ± 5 Ma fission tracks muscovite

b) Construct a P-T-t plot of the burial-uplift trajectory with depth (in km). Label time (t) on the P-T curve. Assume 3.4 km/kb.

c) Discuss the significance of your results.

CHAPTER 5

1. Because of their buoyancy, continents tend to resist subduction. The pull of a descending lithospheric slab can be expressed as a force per unit length F_L ,

$$F_L = [(\rho_a - \rho_c)(h_c - h_s)]dg \quad (5a)$$

where ρ_c and ρ_a are the densities of the crust and asthenosphere, respectively, g the gravitational acceleration at the Earth's surface, h_c the crustal thickness, and d the length of continental crust subducted. The subcontinental lithosphere (SCL) contributes to subduction by pulling down a fraction of the lower continental crust h_s . Subduction of continental lithosphere should stop when the force pulling it down (F_L) is balanced by the buoyancy of the continental crust being subducted.

Estimates of the variables in equation 5a are as follows:

$$\begin{aligned} \rho_a &= 3.4 \text{ gm/cm}^3 \\ \rho_c &= 2.8 \text{ gm/cm}^3 \\ h_c &= 40 \text{ km} \\ h_s &= 10 \text{ km} \end{aligned}$$

h_s is estimated from density and thermal relationships between the crust and SCL. The pull of the descending slab depends on various assumptions, four models of which are as follows:

- 1) $F_L = 1.8 \times 10^{16}$ dyne/cm: slab pull to a depth of 350 km
- 2) $F_L = 3.2 \times 10^{16}$ dyne/cm: slab pull to a depth of 700 km

- 3) $F_L = 5.1 \times 10^{16}$ dyne/cm: slab pull to a depth of 700 km
with excess mass caused by the olivine-spinel phase change
- 4) $F_L = 7 \times 10^{16}$ dyne/cm: case three with pull to a depth of 1500 km
and excess mass added at the 670 km discontinuity

a) Calculate d for each of the four model forces above.

b) One of the uncertainties in this calculation is h_s . Construct a plot of h_s versus d for each model (show all four on the same plot). Assuming an upper limit for h_s of 20 km, label the maximum d values for each model.

c) Seismic data indicate that it is very unlikely that continental crust is subducted more than 50-100 km, and only then during a continent-continent collision. List and discuss one or more assumptions made in models 3 and 4 that could be changed to prevent the deep subduction of continental crust.

2. It has been suggested that hotspot density at the Earth's surface can be related to plate speed (v in cm/y) by,

$$\rho_h = 141[e^{-(v/4.8)}] \quad (5b)$$

where ρ_h is the relative hotspot density (number of hotspots/unit area).

a) Using the map in the pocket of the book, the hotspot distribution in Fig. 5.21, and the plate speeds given in Fig. 3.17, compare the observed hotspot densities with those calculated by equation 5b for the Pacific, Indian, North American, and African plates.

b) Discuss and suggest reasons for any differences between observed and calculated hotspot densities.

CHAPTER 6

1. Empirical data from modern oceanic plates indicate that C_o , the rate of generation of new oceanic lithosphere can be expressed as

$$C_o = 2A/t_m \quad (6a)$$

where A is the area of the ocean floor and t_m , the maximum age of the ocean floor (180 Ma today). C_o can also be expressed as the product of total ridge length L and average spreading rate V_s ,

$$C_o = LV_s \quad (6b)$$

a) For the present ridge system, $L = 50,000 \text{ km}$, calculate C_o (km^2/y) and V_s (cm/y).

b) In the Archean, C_o may have been considerably greater due to more vigorous convection and greater heat production in the Earth. Assuming spreading rate V_s is directly proportional to heat production, estimate C_o at 3.5 Ga with a heat production rate 3 times the present rate. How long would the ridge system be if this increased lithosphere production rate were due totally to an increase in L ? What is V_s if it were due totally to faster spreading rates? Construct a plot of C_o (x axis) versus L and V_s (one on each y axis).

c) What would t_m be if there was a negligible amount of continental crust at this time?

d) Discuss the significance of these calculations and the plot in terms of plate tectonics in the Archean.

2. Experimental and theoretical data suggest that t , the time required for a mantle plume to rise from its source to the base of the lithosphere can be estimated from,

$$t^7 = (d/0.16)^5 (\eta/g\Delta\rho)^4 (1/R)^3 \quad (\text{cgs units}) \quad (6c)$$

where d is depth, $\Delta\rho$ the density contrast of the plume and surrounding mantle, η the average viscosity of the mantle, g the acceleration of gravity, and R the rate at which the plume is being fed from its source.

a) Calculate the rise time of a plume generated in the D'' layer at a depth of 2900 km, with $\Delta\rho = 0.1 \text{ gm/cm}^3$ and $\eta = 10^{22}$ poise, the average mantle viscosity. Assume an R value equal to the production rate of basalts at the Reunion hotspot of $0.4 \text{ km}^3/\text{y}$. Make similar calculations for the following hotspot-related flood basalts with their corresponding eruption rates: Deccan traps ($2 \text{ km}^3/\text{y}$), Iceland ($1.5 \text{ km}^3/\text{y}$), Parana basalts ($0.75 \text{ km}^3/\text{y}$), and Karoo basalts ($1 \text{ km}^3/\text{y}$).

b) Make a graph of rise time versus density contrast and use a paint program to place a pattern on the region of reasonable ranges in $\Delta\rho$ in the mantle ($0.01\text{--}0.2 \text{ gm/cm}^3$). Discuss the significance of the graph and of the plume rise times in terms of the estimated rates of mantle turnover by convection ($\sim 10^9$ years).

CHAPTER 7

1. A young alkali basalt in the Andes has a Rb content of 100 ppm and a Sr content of 500 ppm. Employing the data given below, calculate the Rb/Sr ratio of a lherzolite parent assuming 10% partial melting. How does the calculated Rb/Sr ratio

compare to the probable ratio of unfractionated mantle (0.03) and discuss the results?

	Mode Fraction (m)	Melt Fraction (p)	K_d^{Rb}	K_d^{Sr}
olivine	0.60	0.25	0.01	0.014
opx	0.25	0.20	0.02	0.02
cpx	0.15	0.55	0.33	0.01

Note that for batch melting,

$$C^L/C_o = 1/[D_o + F(1-P)] \quad (7a)$$

where C^L = concentration of an element in the melt

C_o = concentration of an element in the source

D_o = Bulk distribution coefficient which is equal to the weighted sum of mineral/liquid distribution coefficients (K_d), such that,

$$D_o = m^1 K_d^1 + m^2 K_d^2 + \dots m^i K_d^i \quad (7b)$$

m^i values are mass fractions of each mineral i .

F = fraction of melting

P = Bulk distribution coefficient of melting where,

$$P = p^1 K_d^1 + p^2 K_d^2 + \dots p^i K_d^i \quad (7c)$$

p^i values are the fractions of phases entering the melt.

2. Using the mantle Rb/Sr ratio calculated in problem 1 and assuming that this mantle source fractionated from undepleted mantle, calculate the age of this fractionation and the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the alkali basalt, employing the following data:

$$^{87}\text{Sr}/^{86}\text{Sr} = 2.9 \text{ Rb/Sr}$$

$$^{87}\text{Sr}/^{86}\text{Sr} \text{ of the alkali basalt} = 0.7070$$

$$\text{age of the Earth} = 4.6 \text{ Ga}$$

Plot the relationships on a diagram showing $^{87}\text{Sr}/^{86}\text{Sr}$ on the vertical axis and time (in Ga) on the horizontal axis and discuss the results.

CHAPTER 8

1. Using *Terra Mobilis* (© C. R. Denham and C. R. Scotese, Geoimages, 700 Tanglewood, Arlington, TX 76012) or another plate tectonics program, plot the paths of India and Australia from 200 Ma to the present.

a) From this plot, calculate the rates of plate motion for each continent averaged over 20 My intervals and compare these to average modern plate velocities given in Fig. 3.17.

b) Construct an x-y plot of the time integrated velocities (20 My windows) of both continents and discuss the results in terms of their plate tectonic histories.

2. a) Again using *Terra Mobilis*, calculate the rate of growth of Asia from the Pennsylvanian (300 Ma) to the present. Calculate the rates every 25 My and construct a plot of age versus rate.

b) Discuss your results and indicate the major uncertainties that make your rate estimates either maximum or minimum values.

1. Using *Terra Mobilis* (© C. R. Denham and C. R. Scotese, Geoimages, 700 Tanglewood, Arlington, TX 76012) or another plate tectonics program, plot the paths of India and Australia from 200 Ma to the present.

CHAPTER 9

1. Compare the rates of plate motion during the Proterozoic (2.3–1.7 Ga) with those of the Phanerozoic for North America using the data in Figures 5.20 and 9.26. To do this, estimate rates averaged for 20 My increments. Interpolate ages on the APW paths in Fig. 9.26.

a) Plot the average plate velocity for North America for 20 My increments versus time from the present back to 300 Ma and from 2300 to 1000 Ma.

b) Discuss the distribution of data on your plot in terms of the plate tectonic history of North America. How might the results bear on supercontinental cycles?

2. Sediment associations on the cratons suggest that the average elevation of continents above sea level (*i.e.*, the freeboard) has remained approximately constant since the Late Archean. The mean depth of ocean ridges (*d*) is related to continental freeboard (*e*) by the following expression,

$$d = [\rho_c h_c + \rho_a h_L [k-1] - \rho_a h_c + \rho_a \rho] / (1 - \rho_a) \quad (9a)$$

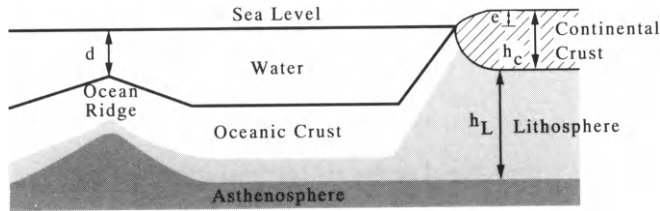
where the *h* terms are thicknesses and ρ terms densities as defined in the accompanying table and drawing. *K* is defined as follows,

$$K = 1 + \alpha(T_a - T_L) \quad (9b)$$

where α is the coefficient of expansion of the lithosphere (approx. $40 \times 10^6/^\circ\text{C}$), and T terms are the average temperatures (deg. C) of asthenosphere and lithosphere, respectively.

a) From this expression, calculate the average thickness of continental lithosphere today ($h_c + h_L$) and compare your results to estimates based on seismic wave data (150–200 km).

b) Assuming the Late Archean lithosphere is only 50 km thick, calculate the average depth of ocean ridges (d) to maintain the same continental freeboard as we have today. Make a plot of d versus time assuming a direct correlation of curve shape to the cooling curve of the Earth in Figure 2.9.



c) Discuss the assumptions made in your calculations and the significance of the results.

TABLE 9.3 Data for Problem 9.2

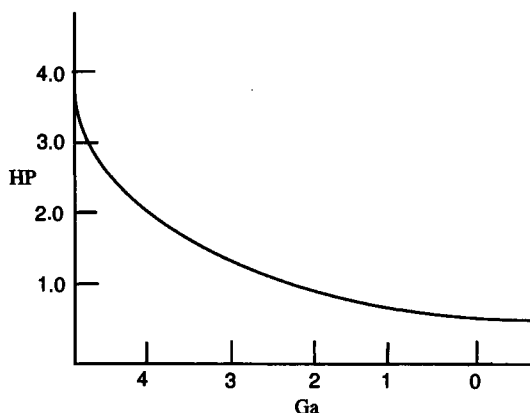
	Late Archean	Present
T_a Asthenosphere temperature	1500°C	1300°C
T_L Lithosphere temperature	800°C	800°C
h_c Continent thickness	35 km	35 km
h_L Subcont lithosphere thickness	50 km	
e Freeboard	300 m	300 m
d Oceanic depth at ocean ridge		1 km
ρ_L Density lithosphere	3.4 gm/cm ³	3.4 gm/cm ³
ρ_a Density Asthenosphere	3.6 gm/cm ³	3.7 gm/cm ³
ρ_c Density Continental Crust	2.8 gm/cm ³	2.8 gm/cm ³

CHAPTER 10

1. Estimates of the growth rate of continental crust (G) during the Phanerozoic generally fall in the range of 0.8 to 1.2 km³/y.

a) Assuming that this growth rate varied as the relative heat production rate in the Earth (HP) in the figure below, plot G as a function of time, assuming 1 km³/y

growth rate at zero time. What are the predicted growth rates of continents (in km^3/y) during the Archean and how do they compare with rates portrayed in Fig. 10.6? Discuss the differences.



b) Estimated volumes and crustal formation ages of various segments of continental crust are given in the table below. Calculate the rates of continental growth for each and plot these points on your graphs in part a. Plot HP on one y axis and G on the other. Discuss the results.

	vol ($\text{km}^3 \times 10^8$)	Age (Ga)
Canadian Shield	4.0	3.0 – 2.7
Baltic Shield	2.5	2.0 – 1.85
Central & SW USA	3.0	1.85 – 1.65
Arabian-Nubian Shield	3.0	0.9 – 0.6
Eastern Australia	1.5	0.4 – 0.2
Cordilleran accretionary terranes, NW North America	1.9	0.15 – 0

2. a) Estimate the potential rate of accretion of Pangaea (in km^3/y) since it began to breakup 200 Ma. Assume the modern rates of arc and oceanic island-plateau growth of $1.1 \text{ km}^3/\text{y}$ and $0.2 \text{ km}^3/\text{y}$, respectively. In addition to these sources of potential growth, sediment on the ocean floor may be accreted to Pangaea. The total volume of sediments on the sea floor today is about $9.4 \times 10^7 \text{ km}^3$ and the mean age of these sediments is 60 Ma.

b) Compare your rate of potential growth to that estimated from geologic maps and from Nd isotopic studies. These data suggest that about $25 \times 10^6 \text{ km}^2$ were added to Pangaea over the last 200 My. Because much of this was oceanic arcs, assume an average crustal thickness of 25 km.

c) Discuss the significance of your results.

3. Using the experimental data below for the solidi of granite magma, discuss the probability that granites form on Venus. To do this, plot the data on a depth (km)–

temperature (°C) graph like Figure 10.25. Transfer a continental geotherm (Fig. 3.14) and the Venusian geotherm (Fig. 10.25) to your diagram. Contrast granite magma production and emplacement on both planets. Is it more or less likely that felsic magmas will be extruded on Venus than on Earth and why?

Granite Solidi Pressures and Temperatures		
Pressure (kbar)	Wet (°C) (Earth)	Dry (°C) (Venus)
0	1150	1150
1	950	1160
2	825	1165
3	780	1175
5	725	1190
10	700	1250
11	680	1275

CHAPTER 11

1. a) It is unlikely that much continental crust was above sea level prior to the Late Archean. How would this affect the carbonate-silicate weathering cycle? Construct a computer-generated flow diagram of the Early Archean carbonate-silicate cycle and discuss how changes in input/output of CO₂ might affect both the composition of the atmosphere, and the deposition of carbonates.

b) Assume CO₂ degassing of the Earth follows the exponential relation with time.

$$V = V_0 e^{wt} \quad (11a)$$

where V and V_0 are past and present degassing rates, t is time (age), and w is a constant equivalent to 0.4. Also, assume that to a first approximation continental growth rate and degassing rate are related by,

$$dA/dt = -c(dV/dt), \quad (11b)$$

where c is a constant such that $cV_0 = 0.2$. A and A_0 are continental areas past and present ($A_0 = 2 \times 10^8 \text{ km}^2$). Derive an expression relating A/A_0 to t and plot A/A_0 versus t on an x-y graph (t in 100 My increments from 4 to 0 Ga).

c) Today mean elevation of the continents (E in km) relative to sea level ($E = 0$) is related to continental area (A in km^2) by,

$$E = -1.5 + 1.1 \log A \quad (11c)$$

Using the expression you derived in part b), when should continents have first emerged above sea level? Is your answer consistent with the geologic record?

Answer the same questions if continents grew episodically according to curve 4 in Figure 10.6. Discuss the assumptions and results of the calculations in terms of the carbonate-silicate weathering cycle.

2. In the absence of an atmosphere and oceans, the flux of solar energy incident on a planetary surface (S) and the albedo (a) essentially control the surface temperature as follows,

$$T(^{\circ}\text{K}) = [\{S(1-a)\}/4\sigma]^{1/4} \quad (11d)$$

a) Using the data in Table 11.7, calculate the surface temperature as a function of time from 4.5 Ga to the present for Earth and Venus, for the case where neither has an atmosphere or oceans. Plot your results on a temperature versus time graph and compare calculated present temperatures to those observed (express temperature as degrees C, log scale on the y-axis) σ is the Boltzman constant ($5.67 \times 10^{-5} \text{ erg/cm}^2 \text{ sec deg}^4$).

TABLE 11.7

	Present Mean Surface Temperature $^{\circ}\text{C}$	Present Albedo (a)	Albedo a at 4.5 Ga	Incident Solar energy flux (S) ($[\text{erg/cm}^2 \text{ sec}] \times 10^6$)	
				Today	4.5 Ga
Earth	20	0.33	0.1	1.4	1.0
Venus	500	0.76	0.1	2.6	2.0

b) It is likely that CO_2 degassed and accumulated rapidly in the atmosphere of both planets. For the purpose of this problem, assume that this early atmosphere heated to an average temperature of 10°C for the Earth and 50°C for Venus by 4 Ga. Using the results from part a), calculate the greenhouse contribution (in %) to atmospheric warming at 4 Ga and today for each planet. Assume both planetary surfaces began at the temperatures calculated in part a) at 4.5 Ga. Plot results on your temperature-time graph in part a) and also plot them on a time versus percent greenhouse effect graph.

c) Discuss the significance of your calculations and graphs.

3. a) The cumulative number of small asteroids (N) that could potentially collide with an inner planet is given by the following approximate empirical relationship,

$$N = ke^{\gamma M} \quad (11e)$$

where M is the absolute magnitude of the asteroid and γ and k are constants ($k = 6 \times 10^{-5}$ and $\gamma = 0.8$). Also, it has been noted that asteroid diameter (d) and magnitude are related by,

$$\log d = 3.12 - 0.2M - 0.5 \log 0.037 \quad (11f)$$

Calculate the size and mass of the largest possible asteroid that may have collided

with the Earth at the Cretaceous–Tertiary boundary. To do this, assume $N = 1$ in eq. 11e.

b) Compare your calculated mass to that estimated from the surface Ir density at Gubbio, Italy of 3.4×10^{14} kg for an asteroid with a density of 1.7 gm/cm^3 . What size of asteroid is needed to account for the mass estimated from the Gubbio data?

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