Plate Tectonics and Crustal Evolution

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New Mexico Institute of Mining and Technology

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Preface

This book has grown out of a course I teach with the same title. The rapid accumulation of data related to sea-floor spreading and plate tectonics in the last decade has necessitated continued updating of the course. Although new data are still coming in, the rate of increase of fundamentally important data has decreased somewhat and the time seems right for a textbook to accompany the course. The book is written for an advanced undergraduate or graduate-level student. 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 it in a systematic manner addressing problems related to the evolution of the earth's crust over the last 3.5 b. years. The role of plate tectonics in the geologic past is examined in light of existing geologic evidence and examples of plate reconstructions are discussed.

In order to keep the book a reasonable length and so as not to duplicate extensively information 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 geologic, geochemical, and geophysical data are gathered are only briefly mentioned as books on these subjects are 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 many of the 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.

It is not possible to acknowledge everyone who has influenced my opinions in writing the book. The following individuals, however, gave freely of their time to review and criticize one or more chapters and I gratefully acknowledge their help in arriving at a final version: Fred Anderson, Richard Armstrong, Gale Billings, James Case, Peter Coney, Stanley Hart, Peter Lipman, John MacMillan, Paul Mueller, Denis Shaw, and Ross Taylor. It should be pointed out that the interpretations I present in the book do not necessarily reflect those of any of the above. I am also grateful to those publishers and authors who have allowed reproduction of many of the figures used in the book. Dennis Umshler, Charles O'Melveny, Michael Graham, George Ross, and Susan Williams are acknowledged for their assistance in compiling data and references and Carolyn Condie for her help in editing. Rosalie Anaya, Jeanette Chavez, Rosie Trujillo, and Petra Apodaca typed the manuscript at various stages.

Contents

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 dating 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 Mars 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 last 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 has evolved. Most 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 earth's 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 appear to have been added partially to continental margins, causing the continents to grow laterally, and then partially returned to the mantle (the region between the crust and the earth's core), 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 has subsequently grown are not well known, but the large amount of data currently available allows some 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 sea-floor spreading and plate tectonics that have so profoundly influenced geologic thinking in the last decade have also provided valuable insight into the mechanisms by which the crust has evolved. One of the major problems regarding crustal evolution (addressed in Chapter 10) is that of discovering when in geologic time plate-tectonic and sea-floor 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 some new unifying theory or theories are proposed that explain a great deal of the accumulated data. Physics and chemistry underwent such revolutions near or before the beginning of the 20th century, whereas the earth sciences may be just entering such a revolution. As with scientific revolutions in other fields, new ideas and interpretations in the earth sciences do not invalidate earlier observations. On the contrary, the theories of sea-floor spreading and plate tectonics offer for the first time a unified explanation for heretofore seemingly unrelated observations in the fields of geology, paleontology, and geophysics.

At the onset, it is necessary to briefly discuss sea-floor spreading and plate tectonics and to introduce a few terms that will be used throughout the book. Most geophysical data suggest that today the surface of the earth is composed of rigid plates 50-150 km thick, known collectively as the lithosphere. These plates rest on a hotter, more dense layer that deforms plastically and is known as the asthenosphere (Fig. 1-1). The upper part of the lithosphere (6-40 km deep) is composed of the crust. An important part of the sea-floor spreading theory is that new lithosphere and crust are continually being created at oceanic rises by injection and eruption of magma derived from the asthenosphere. This lithosphere spreads laterally away from rises and is finally consumed by the asthenosphere in subduction zones. The lithosphere can be considered, to a first approximation, as a mosaic of plates bounded by oceanic rises, subduction zones, and transform faults (boundaries along which plates slide by each other) (Fig. 1-1). Intracontinental compressive zones (a less common type of plate boundary) are not depicted in Fig. 1-1. The study of the interactions of lithospheric plates is known as plate tectonics.

THE APPROACH

The general format of the book will evolve from one primarily of presentation of data critical to models for the origin and evolution of the crust to one of interpretation and speculation. Chapter 2 is concerned chiefly with the gross physical and chemical features of the earth; it serves as a basic framework for later discussions. Hypotheses regarding the origin of the earth (and moon) are also briefly reviewed in this chapter.

An account of the detailed structure and composition of the mantle is given in Chapter 3. This provides important information bearing both on the source of the crust and on sea-floor spreading, which appears to result from dynamic processes in the upper mantle. The seismic, heat flow, gravity, magnetic, and electrical properties of the crust, together with a discussion of crustal composition, are presented in Chapter 4. In Chapter 5 the chief methods of radiometric dating used to define crustal age provinces are discussed as are the types of events that can be dated. A survey of both Phanerozoic and Precambrian crustal provinces is then presented in which overall structure, rock distributions, and compositions are considered.

In Chapter 6 the theory of sea-floor spreading is presented and the evidences that led to its formulation are reviewed. Lithospheric plates are described and hypotheses for the causes of sea-floor spreading are discussed. Magma associations on the earth are described in Chapter 7 with reference to a plate-tectonic framework. The origin of magmas, which is of considerable importance in models for crustal origin and growth, is also considered in light of existing field, experimental, geochemical, and geophysical data.

Chapter 8 is concerned with the principles of plate tectonics and continental drift and presents a discussion of the methods available for reconstruction of plate positions in the geologic past. The relationship of orogeny (mountain building) to plate tectonics is also reviewed in this chapter. In Chapter 9 specific examples of plate reconstructions are given to illustrate methods set forth in Chapter 8, with emphasis on Phanerozoic reconstructions for which the most data are available.

In Chapter 10 an attempt is made to bring together existing ideas on the origin and growth of the crust and on the role of plate tectonics in crustal development as a function of time. Secular compositional changes in the crust and a brief account of the origin of the atmosphere and oceans are also discussed.

Although methods and techniques of acquiring data will not be extensively discussed in the book, it is perhaps appropriate to briefly review at the outset some of the more important methods and at the same time introduce some basic terms.

METHODS AND DEFINITIONS

Seismic

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. 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 at which 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.* These various types of seismic wave motions are illustrated in Fig. 1-2.

Elastic waves are detected by seismometers, which respond to ground displacements, or in some cases to wave velocities. Short-period seismometers are used to detect body waves and long-period seismometers are used for surface waves. Computerized arrays of seismometer stations have recently made 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 (Bott, 1971). The gross features of the earth's interior, as discussed in the next chapter, are determined from travel-time distance studies of body waves traveling 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 time and location of the explosion are known more accurately than are earthquake times and hypocenter locations. The refraction method, which is used both on land and at sea, is based on measuring the travel-times of P waves between shot points and seismic recorders located at various distances apart, usually along straight-line profiles. The method is limited in that very detailed crustal structure cannot be determined and low-velocity layers cannot be detected. The use of supercritical reflections (i.e., reflected waves that have incident angles greater than the critical angle) can enhance the interpretations of refraction data. Vertical incidence reflections occur only at sharp discontinuities and often allow the seismologist to distinguish between sharp and gradational discontinuities.

Travel-time anomaly studies have recently proved 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* (Garland, 1971). Two types of oscillations occur. Torsional oscillations involve particle displacements normal to the earth's radius and spheroidal oscillations, which are radial or tangential displacements. Surface waves are examples of short-period free oscillations. Long-period free oscillations are detected with strain seismometers and earth-tide gravimeters. Free-oscillation studies have recently resulted in improved resolution and detection of interfaces within the earth, as well as direct determination of density and seismic anelasticity of parts of the mantle. Rayleigh- and Love-wave dispersion (i.e., the variation of velocity with wavelength) provides a basis for detailed studies of crustal and upper mantle structure. Free oscillations produced by major earthquakes do not last indefinitely but the vibrational energy is gradually converted to heat as the earth comes to rest. The oscillations are said to be 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*). Low values of Q mean high seismic-wave attenuation. Experimental measurements of Q in rocks indicate a range from about 50 to 1000. Anelasticity in the earth appears to result from some combination of grain boundary damping, stress-induced ordering of crystal defects, and damping caused by vibration of dislocations (Gordon and Nelson, 1966). Q is particularly sensitive to temperature and partial melting.

In oceanic areas, seismic reflection profiling is useful. The method is similar to echo-sounding except that a more powerful acoustic source is used. Reflections from interfaces in the sediment layer on the ocean bottoms and from the sediment-basement layer interface are received by towed hydrophone arrays.

Gravity

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 gals (1 gal = 1cm/sec^2). Gravity is measured with a gravimeter and can be determined both on land and at sea (Bott, 1971; Garland, 1971). Accuracies are typically of about 1 mgal on land and 5-10 mgal at sea. The standard reference for gravity on the earth is the gravitational field of a spheroid and is dependent only on latitude. It has recently become possible to determine accurately the gravity field on the earth from the data derived from the directions and rates of orbital shift of artificial satellites. From such data, it is possible to determine how much the average surface of the earth or *geoid*, which is roughly equal to sea level, actually deviates from a spheroid. Existing data indicate that the earth is pear-shaped with an average equitorial radius of 6378 km and an average polar radius of 6357 km. Gravity distribution on the earth can be accurately calculated from spherical harmonic coefficients of the satellite gravitational data and is discussed in Chapter 3.

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. Next if the attraction of the 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 by Bouguer 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 columns of rock have identical mass 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 thickness of the columns. Both mechanisms probably contribute to isostatic compensation. *Isostatic anomalies* may be calculated by subtracting from Bouguer anomalies the mass distribution within some segment of the upper part of the earth as determined from some combination of the Airy and Pratt compensation mechanisms.

Magnetic

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—i.e., the *declination*, and the angle of dip with the horizontal—i.e., the *inclination*. The inclination becomes vertical at the two magnetic poles. The total magnetic field strength is strongest near the magnetic poles (0.7 oersted at the South Pole) and weakest at the equator (about 0.3 oersted). 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 earth's magnetic field in terms of fluid motions in the outer part of the earth's core. Approximately 90 percent of the present field can be described by a magnetic dipole at the earth's center, which makes an angle of about 11.5 degrees with the rotational axis. A general westward drift of the field is noted at a rate of about 0.18 deg/year.

Local and regional variations in the magnetic field reflect, for the most part, rocks beneath the surface with varying degrees of magnetization. Such variations are measured with fluxgate or proton magnetometers on land, 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 ($1\gamma = 10^{-5}$ oersted). 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 6. Such magnetization is known as remnant magnetization and is measured in the laboratory with spinner or astatic magnetometers. The maximum temperature at which a mineral can become magnetized is known as the *Curie point* tempreature. *Paleomagnetism* is the study of remnant 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 (see Chapter 8). Paleomagnetic studies have shown that the magnetic poles have reversed themselves many times in the geologic past such *reversals* are thought to be produced by instability in the outer core.

Electrical

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 in the earth while longer periods penetrate to greater depths.

Four methods have been used to estimate conductivity distribution in the crust and mantle (Keller, 1971): (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 measuring voltage drops between electrodes. The depth of penetration of this method is limited only to 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 in naturally induced currents caused by magnetic storms. This currently provides the best method for estimating mantle conductivity distributions. Results of these methods may be presented as conductivities ($ohm^{-1} m^{-1}$) or resistivities (ohm-m).

Geothermal

Heat-flow measurements on the earth involve two separate measurements, one of thermal gradient (dT/dx) and one of thermal conductivity (K). From these measurements, heat flow (q) (usually expressed as $\mu cal/cm^2$-sec or as heat-flow units where 1HFU = 1 $\mu cal/cm^2$-sec) is calculated as follows:

$$q = -K\frac{dT}{dx}. \qquad (1\text{-}1)$$

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 dry 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 the 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, 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 anamalously 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 percent 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 U^{235}, U^{238}, Th^{232}, and K^{40}. The concentrations of these elements can be determined by counting the natural radioisotopes with a gamma-ray spectrometer in the lab or in the field. Airborne gamma-ray spectrometers have recently been used to estimate concentrations of U, Th, and K over areas of the crust. Radiogenic heat generation (A) is expressed in 10^{-13} cal/cm^3 sec or heat generation units (HGU) where 1 HGU = 10^{-13} cal/cm^3 sec.

High Pressure-Temperature Studies

In the past decade rapid advances in the field of high-pressure technology allow reconstruction in the laboratory of static pressures up to about 200 kb and more recently to more than 300 kb (Ringwood, 1969a; Kumazawa, *et al.*, 1974). These pressures are equivalent to about 600 km and 1000 km burial depth, respectively. Many 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, it is possible (in conjunction with available geophysical data) to place limitations on the composition, mineralogy, and melting behavior of the crust and upper mantle and to evaluate quantitatively the origin of magmas. From the high pressure and temperature rock-deformation studies, it is possible to understand more fully earthquake mechanisms and flow characteristics within the earth.

Recently it has been feasible to study possible mineral assemblages and compositions of deeper parts of the earth including the core from 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 Petersen, 1969). The pressure and density within the wave can be deduced from measuring the shock and particle velocities. 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 major limitations on the composition of the mantle and core.

Geochemistry, Geochronology, and The Geologic Time Scale

Geochemical data from a variety of rocks and minerals (especially from igneous rocks) provide important information bearing on the composition of the upper mantle and evolutionary changes in the crust and mantle. Alkali, alkaline-earth, and rare-earth elements have proved to be especially valuable in constructing models of magma evolution. Instrumental analytical techniques such as X-ray fluorescence, neutron activation, mass spectrometry, and atomic absorption, as well as rapid wet chemical methods, currently make it possible to analyze large numbers of samples for major and many trace elements with high degrees of precision and accuracy (Energlyn and Brealey, 1971; Wainerdi and Uken, 1971).

Isotopic studies (employing mass spectro-

Table 1-1 The Geologic Time Scale

Relative Durations of Major Geologic Intervals
CENOZOIC
MESOZOIC
PALEOZOIC
PRECAMBRIAN

Era	Period	Epoch	Duration in Millions of Years (Approx.)	Millions of Years Age (Approx.)
	Quaternary	Recent	Approx. last 5,000 years	
		Pleistocene	25	2.5
		Pliocene	45	7
		Miocene	19	26
		Oligocene	12	38
		Eocene	16	54
Cenozoic	Tertiary	Paleocene	11	65
	Cretaceous		71	136
	Jurassic		54	190
Mesozoic	Triassic		35	225
	Permian		55	280
	Carboniferous: Pennsylvanian		45	325
	Carboniferous: Mississippian		20	345
	Devonian		50	395
	Silurian		35	430
	Ordovician		70	500
Paleozoic	Cambrian		70	570
Proterozoic			~1900	2500
Archean			~2100	4600

Millions of Years: 0, 50, 100, 150, 200, 250, 300, 350, 400, 450, 500, 550, 4,600

Modified from Don L. Eicher, *Geologic Time*, (c) 1968, p. 150 by permission of Prentice-Hall, Inc., Englewood Cliffs, N.J.

metry) are important not only in terms of *geochronology* but also for *tracer studies.* Pb and Sr isotopes are especially important in both cases. 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.

Refinements in radiometric dating methods in recent years (see Chapter 5) have improved estimates of the beginning and duration of the various subdivisions of geologic time. A current version of the geologic time scale is shown in Table 1-1. Geologic time is divided into eras, periods, and epochs. Five major eras are recognized; from youngest to oldest, they are the *Cenozoic, Mesozoic, Paleozoic, Proterozoic,* and *Archean.* The three younger eras are collectively known as the *Phanerozoic* and the two older ones—the *Precambrian.* The Precambrian comprises almost 90 percent of geologic time and the oldest dated crust is about 3.8 b. years in age.

Other Methods and 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 (Anderson, 1966). 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 equitorial 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 earth's 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. An especially designed drilling ship, the *Glomar Challenger,* is used as a floating drilling platform. A great deal of information pertaining to sediment ages and lithologies and to the history of the sea floor over the last 150 m. years should become available from these and future cores as they are studied. 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.

Chemical studies of meteorites and lunar samples as well as spectroscopic studies of the sun and other stars provide an important framework upon which to interpret the overall composition of the earth and other planets. Last, but not least, are the conventional and well-established geological methods. Perhaps the most commonly overlooked yet extremely important source of data is from *field geology.* The results of widespread geological mapping on the continents are of critical importance to the evaluation of the roles of sea-floor spreading and plate tectonics in the geologic past. *Stratigraphy* (the correlation of rock units between different geographic areas), *petrology* (the study of the origin of rocks), and *paleontology* (the study of fossils) are other important fields of investigation.

SUGGESTIONS FOR FURTHER READING

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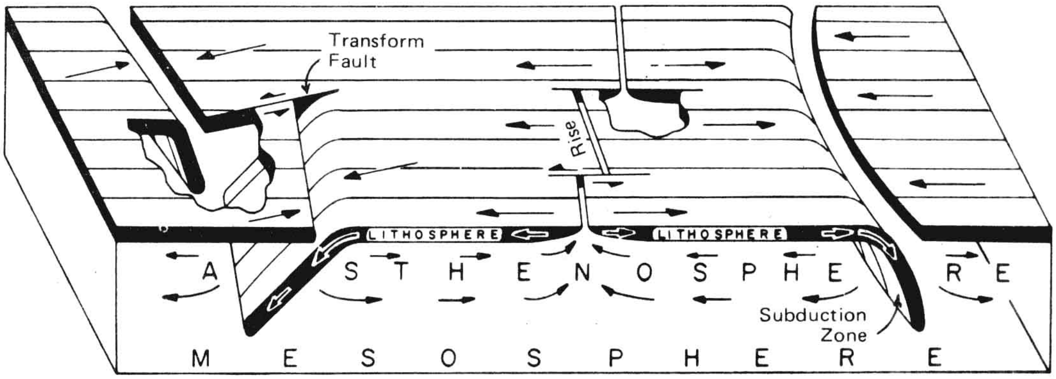
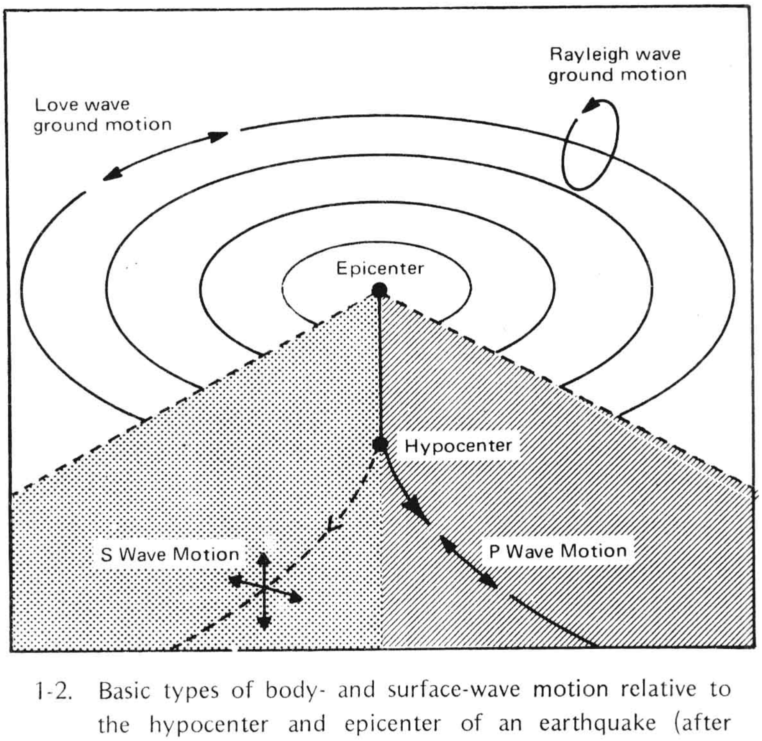


1-1. Schematic three-dimensional diagram showing the major features of sea-floor spreading (after Isacks *et al.*, 1968).



1-2. Basic types of body- and surface-wave motion relative to the hypocenter and epicenter of an earthquake (after Davies, 1968).