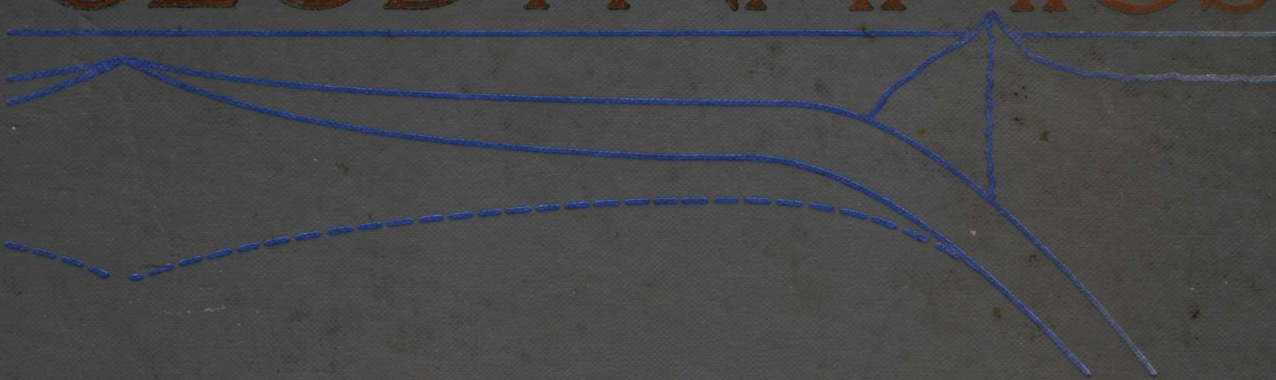
GEODYNAMICS



APPLICATIONS OF CONTINUUM PHYSICS TO GEOLOGICAL PROBLEMS

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GEODYNAMICS
Applications of Continuum Physics to Geological Problems

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preface

This textbook deals with the fundamental physical processes necessary for an understanding of plate tectonics and a variety of geological phenomena. We believe that the appropriate title for this material is *geodynamics*. The contents of this textbook evolved from a series of courses given at Cornell University and UCLA to students with a wide range of backgrounds in geology, geophysics, physics, mathematics, chemistry, and engineering. The level of the students ranged from advanced undergraduate to graduate.

In all cases we present the material with a minimum of mathematical complexity. We have not introduced mathematical concepts unless they are essential to the understanding of physical principles. For example, our treatment of elasticity and fluid mechanics avoids the introduction or use of tensors. We do not believe that tensor notation is necessary for the understanding of these subjects or for most applications to geological problems. However, solving partial differential equations is an essential part of this textbook. Many geological problems involving heat conduction and solid and fluid mechanics require solutions of such classic partial differential equations as Laplace's equation, Poisson's equation, the biharmonic equation, and the diffusion equation. All these equations are derived from first principles in the geological contexts in which they are used. We provide elementary explanations for such important physical properties of matter as solid-state viscosity, thermal coefficient of expansion, specific heat, and permeability. Basic concepts involved in the studies of heat transfer, Newtonian and non-Newtonian fluid behavior, the bending of thin elastic plates, the mechanical behavior of faults, and the interpretation of gravity anomalies are emphasized. Thus it is expected that the student will develop a thorough understanding of such fundamental physical laws as Hooke's law of elasticity, Fourier's law of heat conduction, and Darcy's law for fluid flow in porous media.

The problems are an integral part of this textbook. It is only through solving a substantial number of exercises that an adequate understanding of the underlying physical principles can be developed. Answers to selected problems are provided.

The first chapter reviews plate tectonics; its main purpose is to provide physics, chemistry, and engineering students with the geological background necessary to understand the applications considered throughout the rest of the textbook. We hope that the geology student can also benefit from this summary of numerous geological, seismological, and paleomagnetic observations. Since plate tectonics is a continuously evolving subject, this material may be subject to revision. Chapter 1 also briefly summarizes the geological and geophysical characteristics of the other planets and satellites of the solar system. Chapter 2 introduces the concepts of stress and strain and discusses the measurements of these quantities in the earth's crust. Chapter 3 presents the basic principles of linear elasticity. The bending of thin elastic plates is emphasized and is applied to problems involving the bending of the earth's lithosphere. Chapter 4 deals mainly with heat conduction and the application of this theory to temperatures in the continental crust and the continental and oceanic lithospheres. Heat transfer by convection is briefly discussed and applied to a determination of temperature in the earth's mantle. Surface heat flow measurements are reviewed and interpreted in terms of the theory. The sources of the earth's surface heat flow are discussed. Problems involving the solidification of magmas and extrusive lava flows are also treated. The basic principles involved in the interpretation of gravity measurements are given in Chapter 5. Fluid mechanics is studied in Chapter 6; problems involving mantle convection and postglacial rebound are emphasized. Chapter 7 deals with the rheology of rock or the manner in which it deforms or flows under applied forces. Fundamental processes are discussed from a microscopic point of view. The mechanical behavior of faults is discussed in Chapter 8 with particular attention being paid to observations of displacements along the San Andreas fault. Finally, Chapter 9 discusses the principles of fluid flow in porous media, a subject that finds application to hydrothermal circulations in the oceanic crust and in continental geothermal areas.

The contents of this textbook are intended to provide the material for a coherent one-year course. In order to accomplish this goal, some important aspects of geodynamics have had to be omitted. In particular, the fundamentals of seismology are not included. Thus the wave equation and its solutions are not discussed. Many seismic studies have provided important data relevant to geodynamic processes. Examples include (1) the radial distribution of density in the earth as inferred from the radial profiles of seismic velocities, (2) important information on the locations of plate boundaries and the locations of descending plates at ocean trenches provided by accurate determinations of the epicenters of earthquakes, (3) details of the structure of the continental crust obtained by seismic reflection profiling using artificially generated waves. An adequate treatment of seismology would have required a very considerable expansion of this textbook. Fortunately, there are a number of excellent textbooks on this subject.

A comprehensive study of the spatial and temporal variations of the earth's magnetic field is also considered to be outside the scope of this textbook. A short discussion of the earth's magnetic field relevant to paleomagnetic observations is given in Chapter 1. However, mechanisms for the generation of the earth's magnetic field are not considered.

In writing this textbook, several difficult decisions had to be made. One was the choice of units; we use SI units throughout. This system of units is defined in Appendix 1. We feel there is a strong trend toward the use of SI units in both geology and geophysics. We recognize, however, that many cgs units are widely used. Examples include μcal cm^{-2} s^{-1} for heat flow, kilobar for stress, and milligal for gravity anomalies. For this reason we have often included the equivalent cgs unit in parenthesis after the SI unit, for example, MPa (kbar). Another decision involved the referencing of original work. We do not believe that it is appropriate to include a large number of references in a basic textbook. We have credited those individuals making major contributions to the development of the theory of plate tectonics and continental drift in our brief discussion of the history of this subject in Chapter 1. We also provide references to data. At the end of each chapter a list of recommended reading is given. In many instances these are textbooks and reference books, but in some cases review papers are included. In each case the objective is to provide background material for the chapter or to extend its content.

Many of our colleagues have read all or parts of various drafts of this textbook. We acknowledge the contributions made by Jack Bird, Peter Bird, Muawia Barazangi, Allan Cox, Walter Elsasser, Robert Kay, Suzanne Kay, Mark Langseth, Bruce Marsh, Jay Melosh, John Rundle, Sean Solomon, David Stevenson, Ken Torrance, and David Yuen. We particularly wish to acknowledge the many contributions to our work made by Ron Oxburgh and the excellent manuscript preparation by Tanya Harter.

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chapter 1

PLATE TECTONICS

1-1 INTRODUCTION

Plate tectonics is a model in which the outer shell of the earth is divided into a number of thin, rigid plates that are in relative motion with respect to one another. The relative velocities of the plates are of the order of a few tens of millimeters per year. A large fraction of all earthquakes, volcanic eruptions, and mountain building occurs at plate boundaries. The distribution of the major surface plates is illustrated in Figure 1-1.

The plates are made up of relatively cool rocks and have an average thickness of about 100 km. The plates are being continually created and consumed. At ocean ridges adjacent plates diverge from each other in a process known as *seafloor spreading*. As the adjacent plates diverge, hot mantle rock ascends to fill the gap. The hot, solid mantle rock behaves like a fluid because of solid-state creep processes. As the hot mantle rock cools, it becomes rigid and accretes to the plates, creating new plate area. For this reason ocean ridges are also known as *accreting plate boundaries*. The global system of ocean ridges is denoted by the heavy dark line in Figure 1-1.

Because the surface area of the earth is essentially constant, there must be a complementary process of plate consumption. This occurs at ocean trenches. The surface plates bend and descend into the interior of the earth in a process known as *subduction*. At an ocean trench the two adjacent plates converge, and one descends beneath the other. For this reason ocean trenches are also known as *convergent plate boundaries*. The worldwide distribution of trenches is shown in Figure 1-1 by the lines with triangular symbols, which point in the direction of subduction.

A cross-sectional view of the creation and consumption of a typical plate is illustrated in Figure 1-2. That part of the earth's interior that comprises the plates is referred to as the *lithosphere*. The rocks that make up the lithosphere are relatively cool and rigid; as a result the interiors of the plates do not deform significantly as they move about the surface of the earth. As the plates move away from ocean ridges, they cool and thicken. The solid rocks beneath the lithosphere are sufficiently hot to be able to deform freely; these rocks comprise the *asthenosphere*, which lies below the lithosphere. The lithosphere slides over the asthenosphere with relatively little resistance.

As the rocks of the lithosphere become cooler, their density increases because of thermal contraction. As a result the lithosphere becomes gravitationally unstable with respect to the hot asthenosphere beneath. At the ocean trench the lithosphere bends and sinks into the interior of the earth because of this negative buoyancy. Major faults separate descending lithospheres from adjacent overlying lithospheres. These faults are the sites of a large fraction of the great earthquakes. Examples are the Chilean earthquake in 1960 and the Alaskan earthquake in 1964. These are the largest earthquakes that have occurred since modern seismographs have been available. The locations of the descending lithospheres can be accu-

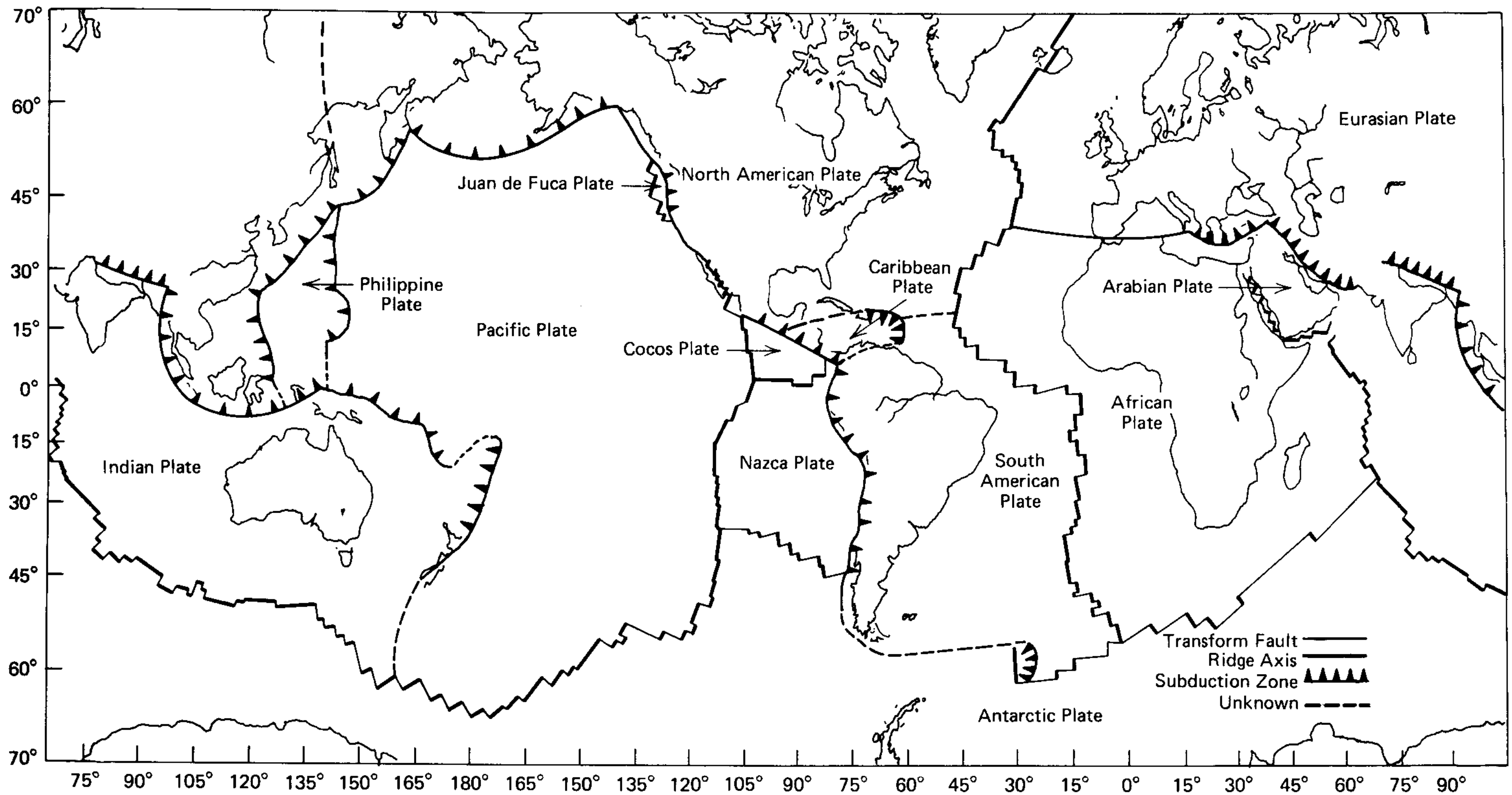

Figure 1-1 Distribution of the major surface plates.

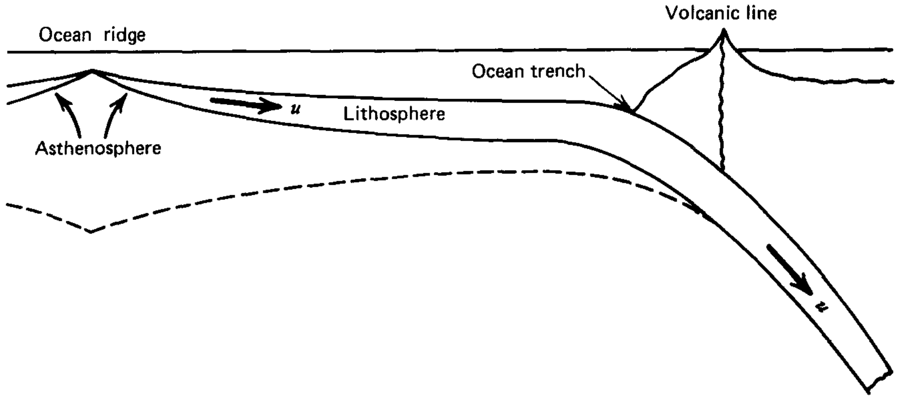

Figure 1-2 Accretion of a lithospheric plate at an ocean ridge and its subduction at an ocean trench. The asthenosphere, which lies beneath the lithosphere, is shown.

rately determined by the earthquakes occurring in the cold, brittle rocks of the lithospheres. These planar zones of earthquakes associated with subduction are known as *Benioff zones*.

Lines of active volcanoes lie parallel to almost all ocean trenches. These volcanoes occur about 150 km above the descending lithosphere. If these volcanoes stand on the seafloor, they form an *island arc*, as typified by the Aleutian Islands in the North Pacific. The remarkable linearity of these volcanic lines is illustrated in the photograph of three volcanic islands in the Aleutian arc in Figure 1-3. If the trench lies adjacent to a continent, the volcanoes grow from the land surface. This is the case in the western United States, where a volcanic line extends from Mt. Baker in the north to Mt. Shasta in the south. Mt. St. Helens, the site of a violent eruption in 1980, forms a part of this volcanic line.

The earth's surface is divided into continents and oceans. The oceans have an average depth of about 5 km, and the continents rise above sea level. The reason for this difference in elevation is the difference in the thickness of the crust. Crustal rocks have a different composition from that of the mantle rocks beneath and are less dense. The crustal rocks are therefore gravitationally stable with respect to the heavier mantle rocks. There is usually a well-defined boundary, the *Moho*, or Mohorovičić discontinuity, between the crust and mantle. A typical thickness for *oceanic crust* is about 6 km; *continental crust* is about 35 km thick. Although oceanic crust is gravitationally stable, it is sufficiently thin so that it does not significantly impede the subduction of the gravitationally unstable oceanic lithosphere. The oceanic lithosphere is continually cycled as it is accreted at ocean ridges and subducted at ocean trenches. Because of this cycling the average age of the ocean floor is about 10^8 years.

On the other hand, the continental crust is sufficiently thick and gravitationally stable so that it is not subducted at an ocean trench. For this reason the rocks of the continental crust, with an average age of about 10^9 years, are much older than the rocks of the oceanic crust. As the lithospheric plates move across the surface of the earth, they carry the continents with them. The relative motion of continents is referred to as *continental drift*.

Much of the historical development leading to plate tectonics concerned the validity of the hypothesis of continental drift, that the relative positions of continents change during geologic time. The similarity in shape between the west coast of Africa and the east coast of South America was noted as early as 1620 by Francis Bacon. This "fit" has led many authors to speculate on how these two continents might have been attached. A detailed exposition of the hypothesis of continen-

Figure 1-3 View looking east from Mt. Moffett on Adak Island in the Aleutians. Also shown are Adagdak volcano in the foreground and Great Sitkin volcano in the distance. (Photograph courtesy of Bruce Marsh.)

tal drift was put forward by Frank B. Taylor in 1910.* The hypothesis was further developed by Alfred Wegener beginning in 1912 and summarized in his book *The Origin of Continents and Oceans*.** As a meteorologist, Wegener was particularly interested in the observation that glaciation had occurred in equatorial regions at the same time that tropical conditions prevailed at high latitudes. This observation in itself could be explained by *polar wander*, a shift of the rotational axis without other surface deformation. However, Wegener also set forth many of the qualitative arguments that the continents had formerly been attached. In addition to the observed fit of continental margins these included the correspondence of geological provinces, continuity of structural features such as relict mountain ranges, and the correspondence of fossil types. Wegener argued that a single supercontinent, Pangaea, had formerly existed. He suggested that tidal forces or forces associated with the rotation of the earth were responsible for the breakup of this continent and the subsequent continental drift.

Further and more detailed qualitative arguments favoring continental drift were presented by Alexander du Toit, particularly in his book *Our Wandering Continents* published in 1937.* Instead of a single supercontinent du Toit argued that there had formerly been a northern continent,

*"Bearing of the Tertiary mountain belt on the origin of the Earth's plan," *Bulletin of the Geological Society of America*, vol. 21, pp. 179–226.

**Dover, New York, 1966, translation of the fourth revised German edition.

*Oliver and Boyd, Edinburgh.

Laurasia, and a southern continent, Gondwanaland, separated by the Tethys Ocean.

During the 1950s extensive exploration of the seafloor led to an improved understanding of the worldwide range of mountains on the sea floor known as mid-ocean ridges. In 1961 Harry Hess* hypothesized that the seafloor was created at the axis of a ridge and moved away from the ridge to form an ocean in a process now referred to as *seafloor spreading*. This process explains the similarity in shape between continental margins. As a continent breaks apart, a new ocean ridge forms. The ocean floor created is formed symmetrically at this ocean ridge, creating a new ocean. This is how the Atlantic Ocean was formed; the mid-Atlantic ridge where the ocean formed now bisects the ocean.

It should be realized, however, that the concept of continental drift won general acceptance by earth scientists only in the period between 1967 and 1970. Although convincing qualitative, primarily geological, arguments had been put forward to support continental drift, almost all earth scientists and, in particular, almost all geophysicists had opposed the hypothesis. Their opposition was mainly based on arguments concerning the rigidity of the mantle and the lack of an adequate driving mechanism.

The propagation of seismic shear waves showed beyond any doubt that the mantle was a solid. An essential question was how horizontal displacements of thousands of kilometers could be accommodated by solid rock. The fluidlike behavior of the earth's mantle had been established in a general way by gravity studies carried out in the latter part of the nineteenth century. Measurements showed that mountain ranges had low-density roots. The lower density of the roots provides a negative relative mass that nearly equals the positive mass of the mountains. This behavior could be explained by the principle of *hydrostatic equilibrium* if the mantle behaved as a fluid. Mountain ranges appear to behave similarly to blocks of wood floating on water.

The fluid behavior of the mantle was established quantitatively by N. A. Haskell in 1935.* Studies of the elevation of beach terraces in Scandinavia showed that the earth's surface was still rebounding from the load of the ice during the last ice age. By treating the mantle as a viscous fluid, Haskell was able to explain the present uplift of Scandinavia if the mantle has a viscosity of about 10^{20} Pa s (10^{21} poise). (Note that SI units are used in this textbook; however, in many cases the values in cgs units are given in parentheses. Details on units are provided in Appendix 1.) Although this is a very large viscosity (water has a viscosity of 10^{-3} Pa s), it leads to a fluid behavior for the mantle during long intervals of geologic time.

In the 1950s theoretical studies had established several mechanisms for the very slow creep of crystalline materials. This creep results in a fluid behavior. In 1965, Robert B. Gordon** showed that solid-state creep quantitatively explained the viscosity determined from observations of postglacial rebound. At temperatures that are a substantial fraction of the melt temperature thermally activated creep processes allow mantle rock to flow at low stress levels on time scales greater than 10^4 years. The rigid lithosphere includes rock that is sufficiently cold to preclude creep on these long time scales.

Forces must act on the lithosphere in order to make the plates move. Wegener suggested that either tidal forces or forces associated with the rotation of the earth caused the motion responsible for continental drift. However, in the 1920s Sir Harold Jeffreys, as summarized in his book *The Earth*,† showed that these forces were insufficient. Some other mechanism had to be found to drive the motion of the plates. Any reasonable mechanism must also have sufficient energy available to

*"History of Ocean Basins," in *Petrologic Studies, A Volume in Honour of A. E. Buddington*, A. E. J. Engle, pp. 599–620, Geological Society of America, 1962.

*"The motion of a viscous fluid under a surface load," Physics, vol. 6, pp. 265–269, vol. 7, pp. 56–61.

**"Diffusion creep in the earth's mantle," Journal of Geophysical Research, vol. 70, pp. 2413–2418.

†Cambridge University Press, first edition 1924, fifth edition 1970.

provide the energy being dissipated in earthquakes, volcanoes, and mountain building. In 1931 Arthur Holmes* hypothesized that thermal convection was capable of driving mantle convection and continental drift. If a fluid is heated from below, or from within, and is cooled from above in the presence of a gravitational field, it becomes gravitationally unstable, and thermal convection can occur. The hot mantle rocks at depth are gravitationally unstable with respect to the colder, more dense rocks in the lithosphere. The result is thermal convection in which the colder rocks descend into the mantle and the hotter rocks ascend toward the surface. The ascent of mantle material at ocean ridges and the descent of the lithosphere into the mantle at ocean trenches are parts of this process. The earth's mantle is being heated by the decay of the radioactive isotopes uranium 235 (^{235}U), uranium 238 (^{238}U), thorium 232 (^{232}Th), and potassium 40 (^{40}K). The volumetric heating from these isotopes drives mantle convection.

During the 1960s independent observations supporting continental drift came from paleomagnetic studies. When a rock is formed, the earth's magnetic field at the time of formation can permanently magnetize the rock. Studies of the orientation of this field can be used to determine the movement of the rock relative to the earth's magnetic poles since the rock's formation. Rocks in a single surface plate that have not been deformed locally show the same position for the earth's magnetic poles. In 1956 Keith Runcorn** showed that rocks in North America and Europe gave different positions for the magnetic poles. He concluded that the differences were the result of continental drift between the two continents.

Paleomagnetic studies also showed that the earth's magnetic field has been subject to episodic reversals. Observations of the magnetic field over the oceans showed a regular striped pattern of *magnetic anomalies* (regions of magnetic field above and below the average field value) lying parallel to the ocean ridges. In 1963 Frederick Vine and Drummond Matthews* correlated the locations of the edges of the striped pattern of magnetic anomalies with the times of magnetic field reversals and were able to obtain quantitative values for the rate of seafloor spreading.

*"Radioactivity and earth movement, XVIII," Transactions of the Geological Society of Glasgow, vol. 18, pp. 559–606.

**"Paleomagnetic comparisons between Europe and North America," Proceedings of the Geological Association of Canada, vol. 8, pp. 77–85.

By the late 1960s the framework for a comprehensive understanding of the geological phenomena and processes of continental drift had been built. The basic hypothesis of plate tectonics was given by Jason Morgan in 1968.** The concept of a mosaic of rigid plates in relative motion with respect to one another was a natural consequence of thermal convection in the mantle. Almost all earthquakes, volcanoes, and mountain building can now be attributed to the interaction of lithospheric plates at their boundaries. Continental drift is an inherent part of plate tectonics. The continents are carried with the plates as they move about the surface of the earth.

Problem 1-1 If the area of the oceanic crust is 3.2×10^8 km^2 and new seafloor is now being created at the rate of 2.8 km^2 yr^{-1}, what is the mean age of the oceanic crust? Assume that the rate of seafloor creation has been constant in the past.

1-2 THE LITHOSPHERE

An essential feature of plate tectonics is that only the outer shell of the earth, the lithosphere, remains rigid during intervals of geologic time. Because of their low temperature, rocks in the lithosphere do not significantly deform on time scales of up to 10^9 years. The rock beneath the lithosphere is sufficiently hot so that solid-state creep can occur. This creep leads to a fluidlike behavior on geologic time scales. In response to forces, the rock beneath the lithosphere flows like a fluid.

*"Magnetic anomalies over oceanic ridges," *Nature*, vol. 199, p. 947.

**"Rises, trenches, great faults, and crustal blocks," *Journal of Geophysical Research*, vol. 73, pp. 1959–1982.

The lower boundary of the lithosphere is defined to be an isotherm (surface of constant temperature). A typical value is approximately 1600°K (~1300°C). Rocks lying above this isotherm are sufficiently cool to behave rigidly, whereas rocks lying below this isotherm are sufficiently hot so that they readily deform. Beneath the ocean basins the lithosphere has a thickness of about 100 km; beneath the continents the thickness is about twice this value. Because the thickness of the lithosphere is about 2 to 4% of the radius of the earth, the lithosphere is a thin shell. This shell is broken up into a number of plates that are in relative motion with respect to one another. The rigidity of the lithosphere ensures, however, that the interiors of the plates do not deform significantly.

The rigidity of the lithosphere allows the plates to transmit elastic stresses during geologic intervals. The plates act as stress guides. Stresses that are applied at the boundaries of a plate can be transmitted throughout the interior of the plate. The ability of the plates to transmit stress over large distances has important implications with regard to the driving mechanism of plate tectonics.

The rigidity of the lithosphere also allows it to bend when subjected to a load. An example is the load applied by a volcanic island. The load of the Hawaiian Islands causes the lithosphere to bend downward around the load, resulting in a region of deeper water around the islands. The elastic bending of the lithosphere under vertical loads can also explain the structure of ocean trenches and some sedimentary basins.

However, the entire lithosphere is not effective in transmitting elastic stresses. Only about the upper half of it is sufficiently rigid so that elastic stresses are not relaxed on time scales of 10^9 years. This fraction of the lithosphere is referred to as the *elastic lithosphere*. Solid-state creep processes relax stresses in the lower, hotter part of the lithosphere. However, this part of the lithosphere remains a coherent part of the plates. A detailed discussion of the difference between the thermal and elastic lithospheres is given in Section 7-10.

1-3 ACCRETING PLATE MARGINS

Lithospheric plates are created at ocean ridges. The two plates on either side of an ocean ridge move away from each other with near constant velocities of a few tens of millimeters per year. As the two plates diverge, hot mantle rock flows upward to fill the gap. The upwelling mantle rock cools by conductive heat loss to the surface. The cooling rock accretes to the base of the spreading plates, becoming part of them; the structure of an accreting plate margin is illustrated in Figure 1-4.

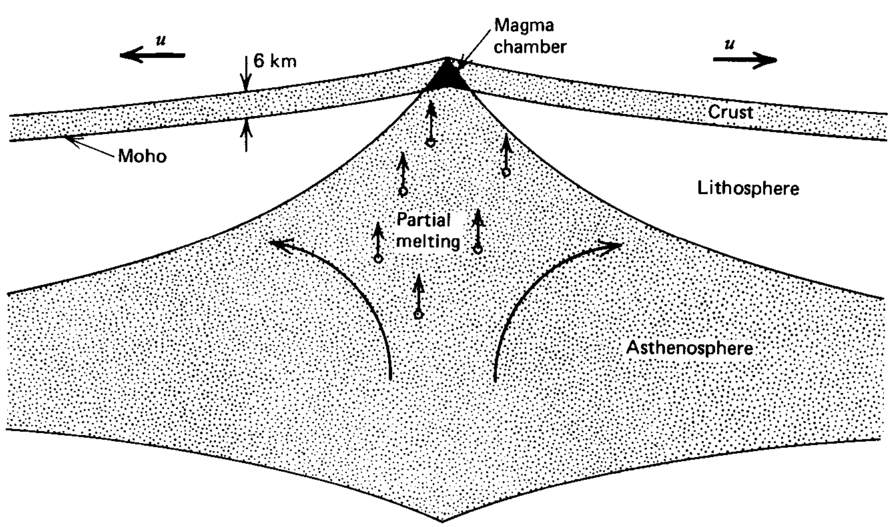

Figure 1-4 An accreting plate margin at an ocean ridge.

As the plates move away from the ocean ridge, they continue to cool and thicken. The elevation of the ocean ridge as a function of distance from the ridge crest can be explained in terms of the temperature distribution in the lithosphere. As the lithosphere cools, it becomes more dense; as a result it sinks downward into the underlying mantle rock. The topographic elevation of the ridge is due to the greater buoyancy of the thinner, hotter lithosphere near the axis of accretion at the ridge crest. The elevation of the ridge also provides a body force that causes the plates to move away from the ridge crest. A component of the gravitational body force on the elevated lithosphere drives the lithosphere away from the accretional boundary; it is one of the important forces driving the plates. The phenomenon is known as *gravitational sliding.*

Ocean ridges are the sites of a large fraction of the earth's volcanism. Because almost all the ridge system is under water, only a small part of this volcanism can be readily observed. The details of the volcanic processes at ocean ridges have been revealed by explorations using submersible vehicles (Figure 1-5). Ridge volcanism can also be seen in Iceland, where the oceanic crust is sufficiently thick so that the ridge crest rises above sea level. The volcanism at ocean ridges is caused by *pressure-release melting.* As the two adjacent plates move apart, hot mantle rock ascends to fill the gap. The temperature of the ascending rock is nearly constant, but its pressure decreases. The

Figure 1-5 Pillow lava near the crest of the East Pacific Rise off the Mexican coast photographed from the submersible Alvin. (R. D. Ballard, Woods Hole Oceanographic Institute.)

pressure p of rock in the mantle is given by the simple hydrostatic equation

$$p = \rho g y \quad (1\text{-}1)$$

where ρ is the density of the mantle rock, g is the acceleration of gravity, and y is the depth. The solidus temperature (the temperature at which the rock first melts) decreases with decreasing pressure. When the temperature of the ascending mantle rock equals the solidus temperature, melting occurs, as illustrated in Figure 1-6. The ascending mantle rock contains a low-melting-point, basaltic component. This component melts to form the oceanic crust.

Problem 1-2 At what depth will ascending mantle rock with a temperature of 1800°K melt if the equation for the solidus temperature T is

$$T(°K) = 1700 + 0.12p \text{ (MPa)}$$

Assume $\rho = 3300 \text{ kg m}^{-3}$, $g = 10 \text{ m s}^{-2}$.

The *magma* (melted rock) produced by partial melting beneath an ocean ridge is lighter than the residual mantle rock, and buoyancy forces drive it upward to the surface in the vicinity of the ridge crest. A large magma chamber is formed. Heat is lost to the seafloor, and this magma solidifies to form the oceanic crust. In some localities slices of oceanic crust and underlying mantle have been brought to the surface. These are known as *ophiolites*; they occur in such locations as Cyprus, Newfoundland, Yemen, and New Guinea. Field studies of ophiolites have provided a detailed understanding of the oceanic crust and underlying mantle. Typical oceanic crust is illustrated in Figure 1-7. The crust is divided into layers 1, 2, and 3, which were originally associated with different seismic velocities but which were subsequently identified compositionally. Layer 1 is composed of sediments that are deposited on the volcanic rocks of layers 2 and 3. The thickness of sediments increases with distance from the ridge crest; a typical thickness is 1 km. Layers 2 and 3 are composed of basaltic rocks of nearly uniform composition. A typical composition of an ocean basalt is given in Table 1-1. The basalt is composed primarily of two rock-forming minerals, plagioclase feldspar and pyroxene. The plagioclase feldspar is 50 to 85% anorthite

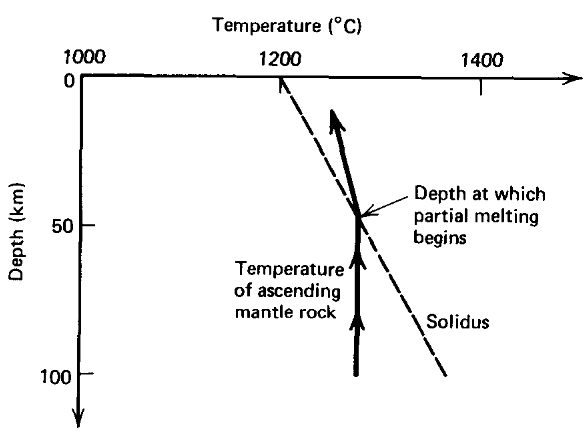

Figure 1-6 The process of pressure-release melting is illustrated. Melting occurs because the nearly isothermal ascending mantle rock encounters pressures low enough so that the associated solidus temperatures are below the rock temperatures.

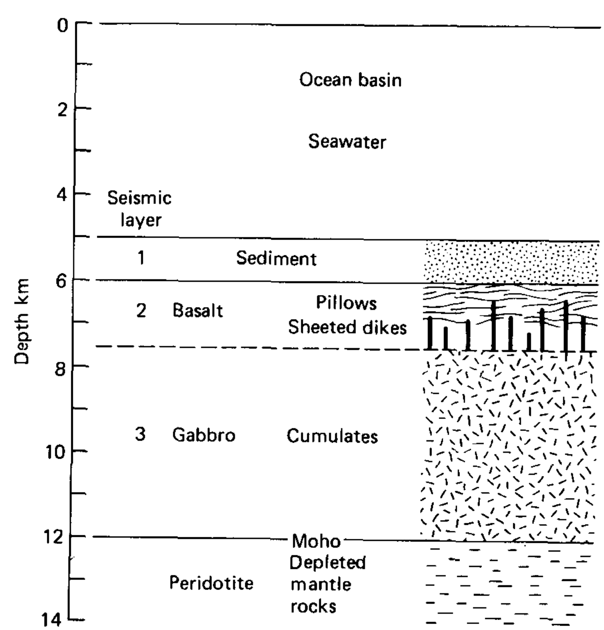

Figure 1-7 Typical structure of the oceanic crust, overlying ocean basin, and underlying depleted mantle rock.