2

Plate Tectonics

2.1 Introduction

During the 1960s there were a wide variety of studies on continental drift and its relationship to mantle convection. One of the major contributors was J. Tuzo Wilson. Wilson (1963a, b, 1965a, b) used a number of geophysical arguments to delineate the general movement of the ocean floor associated with seafloor spreading. He argued that the age progression of the Hawaiian Islands indicated movement of the Pacific plate. He showed that earthquakes on transform faults required seafloor spreading at ridge crests. During this same period other geophysicists outlined the general relations between continental drift and mantle convection (Orowan, 1964, 1965; Tozer, 1965a; Verhoogen, 1965). Turcotte and Oxburgh (1967) developed a boundary layer model for thermal convection and applied it to the mantle. According to this model, the oceanic lithosphere is associated with the cold upper thermal boundary layer of convection in the mantle; ocean ridges are associated with ascending convection in the mantle and ocean trenches are associated with the descending convection of the cold upper thermal boundary layer into the mantle. Despite these apparently convincing arguments, it was only with the advent of plate tectonics in the late 1960s that the concepts of continental drift and mantle convection became generally accepted.

Plate tectonics is a model in which the outer shell of the Earth is broken into a number of thin rigid plates that move with respect to one another. The relative velocities of the plates are of the order of a few tens of millimeters per year. Volcanism and tectonism are concentrated at plate boundaries. The basic hypothesis of plate tectonics was given by Morgan (1968); the kinematics of rigid plate motions were formulated by McKenzie and Parker (1967) and Le Pichon (1968). Plate boundaries intersect at triple junctions and the detailed evolution of these triple junctions was given by McKenzie and Morgan (1969). The concept of rigid plates with deformations primarily concentrated near plate boundaries provided a comprehensive understanding of the global distribution of earthquakes (Isacks et al., 1968).

The distribution of the major surface plates is given in Figure 2.1; ridge axes, subduction zones, and transform faults that make up plate boundaries are also shown. Global data used to define the plate tectonic model are shown in Figures 2.2–2.9. The distribution of global shallow and deep seismicity is shown in Figure 2.2, illustrating the concept of shallow seismicity defining plate boundaries. Figure 2.3 shows the distribution of ages of the ocean crust obtained from the pattern of magnetic anomalies on the seafloor. The distribution of crustal ages confirms that ridges are the source of ocean crust and also establishes the rates of seafloor spreading in plate tectonics. Figures 2.4–2.6 show geoid height variations – the topography of the equilibrium sea surface, which correlates closely with seafloor topography

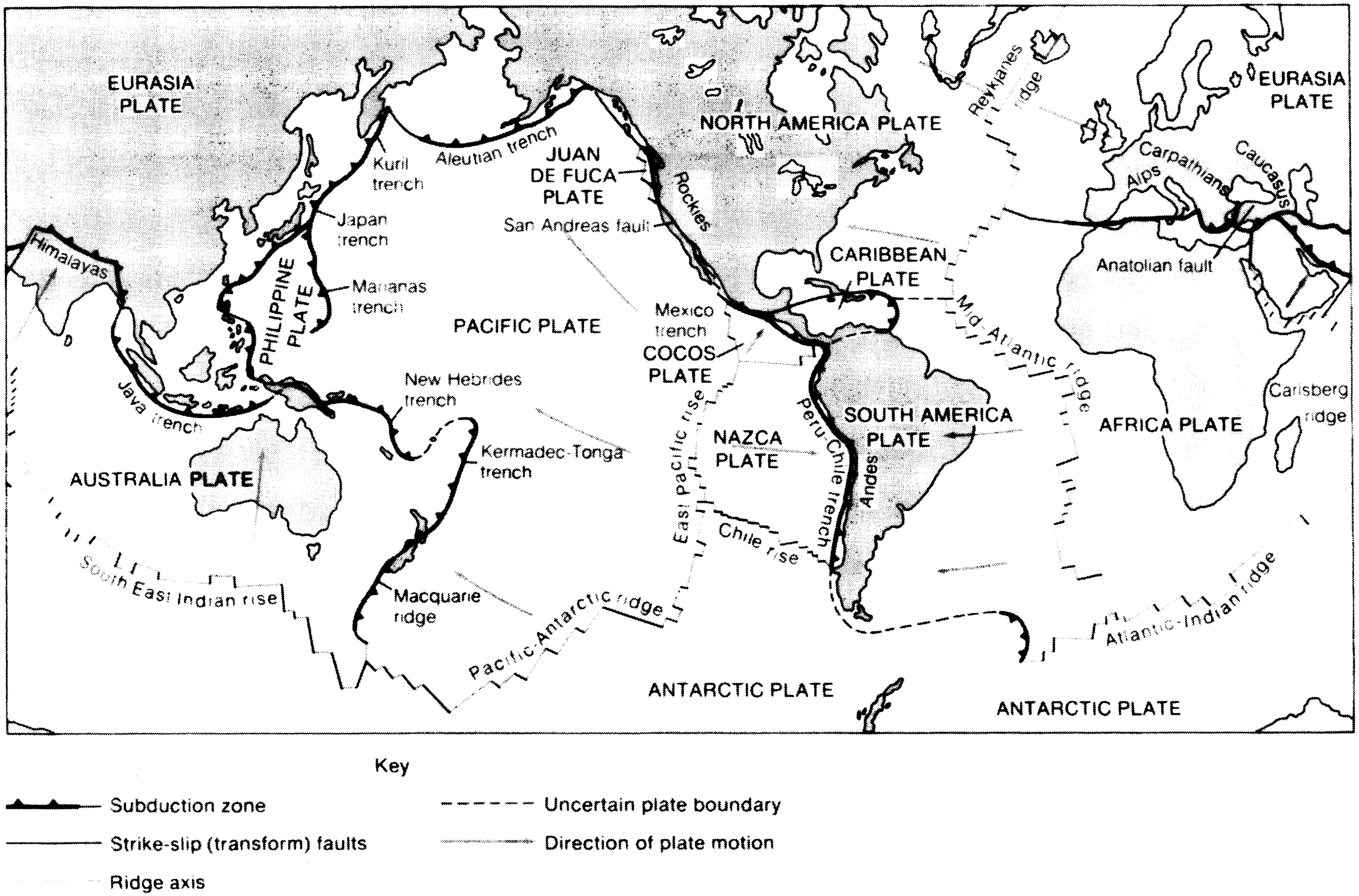

Figure 2.1. Distribution of the major surface plates. The ridge axes, subduction zones, and transform faults that make up the plate boundaries are shown. After Bolt (1993).

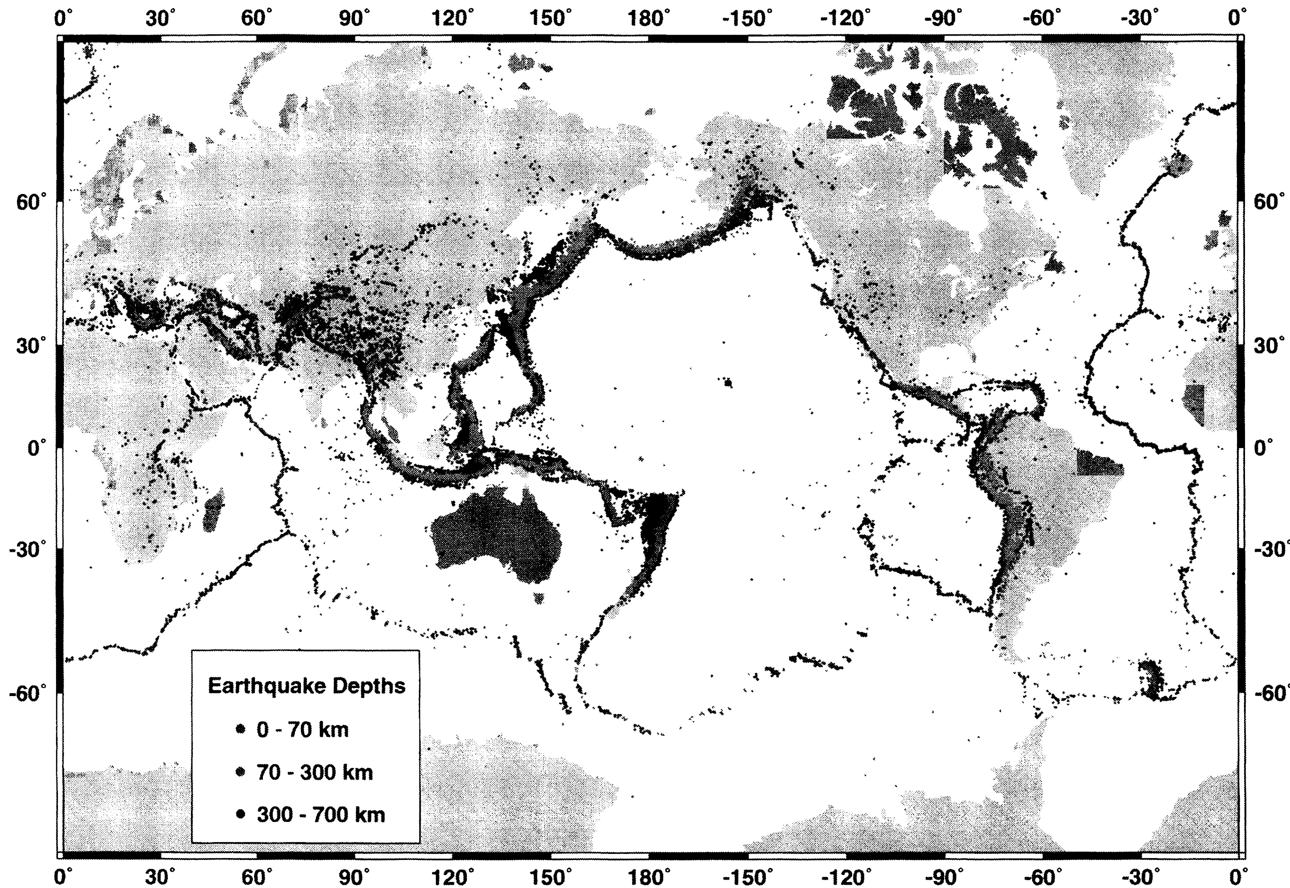

Figure 2.2. The global distribution of both shallow and deep seismicity for well-located earthquakes with magnitude > 5.1. The shallow seismicity closely delineates plate boundaries. Based on Engdahl et al. (1998).

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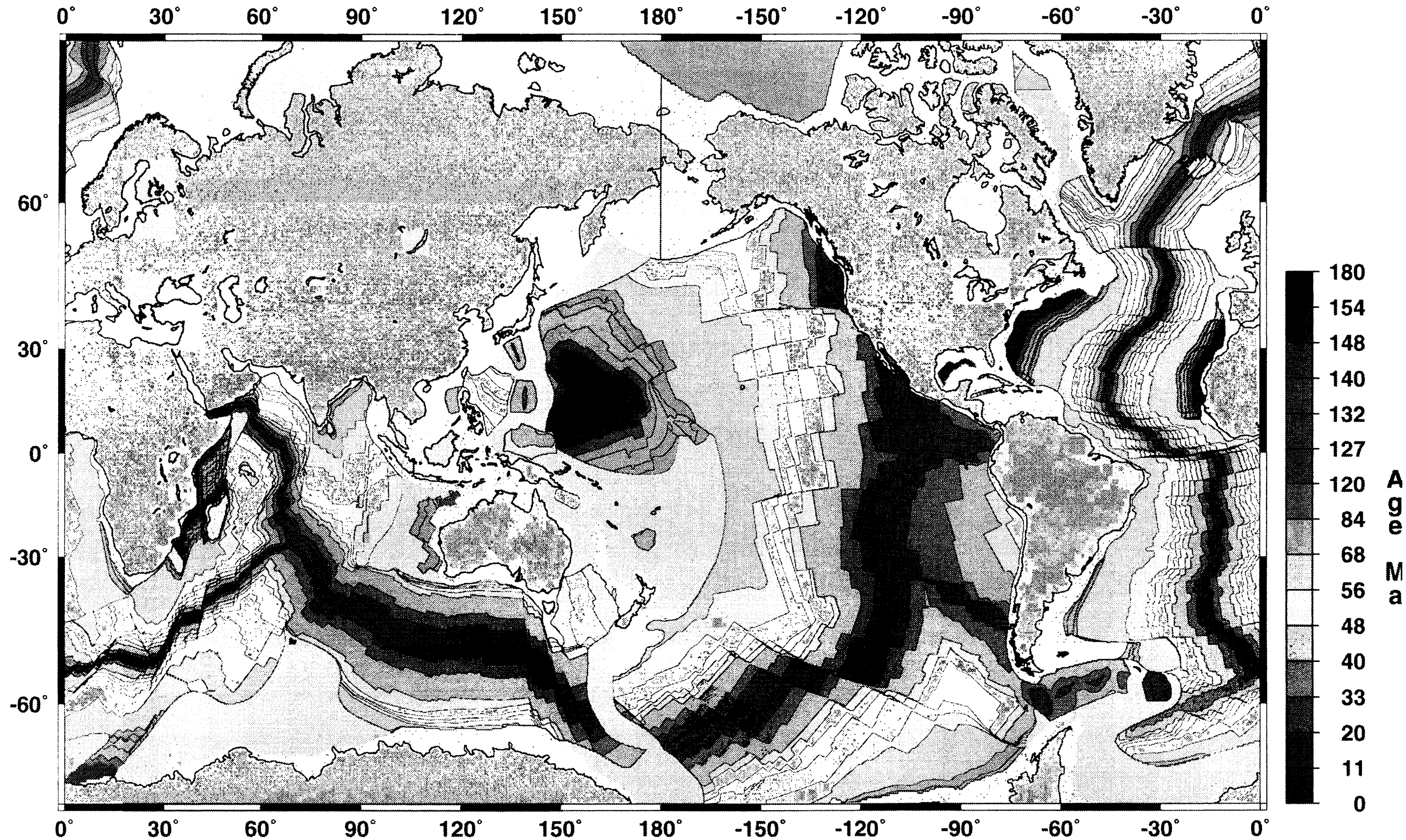

Figure 2.3. Age distribution of the oceanic crust as determined by magnetic anomalies on the seafloor. Based on Mueller et al. (1997).

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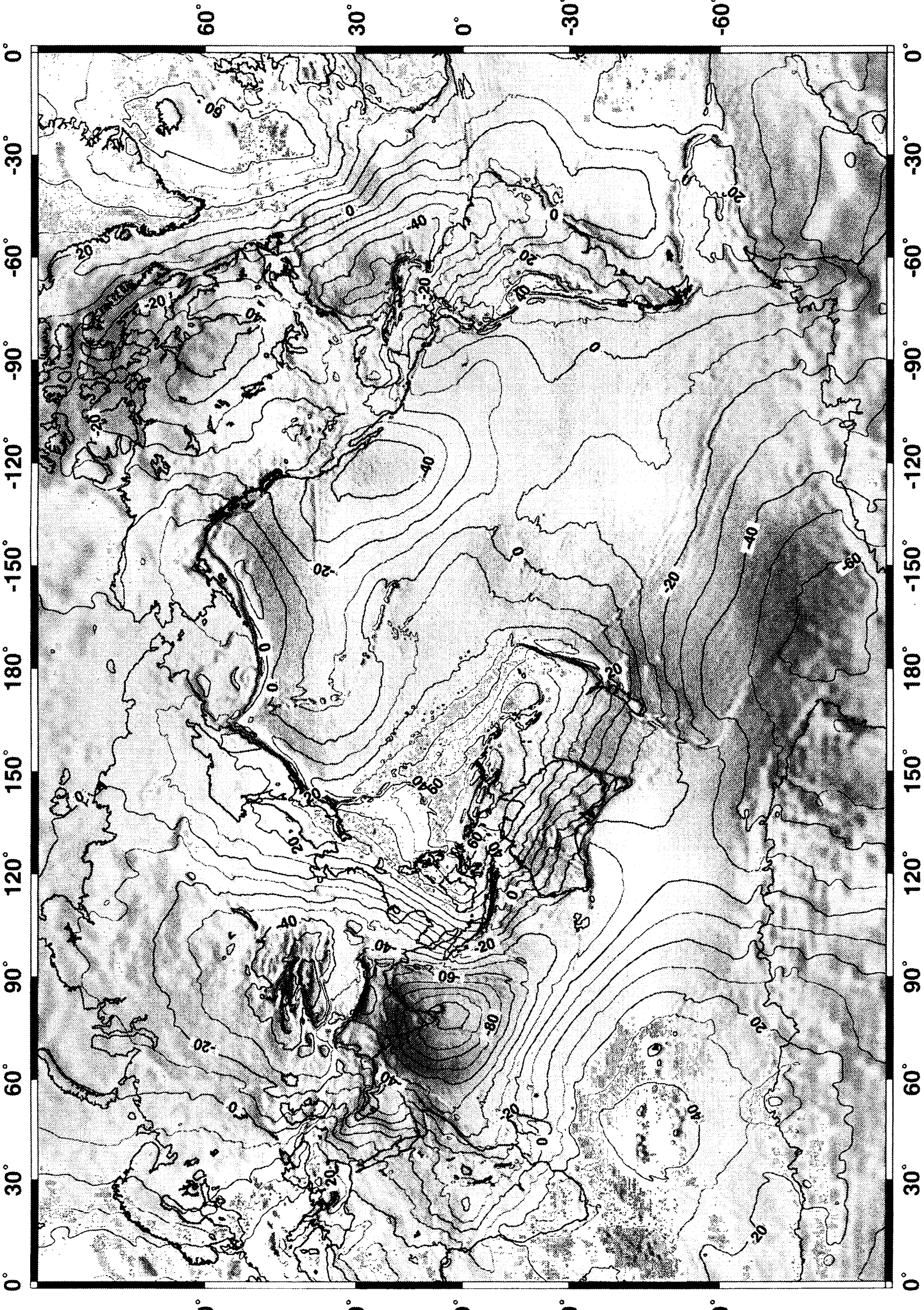

(a)

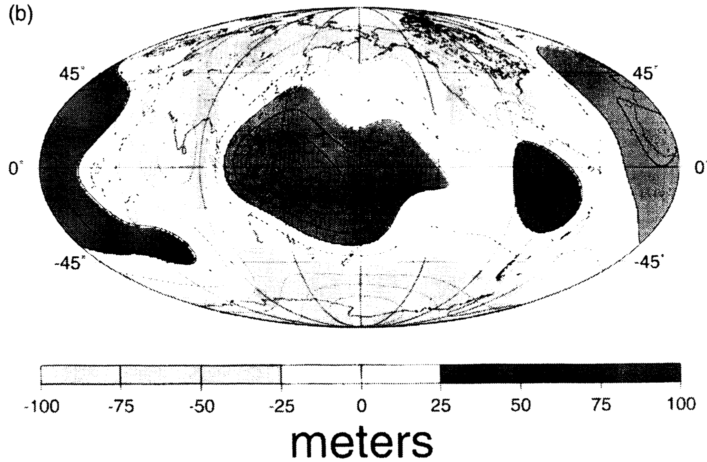

Figure 2.4. (a) Global geoid variations (after Lemoine et al., 1998) and (b) geoid variations complete to spherical harmonic degree 6 (after Ricard et al., 1993). (a) is model EGM96 with respect to the reference ellipsoid WG584. In (b), dotted contours denote negative geoid heights and the dashed contour separates areas of positive and negative geoid height.

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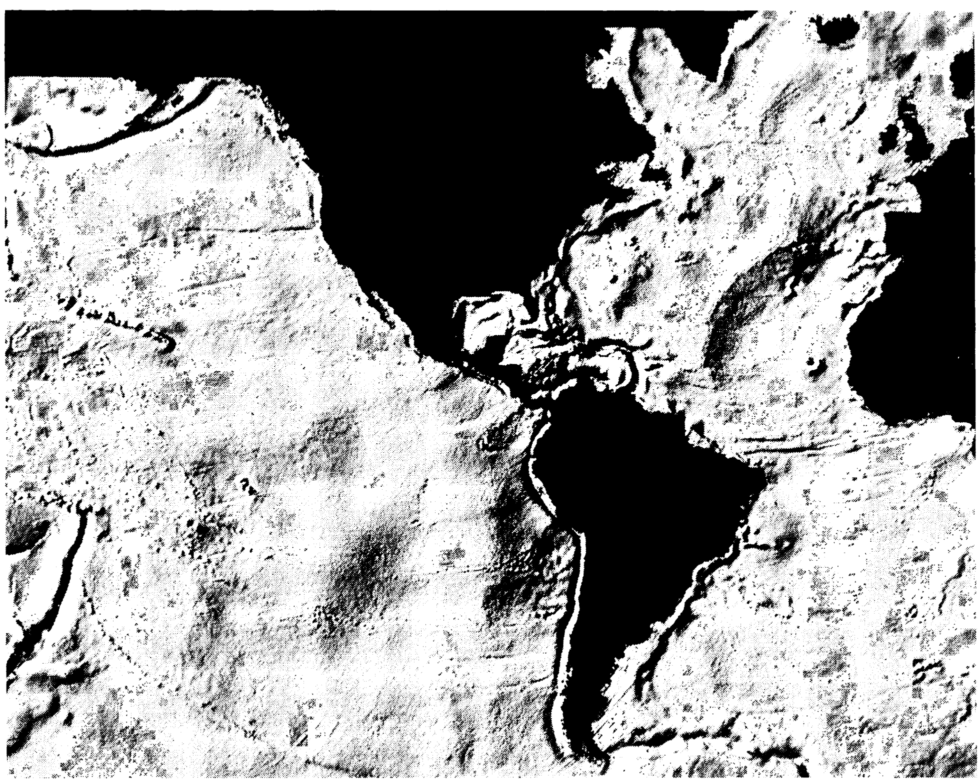

Figure 2.5. Geoid variations over the Atlantic and the eastern Pacific. The long-wavelength components of the global geoid shown in Figure 2.4b (to spherical harmonic degree 6) have been removed. After Marsh (1983).

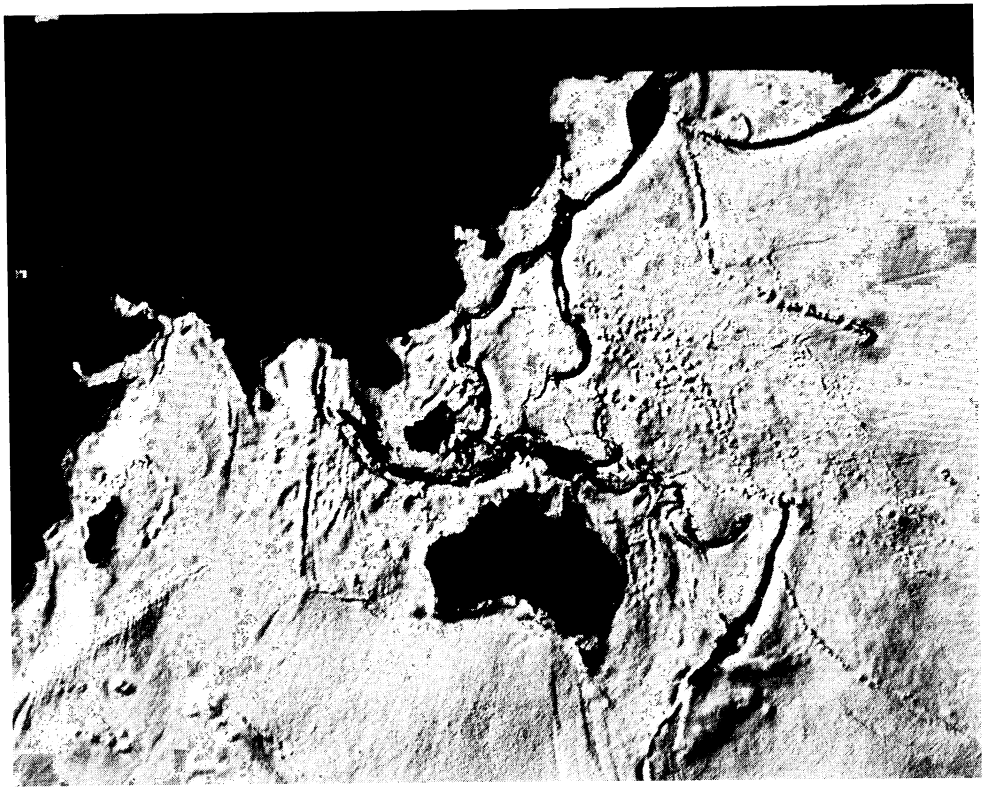

Figure 2.6. Western Pacific geoid. The long-wavelength components of the global geoid shown in Figure 2.4b (to spherical harmonic degree 6) have been removed. After Marsh (1983).

(Figure 2.7). The three types of structures used to define plate boundaries in Figure 2.1 – ridges, trenches, and transform faults – are evident in the geoid and the topography. Figure 2.8 shows the global pattern of heat flow, and Figure 2.9 gives the global locations of volcanoes. Volcanoes, like earthquakes, are strongly clustered at plate boundaries, mainly subduction zones. There are also numerous intraplate volcanoes, many at sites known as hot spots.

The essence of plate tectonics is as follows. The outer portion of the Earth, termed the lithosphere, is made up of relatively cool, stiff rocks and has an average thickness of about 100 km. The lithosphere is divided into a small number of mobile plates that are continuously being created and consumed at their edges. At ocean ridges, adjacent plates move apart 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 accretionary plate boundaries.

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. A cross-sectional view of the creation and consumption of a typical plate is illustrated in Figure 2.10. As the plates move away from ocean ridges, they cool and thicken and their density increases due to thermal

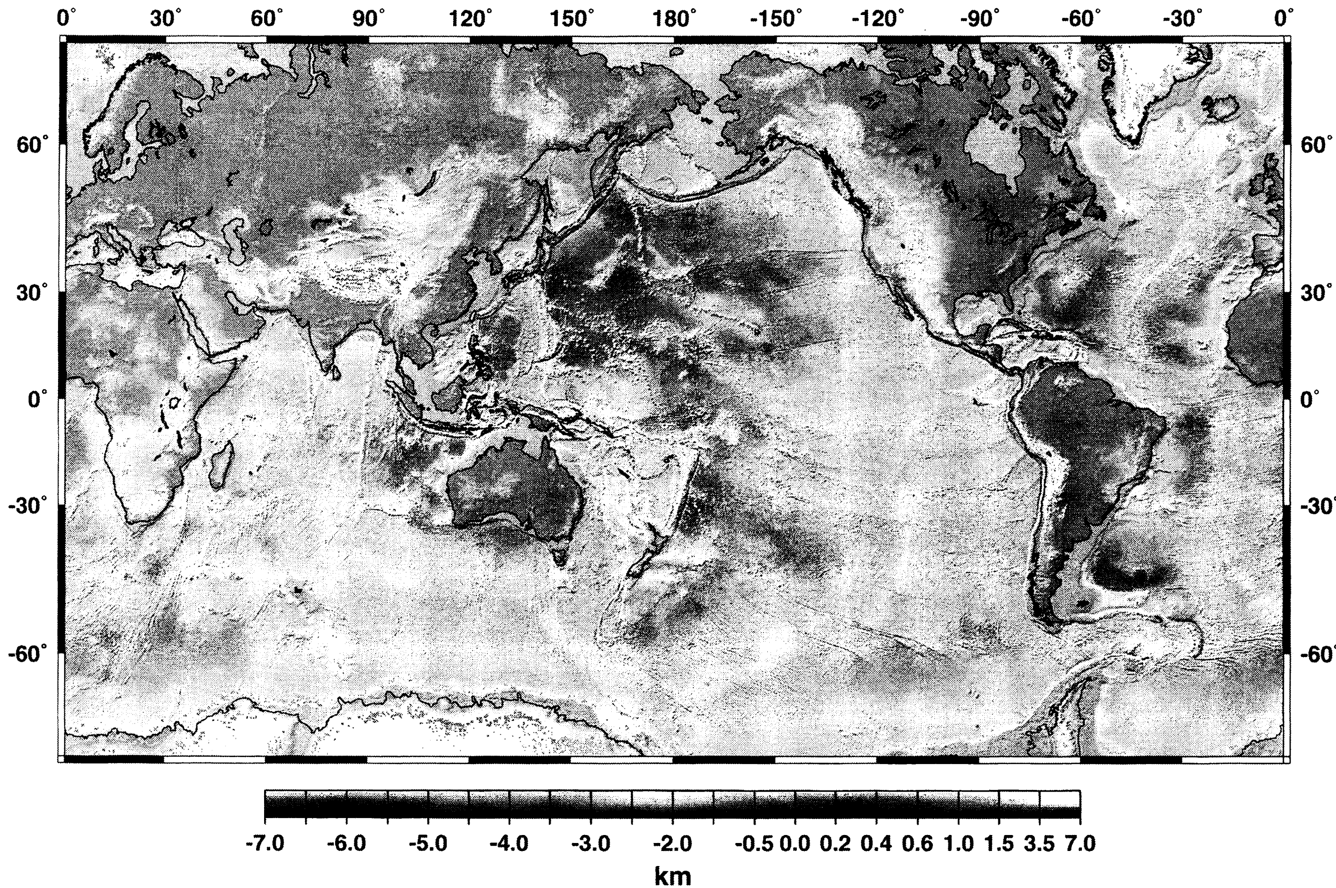

Figure 2.7. Global topography. The mountain range on the seafloor, the system of mid-ocean ridges, is a prominent feature of the Earth's topography. Based on Smith and Sandwell (1997).

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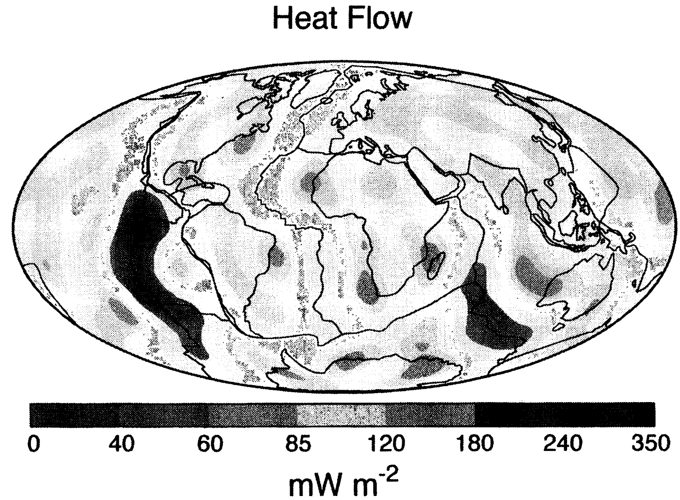

Figure 2.8. Pattern of global heat flux variations complete to spherical harmonic degree 12. After Pollack et al. (1993).

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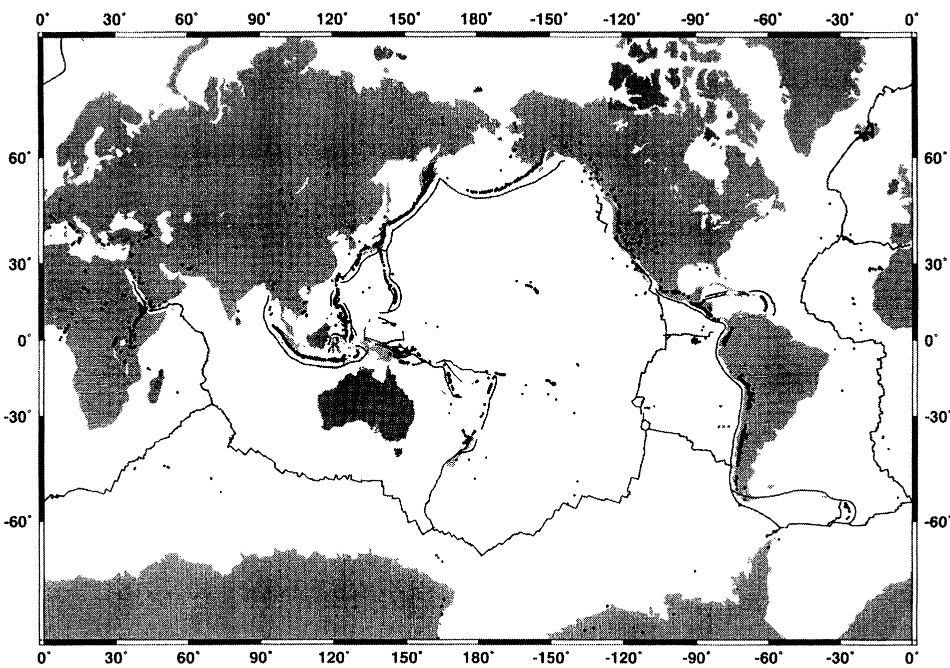


Figure 2.9. Global distribution of volcanoes active in the Quaternary.

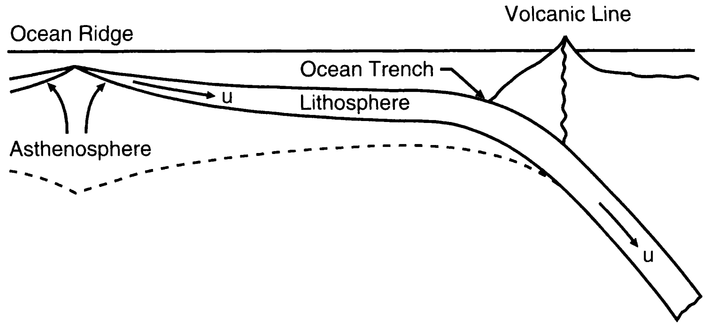


Figure 2.10. Accretion of a lithospheric plate at an ocean ridge (accretional plate margin) and its subduction at an ocean trench (subduction zone). The asthenosphere, which lies beneath the lithosphere, and the volcanic line above the subducting lithosphere are also shown. The plate migrates away from the ridge crest at the spreading velocity u. Since there can be relative motion between the ocean ridge and ocean trench, the velocity of subduction can, in general, be greater or less than u.

contraction. As a result, the lithosphere becomes gravitationally unstable with respect to the warmer asthenosphere beneath. At an ocean trench, the lithosphere bends and sinks into the interior of the Earth because of its negative buoyancy.

2.2 The Lithosphere

An essential feature of plate tectonics is that only the outer shell of the Earth, the lithosphere, remains rigid during long intervals of geologic time. Because of their low temperature, rocks in the lithosphere resist deformation on time scales of up to 10^9 yr. In contrast, the rock beneath the lithosphere is sufficiently hot that solid-state creep occurs. The lithosphere is composed of both mantle and crustal rocks. The oceanic lithosphere has an average thickness of 100 km with the uppermost 6–7 km being the oceanic crust. The oceanic lithosphere participates in the plate tectonic cycle. The continental lithosphere has a typical thickness of about 200 km (some authors argue that the thickness may be twice this value; further discussion will be given in Chapters 3 and 4). Typically, the upper 30 km of the continental lithosphere is continental crust. Because of the buoyancy of the continental crust, the continental lithosphere does not subduct, although it does participate in plate motions.

Because of their higher temperatures, rocks beneath the lithosphere can flow by subsolidus viscous creep. This region is called the asthenosphere. Because the silicic rocks of the continental crust are softer than both mantle rocks and the basaltic rocks of the oceanic crust, rocks within the continental crust, particularly within the lower crust, can also flow viscously while the mantle rocks below behave rigidly. The result is an intracrustal asthenosphere layer within the continental lithosphere.

Although the concept of a lithosphere is straightforward, there is in fact considerable confusion because the precise definition depends on the property being considered. We speak of the mechanical lithosphere, the thermal lithosphere, and the elastic lithosphere. Here each is considered in turn.

(1) *Mechanical lithosphere.* The mechanical lithosphere is defined to be those rocks that remain a coherent part of the plates on geological time scales. A typical definition would be rocks that cannot be deformed more than 1% in 10^8 yr at typical mantle stress levels

(say 1 MPa). The deformation of the mantle is determined by its viscosity, and its viscosity is, in turn, determined by its temperature. Thus the base of the mechanical lithosphere is prescribed by an isotherm, typically 1,400 K. Rocks shallower than the level of this isotherm are sufficiently cool to behave rigidly, whereas rocks lying deeper than this isotherm are sufficiently hot to deform and flow viscously in response to long-term forces. The term "tectonic plate" is most closely related to this definition of the lithosphere.

(2) *Thermal lithosphere*. We will consider the thermal structure of both the oceanic and continental lithospheres in Chapter 4. The oceanic lithosphere is taken to be the upper thermal boundary layer of mantle convection. Since the thermal boundary layer has a continuous variation in temperature, the definition of its thickness is arbitrary. If the temperature difference across the oceanic lithosphere (the thermal boundary layer) is $T_m - T_s$, where T_s is the surface temperature and T_m is the mantle temperature beneath the boundary layer, a typical definition of the base of the thermal boundary layer is the depth at which the temperature T is $T_s + 0.9(T_m - T_s)$. If this temperature is equal to the rheological temperature determining the base of the mechanical lithosphere, then the mechanical and thermal lithospheres are the same. In the remainder of this book we will assume that this is the case, and we will refer to both the mechanical and thermal lithospheres as the lithosphere.

(3) *Elastic lithosphere*. The rigidity of the lithosphere also allows it to flex 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 moat, i.e., 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 that elastic stresses are not relaxed on time scales of 10^9 yr. 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. This lower part of the lithosphere, however, remains a coherent part of the plates.

The strength of the lithosphere allows the plates to transmit elastic stresses over geologic time 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 is a key factor in driving tectonic plates.

2.3 Accretional Plate Margins (Ocean Ridges)

Lithospheric plates are created at ocean ridges (Figures 2.1, 2.5, 2.7, and 2.10). The two plates on either side of an ocean ridge move away from each other with nearly steady 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 2.11.

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, shown in the images of the geoid in Figures 2.5 and 2.6 and in the topography of Figure 2.7, can be explained in terms of the temperature distribution in the lithosphere. As the lithosphere cools, it contracts thermally and its upper surface – the ocean floor – sinks relative to the ocean surface. The topographic elevation of the ridge is due to the lower-density, thinner, and hotter lithosphere near the axis of accretion at the ridge crest. The elevation of

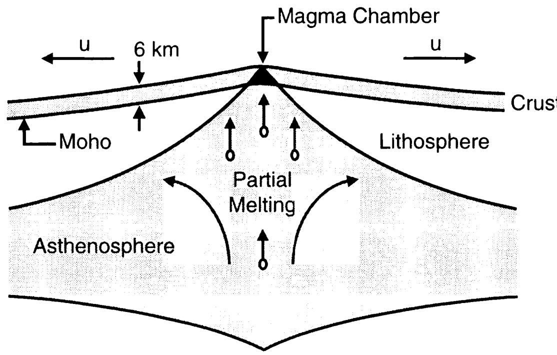


Figure 2.11. Structure at and beneath an accreting plate margin (ocean ridge). Hot flowing mantle rock (asthenosphere) ascends beneath the ridge axis. Pressure-release melting occurs and the resulting magma migrates upward to form the axial magma chamber. The basaltic rocks in this magma chamber solidify to form the 6 km thick ocean crust. Heat loss to the seafloor cools and thickens the oceanic lithosphere and hot asthenospheric rock accretes to it.

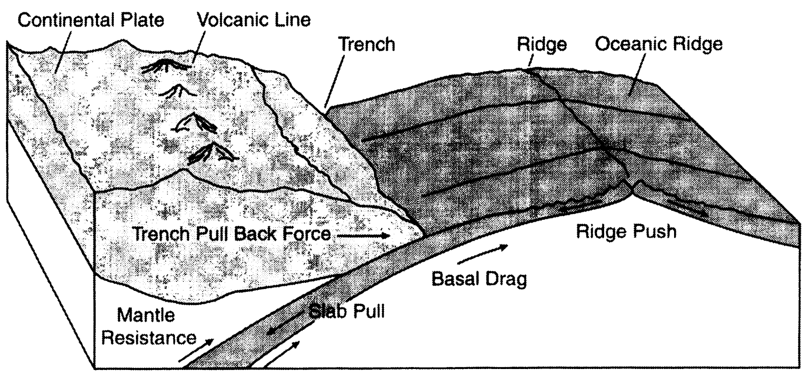


Figure 2.12. Illustration of the major forces acting on the plates.

the ridge also exerts a gravitational body force that drives the lithosphere away from the accretional boundary; it is one of the important forces driving the plates and is known as gravitational sliding or ridge push (Figure 2.12). Conductive cooling of the lithosphere also causes a decrease in the geothermal gradient, which is evident in the pattern of global heat flow (Figure 2.8); heat flow is highest at the ridges and decreases with increasing plate age.

The volume occupied by the ocean ridges displaces seawater. Rates of seafloor spreading vary in time. When rates of seafloor spreading are high, ridge volume is high, seawater is displaced, and the result is an increase in the global sea level. Variations in rates of seafloor spreading are the primary cause for changes in sea level on geological time scales (Hays and Pitman, 1973; Turcotte and Burke, 1978; Schubert and Reymer, 1985). During the Cretaceous ($\approx$80 Ma) the rate of seafloor spreading was about 30% greater than at present, sea level was about 200 m higher than today, and a substantial fraction of the continental interiors was covered by shallow seas.

Ocean ridges generate a large fraction of the Earth's volcanism. Because almost all the ridge system is below sea level, only a small part of this volcanism can be readily observed. Ridge volcanism can be seen in Iceland, where the oceanic crust is sufficiently thick that the ridge crest rises above sea level. The volcanism at ocean ridges is caused by pressure-release melting. The diverging plates induce an upwelling in the mantle. The temperature of the ascending rock decreases slowly with decreasing pressure along an adiabat. The solidus temperature for melting decreases with decreasing pressure at a much faster rate. When the temperature of the ascending mantle rock equals the solidus temperature, melting begins. The ascending mantle rock contains a low-melting-point basaltic component; this component melts first to form the oceanic crust (Figure 2.11).

2.4 Transform Faults

One of the striking features of accretional plate margins is the orthogonal system of ridge segments and transform faults. The ridge segments lie nearly perpendicular to the spreading

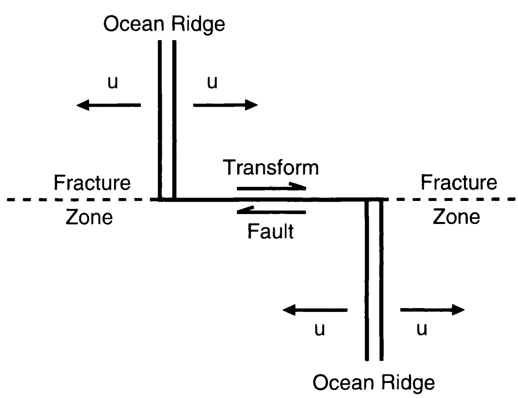


Figure 2.13. Segments of an ocean ridge offset by a transform fault. The fracture zones are extensions of the transform faults into the adjacent plates.

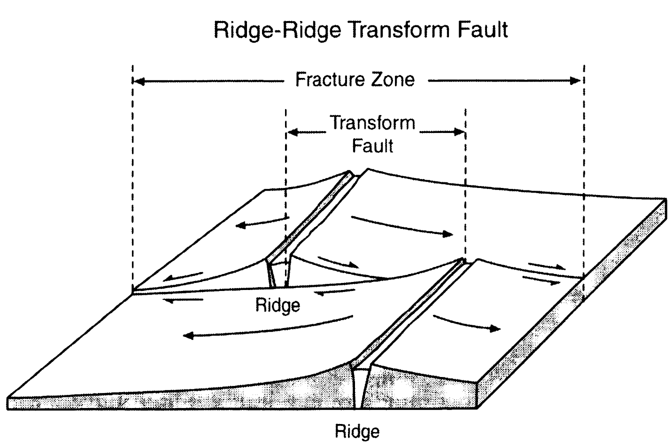


Figure 2.14. Sketch of a ridge–ridge transform fault showing exaggerated differential vertical subsidence across the fault.

direction, whereas the transform faults lie parallel to the spreading direction. This structure is illustrated in Figures 2.1, 2.5, 2.7, 2.13, and 2.14. The relative velocity across a transform fault is twice the spreading velocity. This relative motion results in seismicity on the transform fault between the adjacent ridge segments.

There is also differential vertical motion on transform faults (Figure 2.14). As the seafloor spreads away from a ridge crest, it also subsides. Since the adjacent points on each side of a transform fault usually lie at different distances from the ridge crest where the crust was formed, the rates of subsidence on the two sides differ. The extensions of the transform faults into the adjacent plates are known as fracture zones. Fracture zones are often deep valleys in the seafloor. Transform faults and fracture zones are prominently visible in Figures 2.5, 2.6, and 2.7.

Question 2.1: *Why do accretional plate margins develop the orthogonal ridge segment–transform fault geometry?*

Ocean ridges do not form obliquely to the direction of seafloor spreading. Instead they form the orthogonal ridge–transform system described above. This orthogonal system has been reproduced in laboratory experiments using freezing wax (Oldenburg and Brune, 1972, 1975). Despite the ability to reproduce the orthogonal pattern in the laboratory, the physical reason for the orthogonal pattern remains unclear. A number of authors have suggested that the pattern is associated with the thermal stresses that develop in the cooling oceanic lithosphere (Collette, 1974; Turcotte, 1974; Parmentier and Haxby, 1986; Sandwell, 1986).

A transform fault that connects two segments of an ocean ridge is known as a ridge–ridge transform (Figures 2.13 and 2.14). Transform faults can also connect two segments of an ocean trench or a segment of a ridge with a segment of a trench (Figure 2.15). In some cases one end of a transform fault terminates in a triple junction of three surface plates. An example is the San Andreas fault in California, which accommodates lateral sliding between the Pacific and North American plates (Figure 2.16).

A variety of other complex geometrical patterns are associated with accretional plate margins. In some cases ridge jumps occur and ridge segments propagate (Hey et al., 1980). If the direction of seafloor spreading changes due to plate interactions, the accretional margin can break up into a number of microplates until a new orthogonal pattern of ridges and transforms is established.

2.5 Subduction

As the oceanic lithosphere moves away from an ocean ridge, it cools, thickens, and becomes more dense because of thermal contraction. Even though the basaltic rocks of the oceanic crust are lighter than the underlying mantle rocks, the colder subcrustal rocks in the lithosphere become sufficiently dense to make old oceanic lithosphere heavy enough to be gravitationally unstable with respect to the hot mantle rocks beneath the lithosphere. As a result of this gravitational instability, the oceanic lithosphere founders and begins to sink into the interior of the Earth, creating the ocean trenches shown in Figures 2.5 and 2.6.

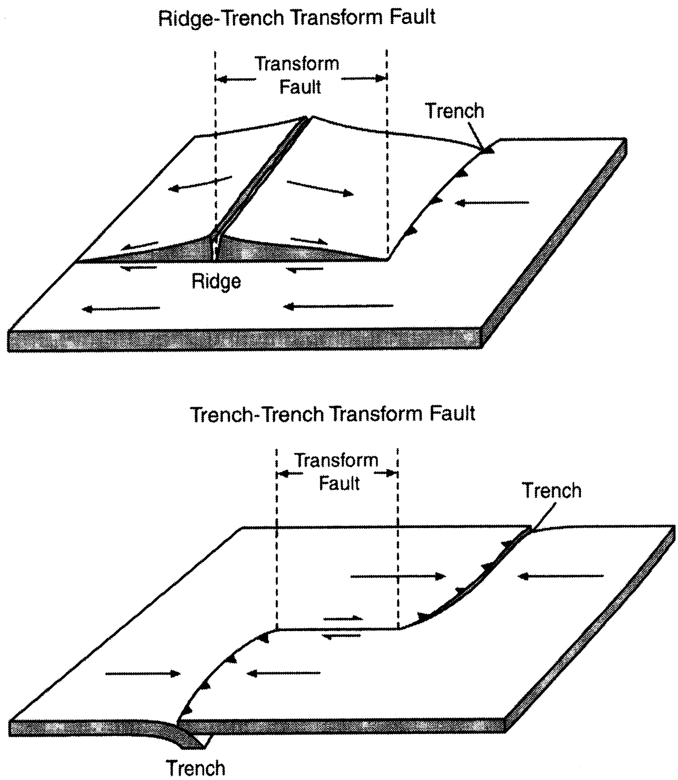


Figure 2.15. Sketch of ridge–trench and trench–trench transform faults.

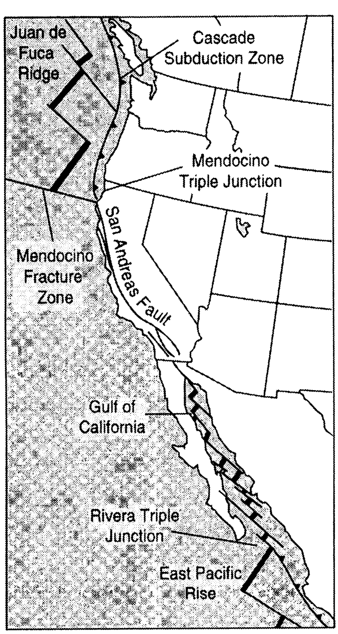


Figure 2.16. The San Andreas transform fault system accommodates lateral sliding between the Pacific and North American plates. The northern terminus of the San Andreas fault is the Mendocino triple junction at the intersection of the Juan de Fuca, Pacific, and North American plates.

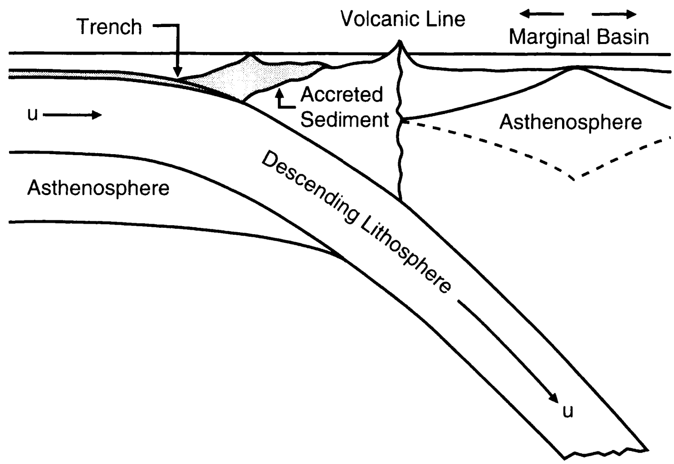


Figure 2.17. Illustration of the subduction of oceanic lithosphere at an ocean trench. The line of volcanic edifices associated with most subduction zones is shown. A substantial fraction of the sediments that coat the basaltic oceanic crust is scraped off during subduction to form an accretionary prism of sediments. Back-arc spreading forms a marginal basin behind some subduction zones.

The subduction of the oceanic lithosphere at an ocean trench is illustrated schematically in Figure 2.17. A chain of volcanoes which lie parallel to the ocean trench is generally associated with subduction. A substantial fraction of the sediments that coat the oceanic crust is scraped off during subduction to form an accretionary prism of sediments (von Huene and Scholl, 1991). In some cases back-arc seafloor spreading forms a marginal basin behind the subduction zone.

The excess density of the rocks of the descending lithosphere results in a downward buoyancy force. Because the lithosphere behaves elastically, it can transmit stresses, i.e., it can act as a stress guide. A portion of the negative buoyancy force acting on the descending plate is transmitted to the surface plate, which is pulled toward the ocean trench. This is slab pull, one of the important forces driving plate tectonics (Figure 2.12).

Ocean trenches are the sites of most of the largest earthquakes. At depths of less than about 55 km, the earthquakes occur on the dipping fault plane that separates the descending lithosphere from the overlying lithosphere (Ruff, 1996). Below about 55 km depth, the earthquakes probably occur within the subducting lithosphere (Comte et al., 1999). Earthquakes at ocean trenches can occur to depths of 660 km or more, depending on variations in the depth of the 660 km seismic discontinuity. Where the mantle is expected to be warmer, earthquakes do not extend as deep (Green and Houston, 1995; Kirby et al., 1996) in accordance with the expected upwarping of the 660 km seismic discontinuity (see Chapters 4, 9, 10, and 13). This seismogenic region, known as the Wadati–Benioff zone (Wadati, 1928, 1934/35; Benioff, 1949; Utsu, 1971), delineates the approximate structure of the descending plate. An early example of the geometry of the Wadati–Benioff zone at two locations along the Tonga arc is shown in Figure 2.18. The projection of the Wadati–Benioff zone onto the surface of the Earth is shown by the systematic horizontal trends in the locations of earthquakes of different depths on the global seismicity map of Figure 2.2. The shapes of the upper boundaries of several descending lithospheres are given in Figure 2.19, on the assumption that the earthquakes in the Wadati–Benioff zone lie on or near the top of the descending slab. The positions of the trenches and the volcanic

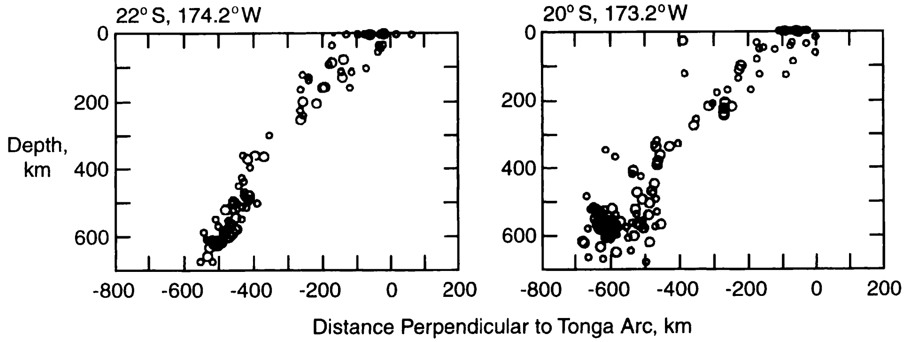


Figure 2.18. Earthquake foci beneath the Tonga arc at two sections oriented perpendicular to the arc. The geographic position corresponds to zero distance on the abscissa. Larger symbols represent more accurate hypocenter locations. The earthquakes were recorded by seismograph stations between 1959 and 1962. Hypocenters are projected from distances ±125 km of each line. The earthquake hypocenters delineate a nearly linear dipping structure, the Wadati–Benioff zone. After Sykes (1966).

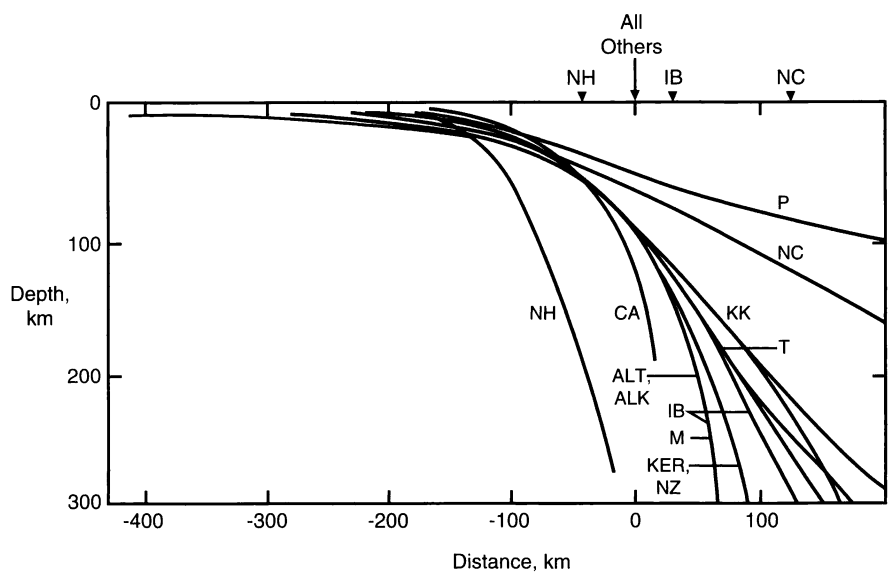


Figure 2.19. The shapes of the upper boundaries of descending lithospheric slabs at several oceanic trenches based on the distributions of earthquakes. The names of the trenches are abbreviated for clarity (NH = New Hebrides, CA = Central America, ALT = Aleutian, ALK = Alaska, M = Mariana, IB = Izu Bonin, KER = Kermadec, NZ = New Zealand, T = Tonga, KK = Kurile-Kamchatka, NC = North Chile, P = Peru). The locations of the volcanic lines are shown by solid triangles (Isacks and Barazangi, 1977); all except NH, IB, and NC lie at a common point (all others).

lines are also shown. Many subducted lithospheres have an angle of dip near 45°. In the New Hebrides the dip is significantly larger, and in Peru and North Chile the angle of dip is smaller.

Along localized segments of some subduction zones, earthquakes at depths of 70–150 km are concentrated on two parallel dipping planes vertically separated by 20–40 km

(Engdahl and Scholz, 1977; Hasegawa et al., 1978a, b; Kawakatsu, 1986; Abers, 1992, 1996; Gorbatov et al., 1994; Kao and Chen, 1994, 1995; Comte et al., 1999) (Figure 2.20). The upper plane of these double seismic zones seems to lie just below the top of the descending slab; the lower plane, therefore, must lie within the slab (Abers, 1996; Comte et al., 1999).

2.5.1 Rheology of Subduction

Question 2.2: *What is the rheology of the lithosphere at a subduction zone?*

The mechanical behavior of the lithosphere at subduction zones has received considerable attention. Two extreme modes of lithospheric deformation that might be associated with subduction are flexure and rupture. Flexure appears to be the best approximation. The oceanic lithosphere bends continuously and maintains its structural integrity as it passes through the subduction zone and creates the large geoid anomalies seen at the trenches in Figures 2.5 and 2.6. Studies of elastic bending at subduction zones are in good agreement with the morphology of some subduction zones seaward of the trench axis (Caldwell et al., 1976; Levitt and Sandwell, 1995). However, there are clearly significant deviations from a simple elastic rheology. Some trenches exhibit a sharp "hinge" near the trench axis; this has been attributed to an elastic–perfectly plastic rheology (McAdoo et al., 1978). Extensional shallow seismicity is generally observed on the forebulge seaward of ocean trenches. Thus the shallow bending lithosphere is undergoing brittle failure, but this failure does not propagate through the lithosphere and it appears to have little effect on the general flexural behavior.

The paired belts of deep seismicity in some subducting slabs (Figure 2.20) provide information on the rheology of these slabs. The upper seismic zone near the upper boundary of the descending lithosphere exhibits down-dip compressional focal mechanisms. The lower seismic zone near the center of the descending lithosphere exhibits down-dip extensional focal mechanisms. These double seismic zones are attributed to the "unbending," i.e., straightening out, of the descending lithosphere (Samowitz and Forsyth, 1981; Kawakatsu, 1986). The double seismic zones are further evidence of the rigidity of the subducted lithosphere. They also indicate that forces on the subducted lithosphere are straightening it out so it usually descends at nearly 45°.

An alternative explanation of the bending of the lithosphere as it approaches a subduction zone is that the bending is a viscous effect (De Bremaecker, 1977; McKenzie, 1977a; Melosh and Raefsky, 1980). Viscous deformation can produce the same morphology of flexure as an elastic rheology so that studies of flexure at trenches cannot differentiate between the two approaches. However, viscous flexure relaxes at long times. The fact that lithospheric flexure is observed in sedimentary basins with ages greater than 10^8 yr (compared with 10^6 yr for subduction) is evidence that the concept of a viscous rheology for lithospheric flexure is inappropriate (Turcotte, 1979). Nevertheless, application of a viscous rheology to the lithosphere may be appropriate for investigating other aspects of the subduction process (Zhang et al., 1985; Vassiliou and Hager, 1988; Zhong and Gurnis, 1994a; Gurnis et al., 1996).

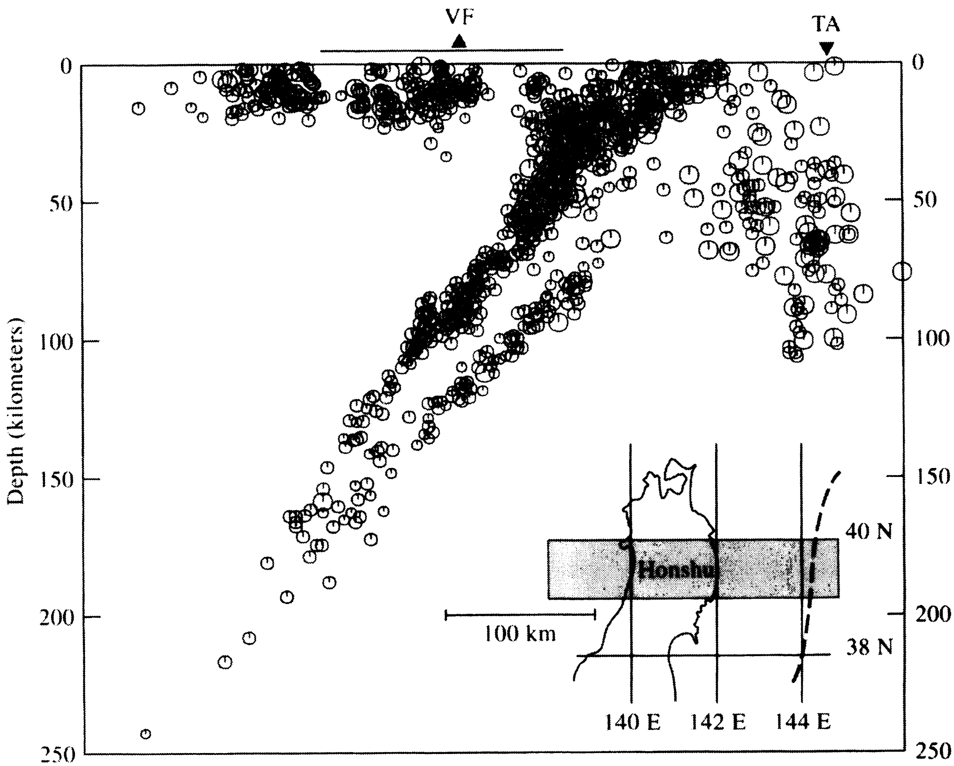


Figure 2.20. Double Benioff zone marking subduction at the Japan arc. Circles are foci of earthquakes recorded in 1975 and 1976. VF – volcanic front, TA – Japan Trench axis. After Hasegawa et al. (1978b). Redrawn from Bolt (1993).

2.5.2 *Dip of Subduction Zones*

Question 2.3: *What determines the subduction dip angle?*

Since the gravitational body force on the subducted lithosphere is downward, it would be expected that the subduction dip angle would tend toward 90°. In fact, as shown in Figure 2.19, the typical dip angle for a subduction zone is near 45°. One explanation is that the oceanic lithosphere is "foundering" and the trench is migrating oceanward. In this case the dip angle is determined by the flow kinematics (Hager and O'Connell, 1981). While this explanation is satisfactory in some cases, it has not been established that all slab dip angles can be explained by the kinematics of mantle flows.

An alternative explanation has been given by Stevenson and Turner (1977), Tovish et al. (1978), and Yokokura (1981). These authors argue that the subducted slab is supported by the induced flow above the slab. The descending lithosphere induces a corner flow in the mantle wedge above it and the pressure forces associated with this corner flow result in a dip angle near 45°.

2.5.3 *Fate of Descending Slabs*

Question 2.4: What is the fate of descending slabs?

One of the key questions in mantle convection is the fate of the descending slab. Earthquakes on the Wadati–Benioff zone terminate near a depth of about 660 km, but termination of seismicity does not imply cessation of slab descent. As will be discussed in the next chapter, 660 km is near the depth of a major seismic discontinuity associated with the solid–solid phase change from spinel to perovskite and magnesiowüstite; this phase change could act as a barrier to the descending lithosphere. In some cases seismic activity spreads out at this depth, and in some cases it does not. Studies of seismic focal mechanisms in the Wadati–Benioff zone give extensional stresses in the upper part of the zone and compressional stresses in the lower part of the zone (Isacks and Molnar, 1971).

Another major mantle phase transition at a depth of about 410 km is associated with the phase change from olivine to β spinel. Theoretical studies indicate that this shallower exothermic phase change will enhance convection (Turcotte and Schubert, 1971). However, similar studies of the effect on convection of the deeper endothermic phase change from spinel to perovskite and magnesiowüstite show that it could inhibit flow through 660 km depth, particularly if there is also a significant stabilizing compositional change at this depth (Schubert et al., 1975; Christensen and Yuen, 1984). The effects of major solid–solid phase transitions on convection in the mantle will be discussed in detail in Chapters 4, 9, and 10.

One of the principal goals of mantle seismic tomography has been to determine the fate of subducted slabs. Numerous seismic investigations of deep earthquakes and mantle structure around subduction zones have sought to resolve the maximum depth subducted lithosphere can be traced into the mantle. There is still much disagreement on this issue, perhaps because there is, in fact, no single depth where all slab penetration terminates. Indeed, surveys of all the pertinent seismic evidence (Lay, 1994a, b, c) come to the conclusion that some slabs penetrate through the transition zone into the lower mantle, while others do not.

An additional question about the fate of descending slabs is the maximum depth of penetration of slabs that enter the lower mantle.

Question 2.5: Do slabs that cross 660 km depth sink all the way to the core–mantle boundary or do they come to rest at some shallower depth?

Seismic tomographic evidence that at least some slabs descend all the way to the bottom of the mantle will be presented in Chapter 3.

2.5.4 *Why Are Island Arcs Arcs?*

Question 2.6: Why do subduction zones have arcuate structures?

One of the striking features of subduction zones is their arcuate structure in map view or planform. Subduction zones are made up of a sequence of arc structures with a clear planform curvature; this is the origin of the term "island arc." A good example is the Aleutian arc, shown in Figures 2.1 and 2.5. Just as accretionary margins are characterized by their orthogonal ridge–transform geometry, subduction zones are characterized by their arc configuration.

Frank (1968a) proposed a simple model for the curvature of island arcs based on a ping-pong ball analogy. If an indentation is made on a ping-pong ball, there is a simple analytical relation between the angle of dip and the radius of the indentation. Frank proposed that this relation also could be used to relate the dip angle of the subducted lithosphere to the planform radius of curvature of the island arc. The assumption was that the rigidity of the subducted lithosphere controlled the geometry of subduction in direct analogy to a ping-pong ball. Clearly this problem is related to the problem of the angle of dip considered in the previous section.

Several authors have tested Frank's hypothesis (DeFazio, 1974; Tovish and Schubert, 1978) and have found that it is a fair approximation in some cases and a poor approximation in other cases. It is generally accepted that the arcuate structure of island arcs can be attributed to the rigidity of the descending plate (Laravie, 1975), but the detailed mechanism remains controversial. Yamaoka et al. (1986) and Yamaoka and Fukao (1987) attribute the island arc cusps to lithospheric buckling. It is clear from seismic observations that the cusps represent tears in the descending lithosphere.

Any completely successful numerical model for mantle convection must reproduce the observed arcuate structure of subduction zones.

2.5.5 Subduction Zone Volcanism

Question 2.7: *What is the mechanism for subduction zone volcanism?*

Volcanism is also associated with subduction (Tatsumi and Eggins, 1995). A line of regularly spaced volcanoes closely parallels the trend of almost all the ocean trenches. These volcanoes may result in an island arc or they may occur within continental crust (Figure 2.21). The volcanoes generally lie above where the descending plate is 125 km deep, as illustrated in Figure 2.17. It is far from obvious why volcanism is associated with subduction. The descending lithosphere is cold compared with the surrounding mantle, and thus it acts as a heat sink rather than as a heat source. The downward flow of the descending slab is expected to entrain flow in the overlying mantle wedge. However, this flow will be primarily downward; thus, magma cannot be produced by pressure-release melting. One possible source of heat is frictional dissipation on the fault plane between the descending lithosphere and the overlying mantle (McKenzie and Sclater, 1968; Oxburgh and Turcotte, 1968; Turcotte and Oxburgh, 1968). However, there are several problems with generating island arc magmas by frictional heating. When rocks are cold, frictional stresses can be high and significant heating can occur. However, when the rocks become hot, the stresses are small, and it may be difficult to produce significant melting simply by frictional heating (Yuen et al., 1978). On the other hand, Kanamori et al. (1998) have used the unusual properties of the 1994 Bolivian earthquake, including a slow rupture velocity, a high stress drop (about 100 MPa), and a low ratio of radiated seismic energy to total strain energy, to infer that melting may have occurred on the fault plane during this earthquake. They suggested a minimum frictional stress of about 55 MPa and calculated a minimum amount of nonradiated seismic

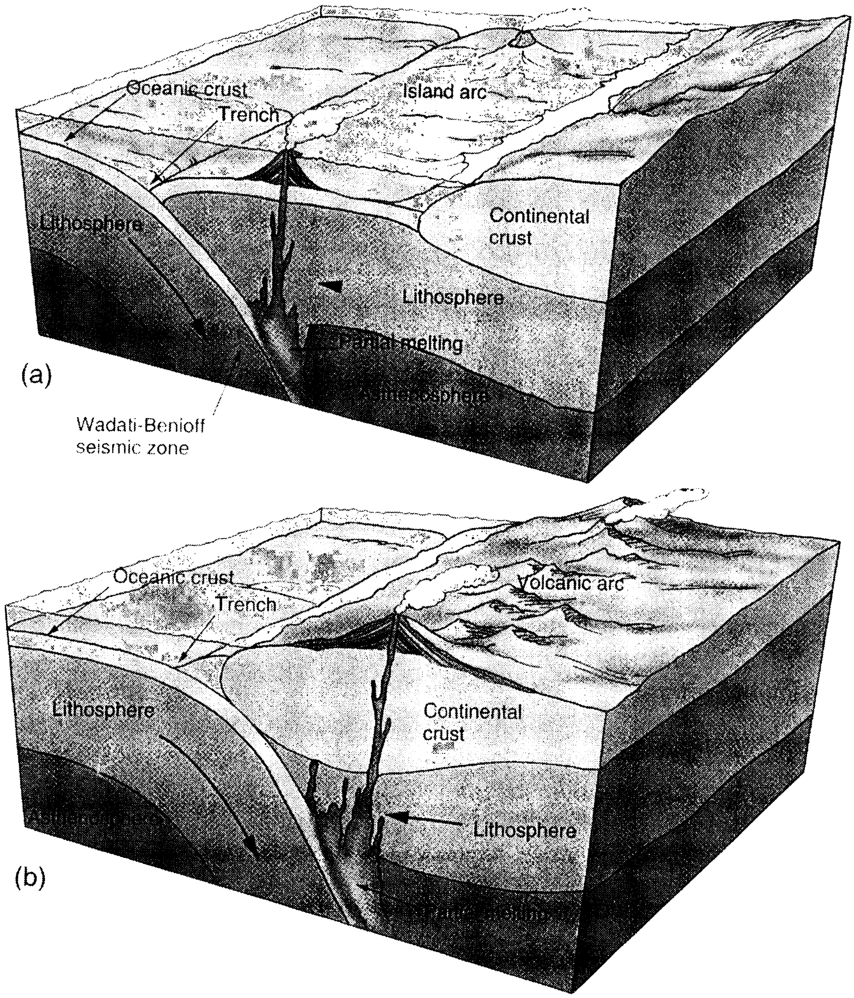


Figure 2.21. Schematic of (a) oceanic lithosphere subducting beneath oceanic lithosphere and the creation of a volcanic island arc, and (b) oceanic lithosphere subducting beneath continental lithosphere and creation of a volcanic chain on the continent. After Tarbuck and Lutgens (1988).

energy equal to about 10^{18} J, sufficient to have melted a layer on the fault plane about 300 mm thick.

One proposed explanation for arc volcanism involves interactions between the descending slab and the induced flow in the overlying mantle wedge, leading to heating of the descending oceanic crust and melting (Marsh, 1979). Many thermal models of the subduction zone have been produced (e.g., Oxburgh and Turcotte, 1970; Toksöz et al., 1971; Turcotte and Schubert, 1973; Hsui and Toksöz, 1979; Hsui et al., 1983; Peacock et al., 1994; Ponko and Peacock, 1995; Iwamori, 1997; Kincaid and Sacks, 1997). All these models show that there is great difficulty in producing enough heat to generate the observed volcanism, since the subducted cold lithospheric slab is a very strong heat sink and depresses the isotherms above the slab.

Water released when hydrated minerals in the subducted oceanic crust are heated can contribute to melting by depressing the solidus temperature of the crustal rocks and adjacent mantle wedge rocks (Anderson et al., 1976; Ringwood, 1977a; Bird, 1978a). However, the bulk of the volcanic rocks at island arcs have near-basaltic compositions and erupt at temperatures very similar to eruption temperatures at accretional margins. Studies of the petrology of island arc magmas (Hawkesworth et al., 1994) indicate that they are primarily the result of the partial melting of fertile mantle rocks in the mantle wedge above the descending slab.

Nevertheless, there is geochemical evidence that the subducted oceanic crust does play an important role in island arc volcanism. Beryllium isotopic studies of volcanic rocks in subduction settings have revealed ^{10}Be enrichments relative to mid-ocean ridge and ocean island basalts that are attributed to sediment subduction (Tera et al., 1986; Sigmarsson et al., 1990). One way to incorporate ^{10}Be from subducted sediments into island arc magmas is through dehydration of the sediments and transport of beryllium with the liberated water (Tatsumi and Isoyama, 1988). Thus, direct melting of the subducted oceanic crust and lithosphere is not required to explain the ^{10}Be excess in island arc volcanic rocks. Other evidence that the subducted oceanic crust is important in island arc magmatism is the location of the surface volcanic lines, which has a direct relationship to the geometry of subduction. In some cases two flaps of slab subduct at different angles, as in the Aleutians. For the shallower dipping slab, the volcanic line is further from the trench, keeping the depth to the slab beneath the volcanic line nearly constant (Kay et al., 1982).

The basic physical processes associated with subduction zone volcanism remain enigmatic, though it is apparent that the subducted oceanic crust triggers this volcanism. However, substantial melting of the subducted crust only occurs when young and relatively hot lithosphere is being subducted (Drummond and Defant, 1990; Kay et al., 1993). The bulk of the volcanism is directly associated with the melting of the mantle wedge similar to the melting beneath an accretional plate margin. A possible explanation for island arc volcanism has been given by Davies and Stevenson (1992). They suggest that "fluids" from the descending oceanic crust induce melting and create sufficient buoyancy in the partially melted mantle wedge rock to generate an ascending flow and further melting through pressure release. This process may be three dimensional with along-strike ascending diapirs associated with individual volcanic centers. Sisson and Bronto (1998) have analyzed the volatile content of primitive magmas from Galunggung volcano in the Indonesian arc and concluded that the magmas were derived from the pressure-release melting of hot mantle peridotite. There is no evidence that volatiles from the subducted oceanic crust were directly involved in the formation of these magmas. We conclude that many aspects of island arc volcanism remain unexplained.

2.5.6 Back-arc Basins

Question 2.8: *Why do back-arc basins form?*

In some subduction zones, a secondary accretionary plate margin lies behind the volcanic line (Karig, 1971). This back-arc spreading is similar to the seafloor spreading that is occurring at ocean ridges. The composition and the structure of the ocean crust that is being created are the same. Behind-arc spreading has created marginal basins such as the Sea of Japan.

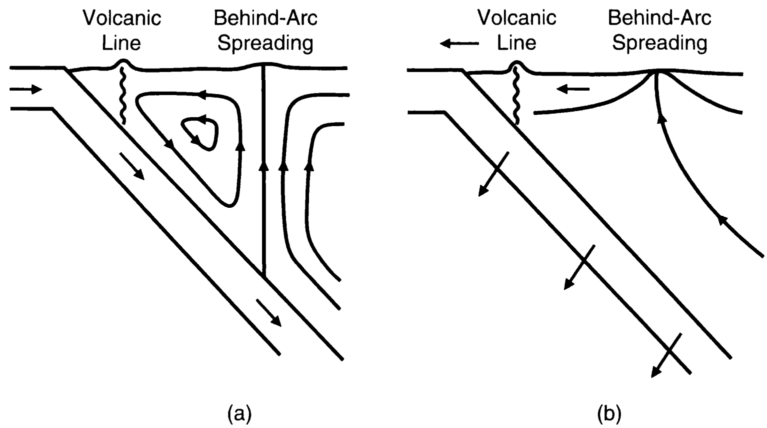


Figure 2.22. Models for the formation of marginal basins. The descending slab, volcanic line, and behind-arc spreading axis are shown. The mantle wedge is the region above the descending slab. (a) Secondary mantle convection induced by the descending lithosphere. (b) Ascending convection generated by the foundering of the sinking lithosphere and the seaward migration of the trench.

A number of explanations have been given for behind-arc spreading (Hynes and Mott, 1985). One hypothesis is that the descending lithosphere induces a secondary convection cell, as illustrated in Figure 2.22a (Toksöz and Hsui, 1978a; Hsui and Toksöz, 1981). An alternative hypothesis is trench rollback, where the ocean trench migrates away from an adjacent continent because of the transverse motion of the descending lithosphere. Behind-arc spreading occurs in response to the rollback, as illustrated in Figure 2.22b (Chase, 1978; Garfunkel et al., 1986).

A number of authors have proposed that there are basically two types of subduction zones (Wilson and Burke, 1972; Molnar and Atwater, 1978; Uyeda and Kanamori, 1979). If the adjacent continent is being driven up against the trench, as in Chile, marginal basins do not develop. If the adjacent continent is stationary relative to the trench, as in the Marianas, the foundering of the lithosphere leads to a series of marginal basins as the trench migrates seaward. Jarrard (1986) has provided a more extensive classification of subduction zones. There is evidence that behind-arc spreading centers are initiated at volcanic lines (Karig, 1971). The lithosphere at the volcanic line may be sufficiently weakened by heating that it fails under tensional stress.

2.6 Hot Spots and Mantle Plumes

Question 2.9: *Are there plumes in the mantle beneath hot spots, and if so, from what depth(s) do they originate?*

Not all volcanism is restricted to the plate margins. Figures 2.4a, 2.5, and 2.6 show large geoid highs over regions of intraplate volcanism, such as Hawaii and Iceland, known as hot spots. The locations of some major hot spots around the globe are given in Figure 2.23. Morgan (1971) attributed hot spot volcanism to mantle plumes. Mantle plumes are quasi-cylindrical concentrated upflows of hot mantle material and they represent a basic form of

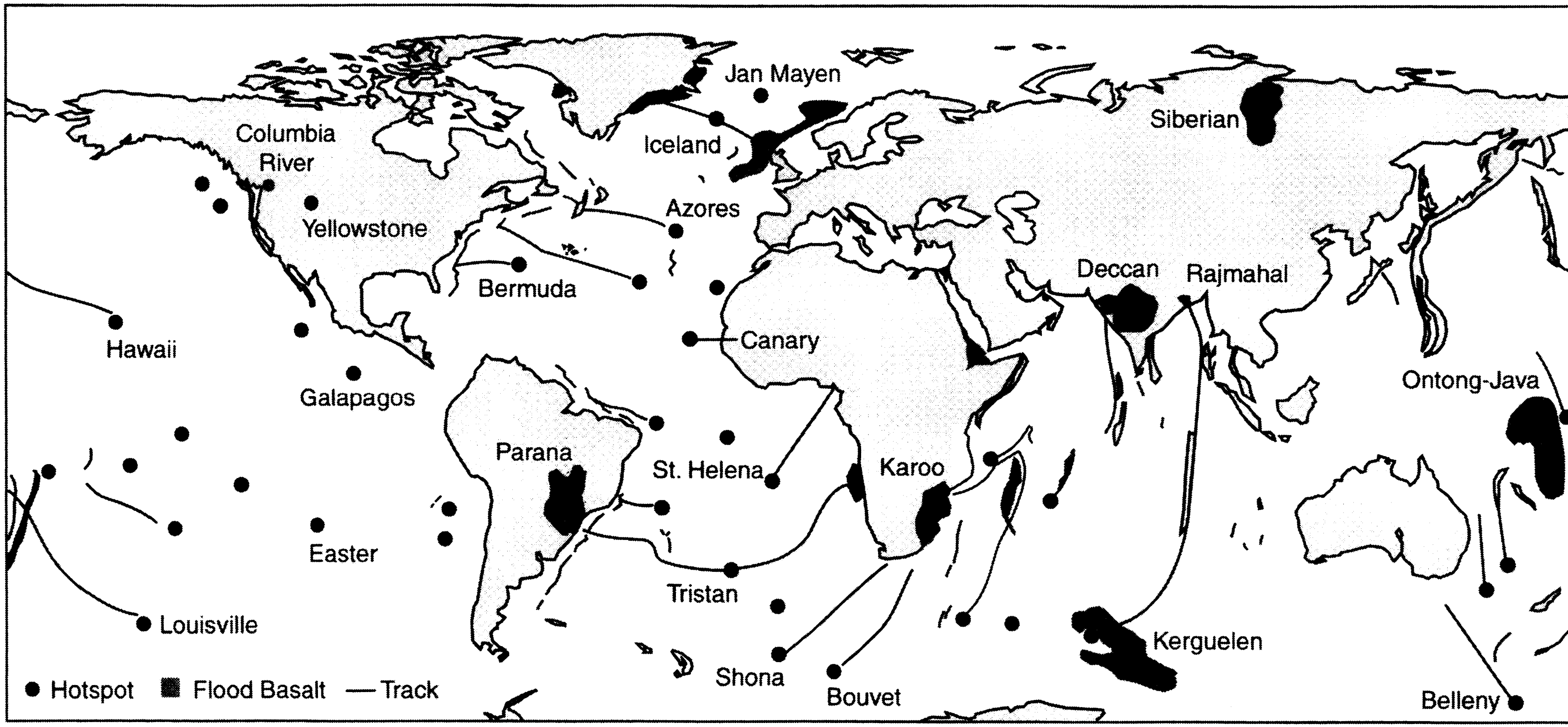


Figure 2.23. Map of major hot spots, hot spot tracks, and flood basalt provinces.

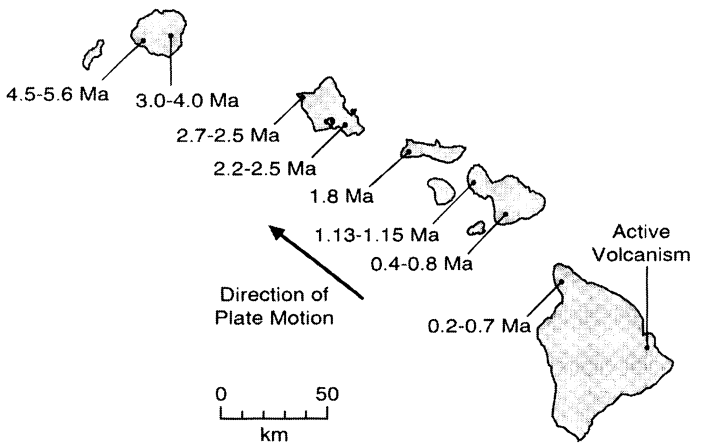
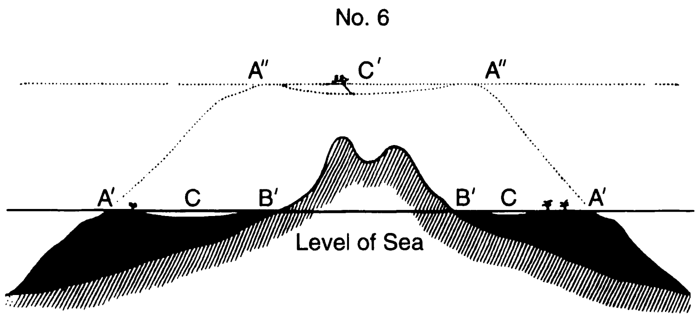


Figure 2.24. Ages in Ma of volcanic rocks on the Hawaiian Island chain. The volcanic rocks on the islands age systematically to the northwest, parallel to the direction of plate motion. The islands are also more eroded to the northwest.

upwelling in the convecting mantle (Bercovici et al., 1989a). Pressure-release melting in the hot ascending plume rock produces the basaltic volcanism that is forming the Hawaiian Island chain. The hypothesis of fixed mantle plumes beneath overriding plates explains the systematic age progression of the Hawaiian-Emperor island-seamount chain (Figure 2.24) and the deepening of the seafloor with increasing age along the chain. Although the ideas of plate tectonics and mantle plumes were unknown to him in the 1830s, Charles Darwin concluded from his geologic observations of ocean islands and their coral reefs (Figure 2.25) that coral reefs and atolls formed through the aging and subsidence of the islands. Darwin's insightful observations and coral reef theory are presented in *The Geology of the Voyage of H.M.S. Beagle, Part I: Structure and Distribution of Coral Reefs* (see Barrett and Freeman, 1987).

It is remarkable that more than 150 years ago, Darwin inferred the subsidence of the ocean floor with age along island chains. He wrote, "Finally, when the two great types of structure, namely barrier-reefs and atolls on the one hand, and fringing-reefs on the other, are laid down on a map, they offer a grand and harmonious picture of the movements which the crust of the earth has undergone within a late period. We there see vast areas rising, with volcanic matter every now and then bursting forth. We see other wide spaces sinking without any volcanic outbursts; and we may feel sure that the movement has been so slow as to have allowed the corals to grow up to the surface, and so widely extended as to have buried over the broad face of the ocean every one of those mountains, above which the atolls now stand like monuments, marking the place of their burial." In an introduction to *The Works of Charles Darwin* (Barrett and Freeman, 1987), J. W. Judd (in 1890) refers to the following excerpt from Darwin's correspondence: " 'It still seems to me a marvellous thing that there should not have been much, and long-continued, subsidence in the beds of the great oceans. I wish some doubly rich millionaire would take it into his head to have borings made in some of the Pacific and Indian atolls, and bring home cores for slicing from a depth of 500 or 600 feet.' " While marine geologists and geophysicists of our day may not be the millionaires of Darwin's musings, they have brought home the cores that Darwin hoped for and have confirmed his theory.



A' A': Outer edges of the barrier-reef at the level of the sea. The cocoa-nut trees represent coral-islets formed on the reef.
C C: The lagoon-channel.
B'B': The shores of the island, generally formed of low alluvial land and of coral detritus from the lagoon-channel.
A''A'': The outer edges of the reef, now forming an atoll.
C': The lagoon of the newly-formed atoll. According to the scale the depth of the lagoon and of the lagoon-channel is exaggerated.

Figure 2.25. A sketch from Darwin's coral reef theory illustrating the formation of a barrier reef through island subsidence.

Plates and plumes are both consequences of mantle convection, but plumes are not required in plate tectonic theory per se. Although mantle plumes are expected to exist on theoretical grounds, and plumes do occur in relevant laboratory and numerical experiments on convection, direct observational evidence that mantle plumes exist beneath hot spots remains elusive. Seismic tomography holds the best promise to discover mantle plumes through the seismic velocity anomaly that must be associated with the hot upwelling plume material. An example of the use of seismic tomography to image mantle plumes, which illustrates the promise as well as the practical difficulties involved, is shown in Figure 2.26 from a study of mantle structure beneath the Iceland hot spot by Wolfe et al. (1997). The most likely source of mantle plumes is the hot material in the thermal boundary layer at the base of the mantle. Alternatively, if the phase transition at 660 km depth is a boundary that physically separates the upper mantle from the lower mantle, then upper mantle plumes could also originate at this depth. Chapter 11 will be devoted to a detailed discussion of hot spots and mantle plumes.

2.7 Continents

2.7.1 Composition

As described in the previous sections, the development of plate tectonics involved primarily the ocean basins. Yet the vast majority of geological data concerns the continents. There is little evidence for plate tectonics in the continents, and this is certainly one reason why most geologists did not accept the arguments in favor of continental drift and mantle convection for so long.

Question 2.10: *How were the continents formed?*

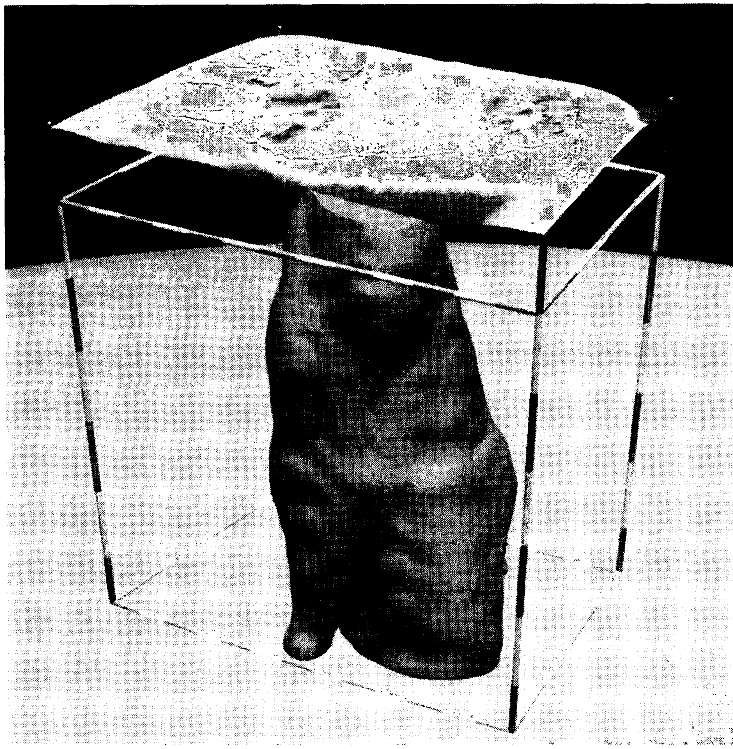


Figure 2.26. Seismically delineated plume structure beneath central Iceland (Wolfe et al., 1997).

For a color version of this figure, see plate section.

The surface rocks of the continental crust are much older than the rocks of the oceanic crust. They also have a much more silicic composition. The continents include not only the area above sea level but also the continental margins. It is difficult to provide an absolute definition of the division between oceanic and continental crust. In most cases it is appropriate to define the transition to occur at an ocean depth of 3 km. The area of the continents including the margins is about 1.9×10^8 km^2, or 37% of the surface of the Earth. Schubert and Sandwell (1989) have provided estimates for the volume of the continents.

The rocks that make up the continental crust are, in bulk, more silicic and therefore less dense than the basaltic rocks of the oceanic crust. This difference makes the continental lithosphere gravitationally stable and prevents it from being subducted. Although continental crust is not destroyed by subduction, it can be recycled indirectly by the subduction of sediments or by delamination.

It is relatively easy to estimate the composition of the upper continental crust but it is difficult to estimate the composition of the crust as a whole. Direct evidence for the composition of the lower continental crust comes from surface exposures of high-grade metamorphic rocks and lower crustal xenoliths transported to the surface in diatremes and magma flows. Indirect evidence for the composition of the lower crust comes from comparisons between seismic velocities and laboratory studies of relevant minerals (Gao et al., 1998). Estimates of the bulk composition of the continental crust are given in Table 2.1. In Table 2.2 the average composition from Table 2.1 is compared with a typical basalt composition. Also included

Table 2.1. ***Estimates for the Composition of the Bulk Continental Crust***

	1	2	3	4	5	6	7	8	9	10	Average
SiO_2	61.9	63.9	57.8	61.9	62.5	63.8	63.2	57.3	63.2	59.1	61.5
TiO_2	1.1	0.8	1.2	0.8	0.7	0.7	0.6	0.9	0.7	0.7	0.8
Al_2O_3	16.7	15.4	15.2	15.6	15.6	16.0	16.1	15.9	14.8	15.8	15.7
FeO	6.9	6.1	7.6	6.2	5.5	5.3	4.9	9.1	5.6	6.6	6.4
MgO	3.5	3.1	5.6	3.1	3.2	2.8	2.8	5.3	3.15	4.4	3.7
CaO	3.4	4.2	7.5	5.7	6.0	4.7	4.7	7.4	4.66	6.4	5.5
Na_2O	2.2	3.4	3.0	3.1	3.4	4.0	4.2	3.1	3.29	3.2	3.3
K_2O	4.2	3.0	2.0	2.9	2.3	2.7	2.1	1.1	2.34	1.9	2.5

Note: 1. Goldschmidt (1933), 2. Vinogradov (1962), 3. Pakiser and Robinson (1966), 4. Ronov and Yaroshevsky (1969), 5. Holland and Lambert (1972), 6. Smithson (1978), 7. Weaver and Tarney (1984), 8. Taylor and McLennan (1985), 9. Shaw et al. (1986), 10. Rudnick and Fountain (1995).

Table 2.2. ***Average Composition of Basalts (Nockolds, 1954), the Mean Composition of the Continental Crust from Table 2.1, and the Mean Composition of Archean and Post-Archean Clastic Sediments (Taylor and McLennan, 1985, p. 99)***

	Basalt	Average Continental Crust	Archean Clastic Sediments	Post-Archean Clastic Sediments
SiO_2	50.8	61.5	65.9	70.4
TiO_2	2.0	0.8	0.6	0.7
Al_2O_3	14.1	15.7	14.9	14.3
FeO	9.0	6.4	6.4	5.3
MgO	6.3	3.7	3.6	2.3
CaO	10.4	5.5	3.3	2.0
Na_2O	2.2	3.3	2.9	1.8
K_2O	0.8	2.5	2.2	3.0

in Table 2.2 are the mean compositions of very old Archean ($\sim$2 + Gyr) and post-Archean clastic sediments. Estimates of the mean composition of the continental crust are clearly more basic (less silicic) than the composition of the upper continental crust, but they do not approach a basaltic composition. The fact that upper continental crust is more silicic than lower continental crust is consistent with the remelting hypothesis for the origin of granites. However, if the only mantle melt responsible for forming the continental crust is basalt, then the mean composition of the continental crust should be basaltic. This is clearly not the case. There is also some evidence that the continental crust has become more silicic with time (Ronov, 1972). This change is supported by the comparison between Archean and post-Archean sediments given in Table 2.2.

2.7.2 Delamination and Recycling of the Continents

Question 2.11: *Does delamination play an important role in recycling continental crust and lithosphere?*

There is no evidence that continental lithosphere is subducted. This is generally attributed to the buoyancy of the continental crust, which results in the continental lithosphere being gravitationally stable. However, the mantle portion of the continental lithosphere is sufficiently cold and dense to be gravitationally unstable. Thus it is possible for the lower part of the continental lithosphere, including the lower continental crust, to delaminate and sink into the lower mantle. This is partial subduction.

Continental delamination was proposed and studied by Bird (1978b, 1979) and Bird and Baumgardner (1981). These authors suggested that delamination is occurring beneath the Himalaya and Zagros collision zones and is also responsible for the elevation of the Colorado Plateau. Hildebrand and Bowring (1999) have argued in favor of delamination at collision zones. McKenzie and O'Nions (1983) have suggested that delamination occurs at island arcs. The zone of high seismic velocities found beneath the Transverse Ranges in California by Humphreys and Clayton (1990) can be interpreted to be the result of delamination. The structure of the Alps can also be associated with crustal and lithospheric delamination (Butler, 1986; Laubscher, 1988). Sacks and Secor (1990) have suggested delamination in continental collision zones. Delamination can be associated with certain types of magmatism (Kay and Kay, 1991, 1993).

There are a number of continental areas in which the mantle lithosphere is absent. One example is the western United States. Crustal doubling such as in Tibet has also been attributed to the absence of mantle lithosphere beneath Asia (Molnar and Tapponnier, 1981). Plateau uplifts such as the Altiplano in the Andes are associated with the absence of mantle lithosphere. In the Puna section of the Altiplano there is direct geochemical evidence supporting the delamination of the lower continental crust (Kay and Kay, 1993). Delamination is an efficient mechanism for the removal of continental lithosphere – for example, in the western United States. Alternative mechanisms for thinning the lithosphere include heat transfer from an impinging plume and heat transport by magmas. The former process may be very slow (Emerman and Turcotte, 1983) and the latter one requires very large volumes of magma (Lachenbruch and Sass, 1978). Moore et al. (1998b) have modeled the plume–lithosphere interaction at the Hawaiian Swell and have shown how three-dimensional drip instabilities of the lower lithosphere can lead to rapid (10 Myr time scale) lithospheric thinning by a mantle plume.

A major unknown is whether delamination includes the lower continental crust. It is well documented that a soft intracrustal zone exists in orogenic zones (Hadley and Kanamori, 1977; Mueller, 1977; Eaton, 1980; Yeats, 1981; Turcotte et al., 1984). The presence of a soft layer at an intermediate depth in the crust can be attributed to the presence of quartz (Kirby and McCormick, 1979). Delamination at this intracrustal weak zone can explain intracrustal decollements in the Alps (Oxburgh, 1972) and in the southern Appalachians (Cook et al., 1979).

Direct estimates for the rate of recycling of continental crust on the Earth have been given by several authors based on various lines of reasoning. Armstrong (1981) gave a rate of $2 \pm 1\,\mathrm{km^3\,yr^{-1}}$, DePaolo (1983) gave $2.5 \pm 1.2\,\mathrm{km^3\,yr^{-1}}$, Reymer and Schubert (1984) gave $0.59\,\mathrm{km^3\,yr^{-1}}$, and Turcotte (1989a) gave $1.49\,\mathrm{km^3\,yr^{-1}}$. These rates are about 10% of the rate of production of oceanic crust on the Earth, which is $17 \pm 2\,\mathrm{km^3\,yr^{-1}}$ (Turcotte and Schubert, 1982, p. 166). The only mechanism that can recycle these relatively large volumes of continental crust is crustal delamination (Turcotte, 1989a, b).

Bird (1979) hypothesized that delamination occurred in a manner like subduction as a consequence of lithospheric flexure beneath a decollement similar to the flexure of the oceanic lithosphere seaward of an ocean trench. However, this type of delamination would result in large surface gravity signatures that are not observed. Another model for delamination has been given by Houseman et al. (1981), Houseman and England (1986), and England and Houseman (1986). These authors treat the continental lithosphere as a viscous fluid which thickens in a collision zone. The thick cold lithosphere is gravitationally unstable and sinks or delaminates. Pari and Peltier (1996) proposed that continuous mantle downwelling occurs beneath the continents.

Turcotte (1983) proposed the alternative mechanism of lithospheric stoping illustrated in Figure 2.27. Soft mantle rock penetrates the continental crust in a zone of volcanism; possible sites would be the volcanic lines associated with subduction zones adjacent to continents and continental rifts; this penetration occurs at mid-crustal levels where the rocks have the softest rheology (Figure 2.27a). Eventually the continental lithosphere fails along a pre-existing zone of weakness (e.g., a fault) (Figure 2.27b). The decoupled block of continental lithosphere (including the lower crust) sinks into the mantle and is replaced by hot asthenosphere mantle rock. This lithospheric stoping process continues away from the initial zone of volcanism.

There are other mechanisms by which continental crust can be recycled into the mantle (McLennan, 1988). Several are associated with subduction zones. Sediments from the continental crust coat the oceanic crust that is being subducted. Some of these sediments are scraped off in the oceanic trench and form the accretionary prism as illustrated in Figure 2.17, but some sediments are entrained in the subducted oceanic crust (Karig and Kay, 1981). There is also some evidence that the subducted lithosphere can erode and entrain some of the overlying continental crust; however, the estimated volume is small. Also some of the subducted entrained material is returned to the crust in the island arc volcanism. Schubert and Sandwell (1989) have suggested that slivers of continental crust broken off from the

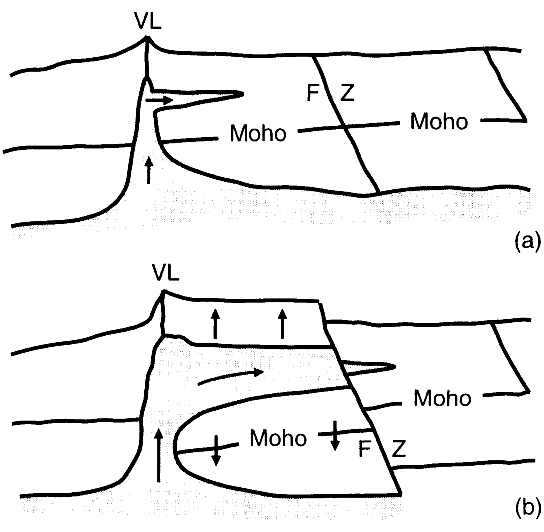


Figure 2.27. Delamination of the continental crust by a lithospheric stoping process. (a) The asthenosphere penetrates into the continental crust along a volcanic line (VL) associated with a subduction zone. It then splits the crust behind the volcanic line along an intracrustal (horizontal) zone of weakness. (b) The lower continental crust and mantle lithosphere beneath the penetrating asthenosphere break away along a pre-existing fault (FZ) and delaminate.

continents and trapped in the seafloor could, if sufficiently small, be subducted with the oceanic lithosphere.

2.7.3 Continental Crustal Formation

Question 2.12: *How is continental crust formed?*

Can currently observed processes, i.e., subduction-related volcanism and continental rift or hot spot volcanism, lead to the formation of continental crust, or was the continental crust primarily formed by processes in the Archean that are no longer active? A related question is whether the formation of the continental crust is continuous or episodic. A primary constraint on models for the generation of the continental crust is its silicic composition. As discussed above, the crust has a mean composition that is more silicic than magmas produced in the mantle today. One hypothesis for the formation of the continental crust is that silicic magmas were generated in the mantle in the Archean and that these magmas are responsible for the silicic composition of the continental crust. Brown (1977) suggested that the direct addition of silicic magmas was primarily responsible for the formation of the silicic continental crust.

An alternative hypothesis for the generation of the continental crust has been given by Kay and Kay (1988). The hypothesis consists of three parts: (1) Basaltic volcanism from the mantle associated with island arc volcanics, continental rifts, and hot spots is responsible for the formation of the continental crust. (2) Intracrustal melting and high-temperature metamorphism is responsible for the differentiation of the crust so that the upper crust becomes more silicic and the lower crust becomes more basic. In a paper entitled "No water, no granites – no oceans, no continents," Campbell and Taylor (1983) argue that basaltic magmas from the mantle intruded into a basaltic continental crust in the presence of water can produce the granitic rocks associated with the continental crust. (3) Delamination of substantial quantities of continental lithosphere including the mantle and lower crust returns a substantial fraction of the more basic lower crust to the mantle. The residuum, composed primarily of the upper crust, thus becomes more silicic.

In this model, the basic cycle responsible for the evolution of the continental crust is as follows. The oceanic crust is created at mid-ocean ridges. Selected elements from the continental crust are added to the oceanic crust by solution and deposition. For example, sediments derived from the continents coat the oceanic crust. Some fraction of the sediments and the altered oceanic crust are subducted at ocean trenches. The remainder of the sediments is returned to the continental crust in accretionary prisms. The subducted oceanic crust and entrained sediments are partially melted beneath island arc volcanoes inducing partial melting in the overlying mantle wedge. The result is a near-basaltic composition with trace element and isotopic compositions contaminated with the signature of the altered oceanic crust and entrained continental sediments.

Island arcs can contribute to the formation of continental crust, in two ways. (1) If an island arc stands on oceanic crust, it generates thick crust. If this island arc subsequently collides with a continent, it can add material in the form of exotic terranes. (2) If a subduction zone is adjacent to a continent, then the subduction zone can add mantle-derived magmas directly to the crust (Figure 2.21). This is happening today in the Andes. The importance of island arc processes in forming continental crust has been discussed in detail by Taylor and White (1965), Taylor (1967, 1977), Jakeš and White (1971), Jakeš and Taylor (1974), and Taylor and McLennan (1985). Flood and rift volcanics and hot spots have added a considerable volume of mantle-derived magmas to the continental crust. McKenzie (1984a) argues that the continental crust is being continuously underplated by extensive intrusive magmas.

It is widely accepted that the silicic rocks associated with continents are produced when basaltic magma intrusions partially melt basaltic or other more silicic continental crustal rocks in the presence of water (Huppert and Sparks, 1988; Luais and Hawkesworth, 1994). In order to produce a continental crust that has a mean composition that is more silicic than basalt, it is necessary to remove the residual basic rock. This is done by the delamination of the lower continental crust.

2.8 Plate Motions

The surface plates are rigid to a first approximation and are in relative motion with respect to each other. The relative motion between two adjacent rigid plates can be described by Euler's theorem. The theorem states that any line on the surface of a sphere can be translated to any other position and orientation on the sphere by a single rotation about a suitably chosen axis passing through the center of the sphere. In terms of the Earth this means that a rigid surface plate can be shifted to a new position by a rotation about a uniquely defined axis. The point where this axis intercepts the surface of the Earth is known as the pole of rotation. This is illustrated in Figure 2.28, where plate B is rotating counterclockwise with respect to plate A at the angular velocity ω about the pole of rotation P. Ridge segments lie on lines of longitude emanating from the pole of rotation. Transform faults lie on small circles with their centers at the pole of rotation.

The relative motion between two adjacent plates is completely specified when the latitude and longitude of the pole of rotation together with the angular velocity of rotation are given. These quantities for the NUVEL-1 model of DeMets et al. (1990) are given in Table 2.3. The plate geometry upon which this model is based consists of the 12 rigid plates illustrated in Figure 2.29. The plate rotation vectors are also shown. The best-fitting plate rotation vectors in this model were obtained using 1,122 data points from 22 plate boundaries. The

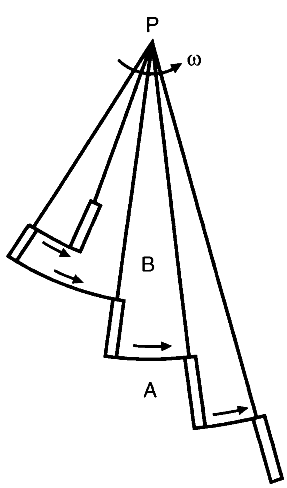


Figure 2.28. Illustration of Euler's theorem. Plate B is moving counterclockwise relative to plate A. The motion is defined by the angular velocity ω about the pole of rotation P. Double lines are ridge segments, and arrows denote direction of motion on transform faults.

data include 277 spreading rate determinations based on magnetic anomalies. An example of the magnetic profiles for the spreading boundary between the Cocos and Pacific plates is given in Figure 2.30. The NUVEL-1 model also uses 232 transform fault azimuths and 724 earthquake slip vectors. The authors found, however, that earthquake slip vectors at ocean trenches systematically misfit whenever convergence is oblique.

Revisions to the geomagnetic time scale have necessitated some recalibration of the NUVEL-1 global plate motion model. The changes consist largely in a reduction of the angular velocities. The revised plate motion model is referred to as NUVEL-1A (DeMets et al., 1994). Angular velocities for model NUVEL-1A are given in Table 2.3.

Table 2.3. ***Euler Vectors for Pairs of Plates Sharing a Boundary Based on the NUVEL-1 Model of DeMets et al. (1990) and the NUVEL-1A Model of DeMets et al. (1994); the First Plate Moves Counterclockwise Relative to the Second Plate***

Plate Pair	Latitude (°N)	Longitude (°E)	ω, NUVEL-1 (deg Myr^{-1})	ω, NUVEL-1A (deg Myr^{-1})
EU-NA	62.4	135.8	0.22	0.21
AF-NA	78.8	38.3	0.25	0.24
AF-EU	21.0	−20.6	0.13	0.12
NA-SA	16.3	−58.1	0.15	0.15
AF-SA	62.5	−39.4	0.32	0.31
AN-SA	86.4	−40.7	0.27	0.26
NA-CA	−74.3	−26.1	0.11	0.10
CA-SA	50.0	−65.3	0.19	0.18
NA-PA	48.7	−78.2	0.78	0.75
CO-PA	36.8	−108.6	2.09	2.00
CO-NA	27.9	−120.7	1.42	1.36
CO-NZ	4.8	−124.3	0.95	0.91
NZ-PA	55.6	−90.1	1.42	1.36
NZ-AN	40.5	−95.9	0.54	0.52
NZ-SA	56.0	−94.0	0.76	0.72
AN-PA	64.3	−84.0	0.91	0.87
PA-AU	−60.1	−178.3	1.12	1.07
EU-PA	61.1	−85.8	0.90	0.86
CO-CA	24.1	−119.4	1.37	1.31
NZ-CA	56.2	−104.6	0.58	0.55
AU-AN	13.2	38.2	0.68	0.65
AF-AN	5.6	−39.2	0.13	0.13
AU-AF	12.4	49.8	0.66	0.63
AU-IN	−5.6	77.1	0.31	0.30
IN-AF	23.6	28.5	0.43	0.41
AR-AF	24.1	24.0	0.42	0.40
IN-EU	24.4	17.7	0.53	0.51
AR-EU	24.6	13.7	0.52	0.50
AU-EU	15.1	40.5	0.72	0.69
IN-AR	3.0	91.5	0.03	0.03

Abbreviations: PA, Pacific; NA, North America; SA, South America; AF, Africa; CO, Cocos; NZ, Nazca; EU, Eurasia; AN, Antarctica; AR, Arabia; IN, India; AU, Australia; CA, Caribbean. See Figures 2.1 and 2.29 for plate geometries.

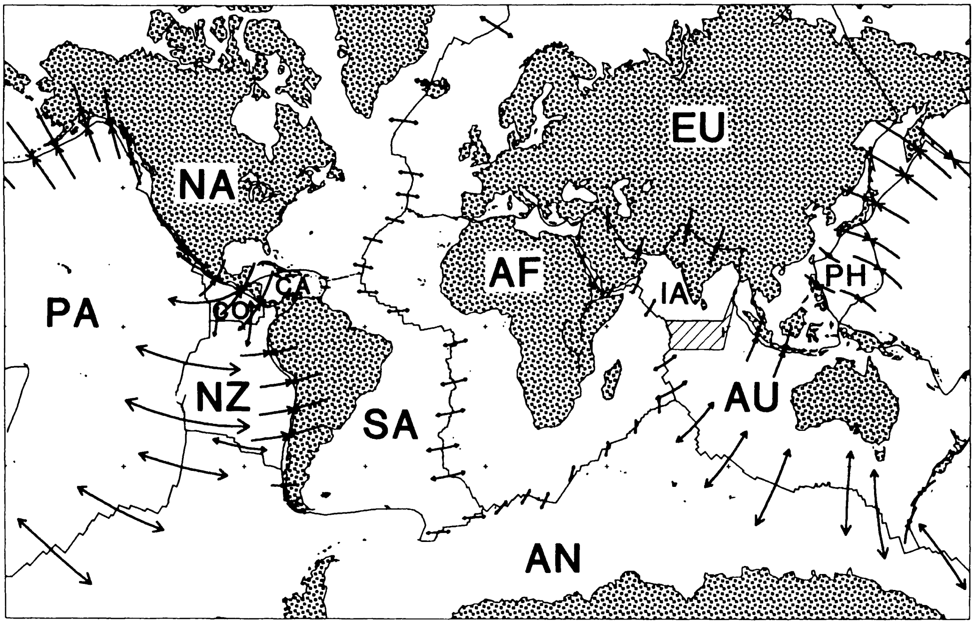


Figure 2.29. The 12 rigid plates used in the NUVEL-1 model of DeMets et al. (1990) are shown. The plate rotation vectors are also shown.

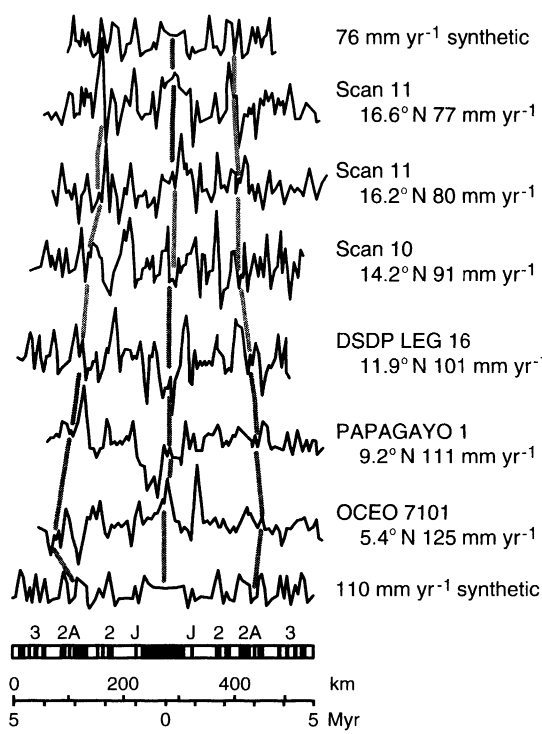


Figure 2.30. Cocos–Pacific magnetism profiles from the National Geodetic Data Center (NGDC) archives. Six profiles are shown with (half) spreading rates from 77 $\mathrm{mm\,yr^{-1}}$ to 125 $\mathrm{mm\,yr^{-1}}$. Synthetic profiles for spreading rates of 76 $\mathrm{mm\,yr^{-1}}$ and 110 $\mathrm{mm\,yr^{-1}}$ are shown. Also shown are the sequence of reversals and the time scale for a spreading rate of 110 $\mathrm{mm\,yr^{-1}}$.

The reduced angular velocities agree better with plate speeds deduced from space geodetic measurements.

The magnitude of the relative velocity u_{rel} between plates at any boundary is

$$u_{\mathrm{rel}} = \omega a \sin \Delta \tag{2.8.1}$$

where a is the radius of the Earth and Δ is the angle subtended at the center of the Earth by the pole of rotation P and the point A on the plate boundary (Figure 2.31a). The angle Δ is related to the colatitude Θ and east longitude Ψ of the rotation pole and the colatitude Θ' and east longitude Ψ' of the point on the plate boundary A by

$$\cos \Delta = \cos \Theta \cos \Theta' + \sin \Theta \sin \Theta' \cos(\Psi - \Psi') \tag{2.8.2}$$

The geometry is illustrated in Figure 2.31b, where s is the surface arc between points A and P, and O is the center of the Earth. With (2.8.1) and (2.8.2) one can find the magnitude of the relative velocity between two plates at any point on the boundary between the two plates, once the latitude and longitude of the point on the boundary have been specified. As a specific example let us determine the magnitude of the relative velocity across the San Andreas fault at San Francisco (37.8°N, 122°W). We assume that the entire relative velocity between the rigid Pacific and North American plates is accommodated on this fault. From the NUVEL-1 model (Table 2.3), we find $\Theta = 41.3°$ and $\Psi = -78.2°$. Since $\Theta' = 52.2°$

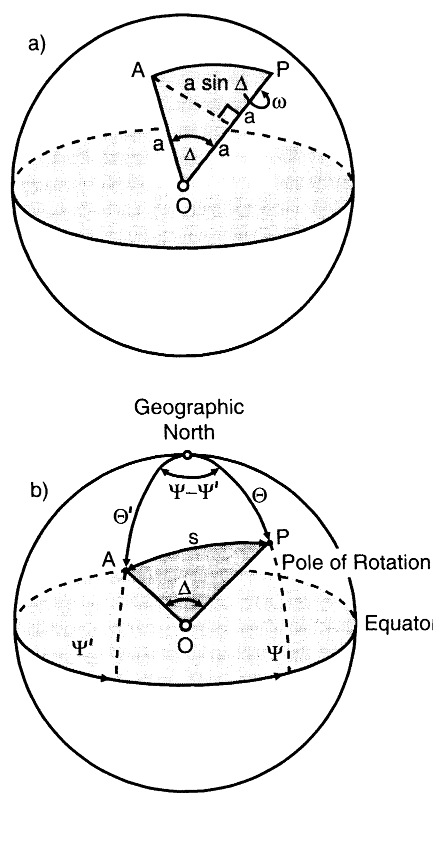


Figure 2.31. (a) Geometry for the determination of the magnitude of the relative plate velocity u_{rel} at a point A on the boundary between two plates in terms of the rate of rotation ω about a pole P. (b) Geometry for determining the angle between a point A on a plate and a pole of rotation.

and $\Psi' = 238°$, we find from (2.8.2) that $\Delta = 33.7°$; with $\omega = 0.78°\ \mathrm{Myr}^{-1}$ we find from (2.8.1) that the magnitude of the relative velocity across the fault is $48\ \mathrm{mm\ yr}^{-1}$.

Global plate motions show that new oceanic crust is being created at accretional margins at the rate of $2.8\ \mathrm{km}^2\ \mathrm{yr}^{-1}$ (Parsons, 1981); due to the conservation of plate area, the rate of subduction of oceanic crust is very close to $2.8\ \mathrm{km}^2\ \mathrm{yr}^{-1}$. Slight differences can be associated with changes in the area of the continents due to continental collisions or continental rifting.

The discussion given above implies that the plates are rigid. This is a reasonable approximation at any instant in time, but as the plates evolve in time, deformation must take place in the interiors of the plates (Dewey, 1975; Gordon, 1998). A configuration of rigid plates with boundaries made up of accretionary margins, subduction zones, and transform faults cannot evolve in time without overlaps and holes. The required interior deformations are generally accommodated by relatively broad plate boundaries. The western United States is an example. Interior plate deformations occur primarily in the "softer" continents.

2.9 The Driving Force for Plate Tectonics

The motions of the surface plates were described in the last section. We now show that these motions provide some information on the forces that drive plate tectonics. The basic question we address is:

***Question 2.13:** What are the forces that drive plate tectonics?*

There are three primary candidates for the forces that drive plate tectonics. These are:

(i) *Slab pull.* The cold subducted lithosphere at ocean trenches is denser than the hotter mantle adjacent to it. This negative buoyancy results in a downward body force. As we have discussed above, the descending lithosphere is attached to the adjacent surface plate. The resulting body force on the surface plate is known as slab pull (Figure 2.12).

(ii) *Ridge push.* The mid-ocean ridges are elevated above the adjacent ocean basins. This results in a body force pushing the adjacent ridge segments apart. This force is also known as gravitational sliding (Figure 2.12).

(iii) *Basal tractions.* If the mantle flow beneath a surface plate is faster than the motion of the plate, the mantle will drag the plate along; the result is a basal traction that will drive the motion of the plate.

Forsyth and Uyeda (1975) summarized the statistics of present-day plate motions and their results are given in Table 2.4. This table gives the perimeter of each plate, the length of plate boundary occupied by ridges and trenches, plate area, and the speed of each plate. By virtue of its size and speed, the Pacific plate alone contains more than two-thirds of the entire kinetic energy of the lithosphere relative to the hot spot reference frame. Together, the Pacific and Indian plates contain more than 90% of the lithospheric kinetic energy, and the three fast-moving large plates, Pacific, India and Nazca, contain in excess of 95% of the kinetic energy of the plates, while occupying only slightly more than half of the Earth's surface area.

Why is the kinetic energy so unevenly partitioned among the plates? A partial explanation is found in Figure 2.32, which shows total plate area, plate area occupied by continents, and

Table 2.4. *Summary of Major Plate Dimensions*

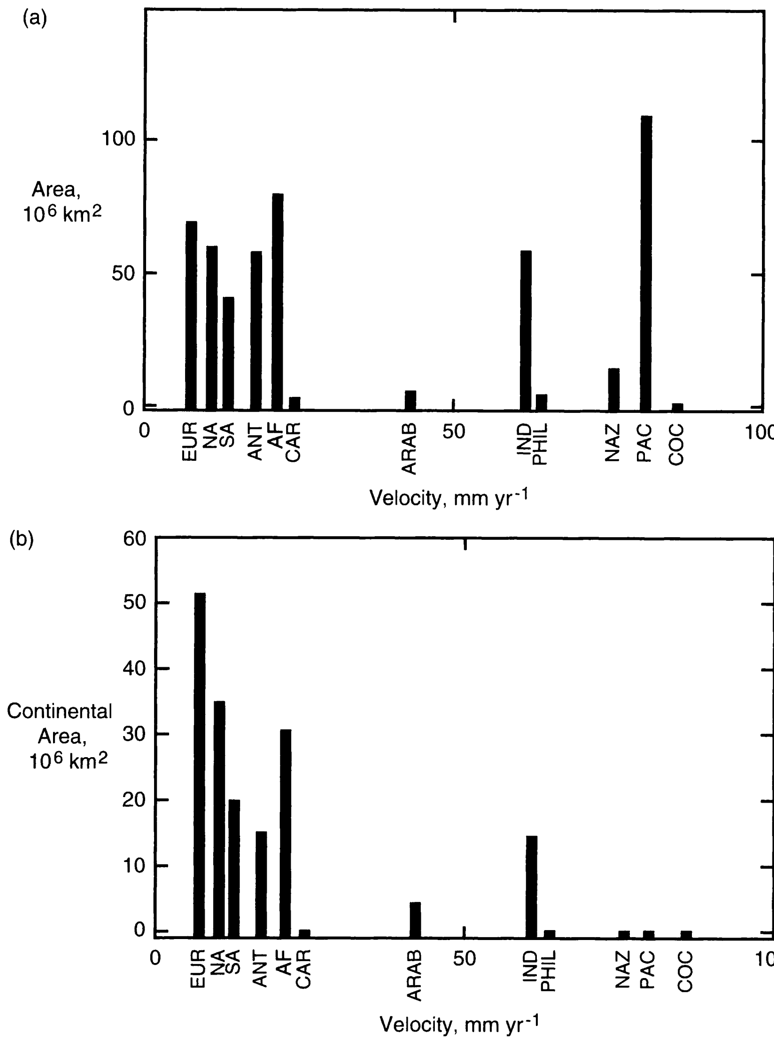
Plate	Area	Continental Area	Average Absolute Velocity	Circumference	Length	
					Ridge	Trench
	(10^6 km^2)	(10^6 km^2)	(mm yr^{-1})	(10^2 km)	(10^2 km)	
NA	60	36	11	388	146	12
SA	41	20	13	305	87	5
PA	108	–	80	499	152	124
AN	59	15	17	356	208	–
IN	60	15	61	420	124	91
AF	79	31	21	418	230	10
EU	69	51	7	421	90	–
NZ	15	–	76	187	76	53
CO	2.9	–	86	88	40	25
CA	3.8	–	24	88	–	–
PH	5.4	–	64	103	–	41
AR	4.9	4.4	42	98	30	–

Note: For plate abbreviations, see Table 2.3. PH, Philippine plate.

fraction of plate circumference occupied by ridges and trenches versus plate velocity for each plate (Forsyth and Uyeda, 1975). According to Figures 2.32b and d, plate velocity is sensitive to the fraction of the plate occupied by continental crust and to the length of subducting slabs attached to the plate. In general, continents occupy slow-moving plates, while fast-moving plates are connected to descending slabs. As seen in Figure 2.32d, the length of subducting boundaries clearly divides the major plates into two groups: the "energy-containing" plates attached to subducting slabs and the "energy-deficient" plates that lack major subducting slabs.

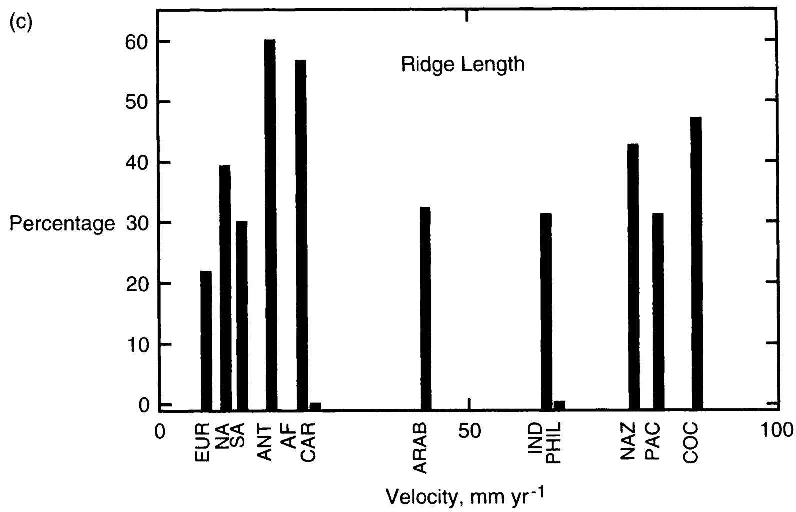
These associations imply that trench pull is the main driving force for plate motions, and that continents tend to slow plates down, probably through increased basal drag. In contrast, Figure 2.32c indicates that ridges do not control plate speeds. This leads to the conclusion that the ridge push force is small compared to the trench pull force. Further conclusions can be drawn from the fact that the small Nazca and Cocos plates are subducted at about the same rates as the large Pacific plate. This is evidence that the basal traction forces on the plates are negligible since all three plates are totally oceanic. The similar velocity of subduction of these oceanic plates is taken as evidence that the downward buoyancy force on the descending lithosphere is nearly balanced by the viscous resistance to its downward motion (Figure 2.12). This near-balance acts as a "velocity governor" on the rate of subduction and the slab pull represents the small excess force between the downward gravitational body force and the viscous resisting force. These conclusions regarding the forces that drive the plates have been generally confirmed by the study of Jurdy and Stefanick (1991), who showed that the stress distributions in the plates inferred from earthquake focal mechanisms and other sources are generally consistent with this picture.

As pointed out above, the plates containing continents move more slowly than the purely oceanic plates. An important question is whether this is due to the lack of subduction zones on the "continental" plates or due to basal tractions associated with the continents. As we will discuss in Chapters 3 and 4, the thickness of the continental lithosphere remains a subject of controversy. If the continental lithosphere has a thickness of $\sim$400 km rather than $\sim$200 km, basal tractions would be expected to be important. Stoddard and Abbott (1996)



have addressed this problem and have concluded that basal continental tractions are not important.

Most studies of the forces that drive the plates have tested various assumptions against either plate velocities or inferred stress directions (e.g., Richardson et al., 1979). Bird (1998) has used a global model of laterally heterogeneous plates of nonlinear rheology separated by faults with low friction to test hypotheses on plate driving mechanisms against both plate velocities and stresses. He concludes that a model in which plates move over a resisting mantle at velocities dictated by their attached subducting slabs does not predict correct stress directions. He also finds that a model in which driving forces result only from elevation differences between rises and trenches and are balanced by basal drag and fault friction fails to predict correct plate velocities. A better match of model predictions to plate velocity and stress measurements occurs for a model in which the mantle supplies a forward basal



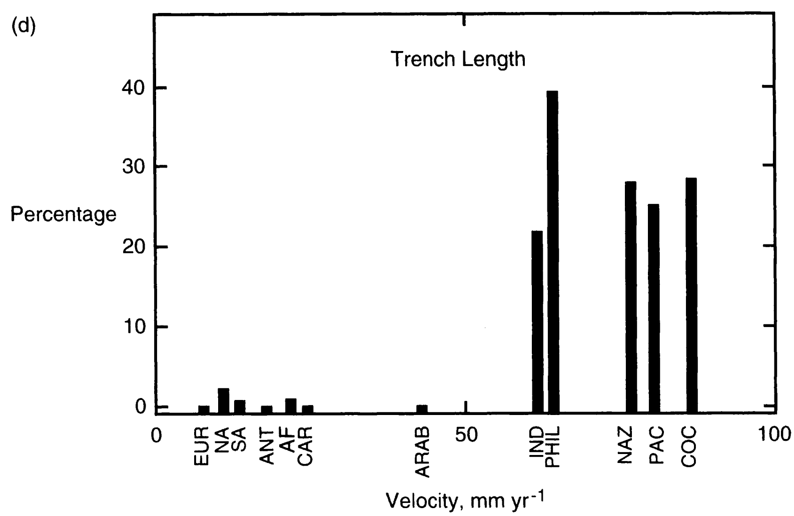


Figure 2.32. Plate tectonic statistics: (a) plate velocity versus area; (b) plate velocity versus continental area; (c) plate velocity versus percentage of perimeter that is ridge; (d) plate velocity versus percentage of perimeter that is trench.

traction to plates with continents, while oceanic plates experience negligible basal shear tractions. Bird's (1998) plate model is coupled to the underlying mantle in an artificial way, i.e., through imposed lower boundary conditions. A complete understanding of the forces that drive the plates will likely require a coupled and self-consistent model of both plates and mantle convection, i.e., a multi-rheological mantle convection model in which plates arise naturally as part of the model.

The rigidity of the surface plates also has a profound effect on the convective flows at depth. As will be shown, the thermal convection of an isoviscous fluid contains only the poloidal surface flows associated with sources of fluid at ascending sites of convection and sinks of fluid at descending regions of convection. With rigid surface plates the surface motion also has a toroidal component consisting of horizontal shearing and rotations about local vertical axes. Analysis of present plate motions by Hager and O'Connell (1978) showed

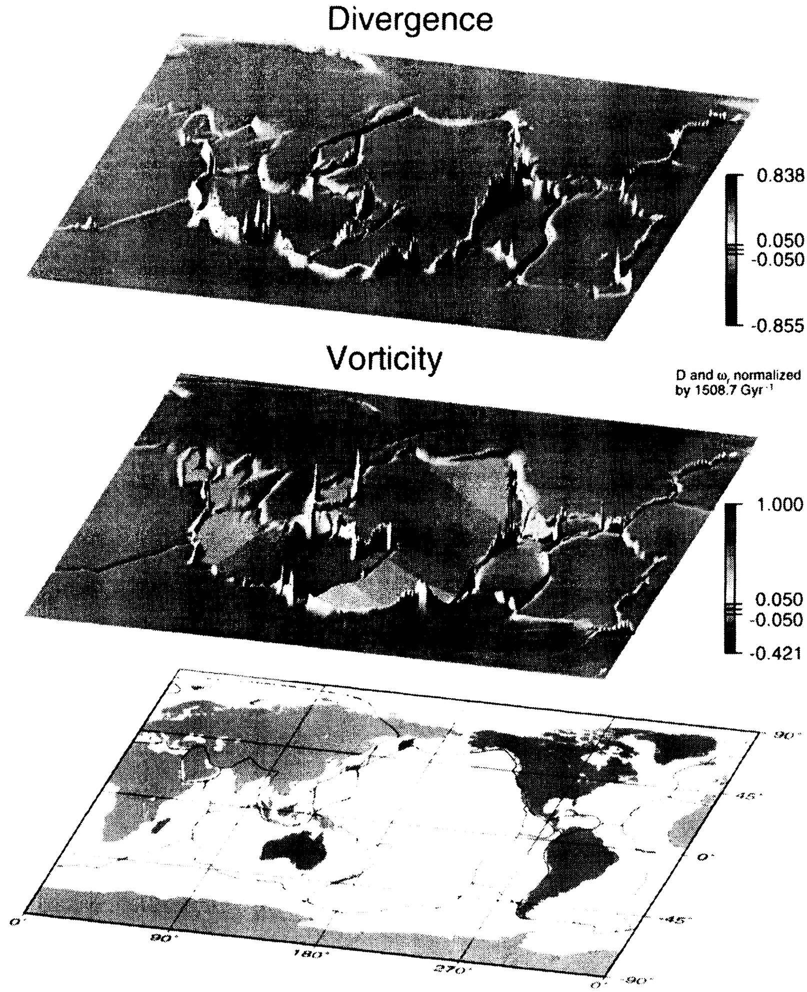


Figure 2.33. The pattern of horizontal divergence and the radial component of vorticity in present plate motions, calculated by Dumoulin et al. (1998).

that the kinetic energy of the toroidal part (strike–slip motion and oblique convergence) of present plate motions is nearly as large as the energy in the poloidal part (normal spreading and convergence).

Question 2.14: *How is toroidal motion generated in mantle convection?*

The origin of the toroidal kinetic energy and the manner in which mantle convection adapts to the presence of this component of kinetic energy are major unsolved problems. Figure 2.33 shows the fields of plate divergence (convergence) and radial vorticity calculated from present plate motions by Dumoulin et al. (1998). Positive and negative divergences are concentrated at ridges and trenches, respectively, as expected. The radial vorticity, a measure of shearing motions and the toroidal energy density, is also concentrated along plate margins, rather than being broadly distributed in plate interiors. This demonstrates that the toroidal motion is associated with plate boundary deformations – shear at transform faults and at oblique subduction zones – rather than spin of plates as a whole. This type of toroidal motion requires strongly nonlinear rheology at plate boundaries, where the energy is drawn from the poloidal motion, shown in the pattern of surface divergence. The question has been the subject of many papers (Gable et al., 1991; O'Connell et al., 1991; Olson and Bercovici, 1991; Ribe, 1992; Lithgow-Bertelloni et al., 1993; Bercovici, 1995a; Weinstein, 1998).

2.10 The Wilson Cycle and the Time Dependence of Plate Tectonics

Wilson (1966) proposed that continental drift is cyclic. In particular he proposed that oceans open and close; this is now known as the Wilson cycle and was based on the opening and closing of the Atlantic Ocean. The Wilson cycle, in its simplest form, is illustrated in Figure 2.34.

Question 2.15: How are accretional plate margins formed?

The first step in the Wilson cycle, illustrated in Figure 2.34, is the breakup of a continent. This occurs on continental rift zones. Present examples are the East African Rift system and the Rio Grande graben. These may or may not break apart to form future oceans. Aulacogens (triple junctions with three rifts connected at about 120°) are believed to play a key role in the initiation of rifting and the breakup of continents (Burke, 1977). Aulacogens are associated with lithospheric swells (Burke and Dewey, 1973). An example of a lithospheric swell on a continent is the Ethiopian swell on the East African Rift. An example of a triple junction is at the southern end of the Red Sea; the three arms are the Red Sea, the Gulf of Aden, and the East African Rift. When a continent opens, two of the rifts separate and become part of an ocean. The third rift aborts and is known as a "failed" arm. Examples of failed arms associated with the opening of the Atlantic Ocean are the St. Lawrence River Valley Rift and the Niger Rift in Africa.

Continental rifts are tensional failures of the continental lithosphere. Both active and passive mechanisms for continental rifting have been proposed (Turcotte and Emerman, 1983). The passive mechanism hypothesizes that the continental lithosphere fails under tensional stresses transmitted through the elastic lithosphere by plate margin forces such as trench pull. In this mechanism volcanism and uplift associated with rifting are secondary processes. The active mechanism hypothesizes that a mantle plume impinges on the base of the continental lithosphere causing volcanism and uplift. In this mechanism the tensional failure of the lithosphere is a secondary process.

The second step in the Wilson cycle is the opening of the ocean illustrated in Figure 2.34. The rift valley splits apart and oceanic crust is formed at an accretional plate boundary. The Red Sea is an example of the initial stages of the opening of an ocean, while the Atlantic Ocean is an example of a mature stage. The margins of an opening ocean are known as passive continental margins in contrast to active continental margins, where subduction is occurring.

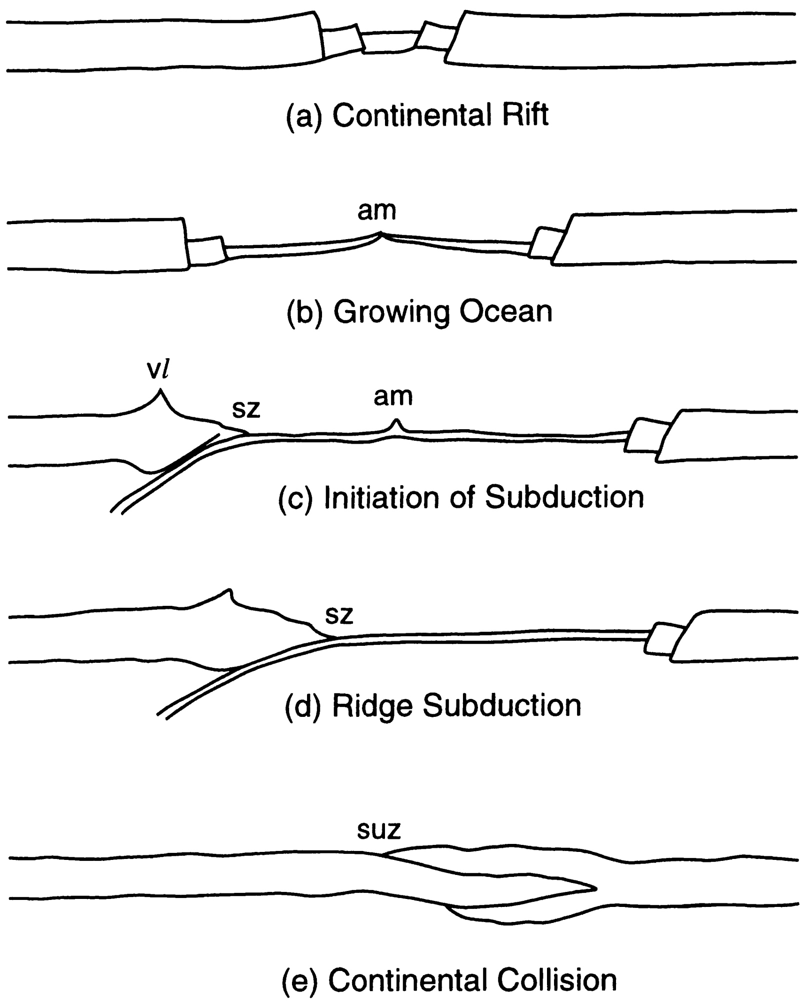


Figure 2.34. Illustration of the Wilson cycle. (a) Initiation of new ocean at a continental rift zone. (b) Opening of the ocean. am – accretional margin. (c) Initiation of subduction. sz – subduction zone, v*l* – volcanic line. (d) Ridge subduction. (e) Continental collision. suz – suture zone.

***Question 2.16:** How are subduction zones formed?*

The third step in the Wilson cycle is the initiation of subduction (Figure 2.34). A passive continental margin is a favored site for the initiation of subduction because it is already a zone of weakness established during rifting. The differential subsidence between aging seafloor and the continental lithosphere provides a source of stress. A further source of stress is the gravitational loading by continental sediments deposited at the passive margin (Cloetingh et al., 1984, 1989; Erickson, 1993; Erickson and Arkani-Hamed, 1993).

Several mechanisms have been proposed for the actual initiation of subduction. McKenzie (1977b) proposed failure of the lithosphere under compressional stress. Thrust faulting at a continental margin leads to the downthrusting of the oceanic lithosphere beneath the continental margin and the initiation of subduction. Turcotte et al. (1977), Kemp and Stevenson (1996), and Schubert and Zhang (1997) have proposed failure of the lithosphere under tension and the foundering of the oceanic lithosphere to generate a subduction zone. Faccenna et al. (1999) have discussed the initiation of subduction at a passive margin. Mueller and Phillips (1991) suggested that in some cases the creation of a new subduction zone could be triggered by the subduction of an aseismic ridge with thick and buoyant oceanic crust at an existing trench causing the cessation of further subduction there. Casey and Dewey (1984) have proposed a complex mechanism for the initiation of subduction involving accretional ridge segments and transform faults when the spreading direction shifts (see also Toth and Gurnis, 1998).

The fourth step in the Wilson cycle, illustrated in Figure 2.34, is ridge subduction. If the velocity of subduction is greater than the velocity of seafloor spreading, the ocean will close and eventually the accretional plate margin will be subducted. A number of authors have considered the thermal consequences of ridge subduction and have examined the geological record for effects on subduction zone volcanism and morphology (DeLong et al., 1978, 1979; Hsui, 1981). Ridge subduction played an important role in the recent geological evolution of the western United States and in the development of the San Andreas fault system (Atwater, 1970).

The fifth and final stage in the Wilson cycle, illustrated in Figure 2.34, is the continental collision that occurs when the ocean closes. This terminates the Wilson cycle. Continental collision is one of the primary mechanisms for the creation of mountains in the continents; the other is subduction (Dewey and Bird, 1970). The Himalayas and the Alps are examples of mountain belts caused by continental collisions, and the Andes is a mountain belt associated with subduction. The boundary between the two plates within the collision zone is known as a suture zone (Burke et al., 1977; Dewey, 1977).

The Himalayas are the result of the continental collision between the Indian subcontinent and Asia. This collision occurred about 45 Ma ago and has been continuing since. The initial collision resulted in a major global reorganization of plate motions that is best documented by the bend in the Hawaiian-Emperor seamount chain seen in Figures 2.6 and 2.7.

Many models have been proposed for the deformation that has resulted in the elevation of the Himalayas and the Tibetan Plateau. A viscous "snowplow" model has been suggested (England and Houseman, 1986). Another is the "flake tectonic" model (Oxburgh, 1972) in which the upper Asian continental crust overrides the Indian continental crust, the lower Asian crust and mantle having been previously delaminated. An additional model has been proposed by Molnar and Tapponnier (1975, 1978) in which India acts as an "indentor" that has driven Asian crust eastward into southeast Asia. Lenardic and Kaula (1995a) argue against the fluid "snowplow" model and in favor of a delayed removal of thickened lithosphere. In fact, the actual deformation associated with continental collision may be some complex combination of all these models.

There is evidence that plate tectonic processes are cyclic and correlate with the Wilson cycle. The evidence, which includes variations in sea level, in the strontium isotope ratios of seawater, and in continental volcanism associated with subduction of continental crust, has been summarized by Worsley et al. (1984), Nicolaysen (1985), Veevers (1989), and Unrug (1992). The summary cartoon given by Veevers (1989), partially reproduced in Figure 2.35, shows cyclic variability in the Paleozoic to recent geological record. Rates of crustal subduction and arc magma production increased during times of continental dispersal and decreased during times of continental aggregation; this is indicated by the correlation of low stand of

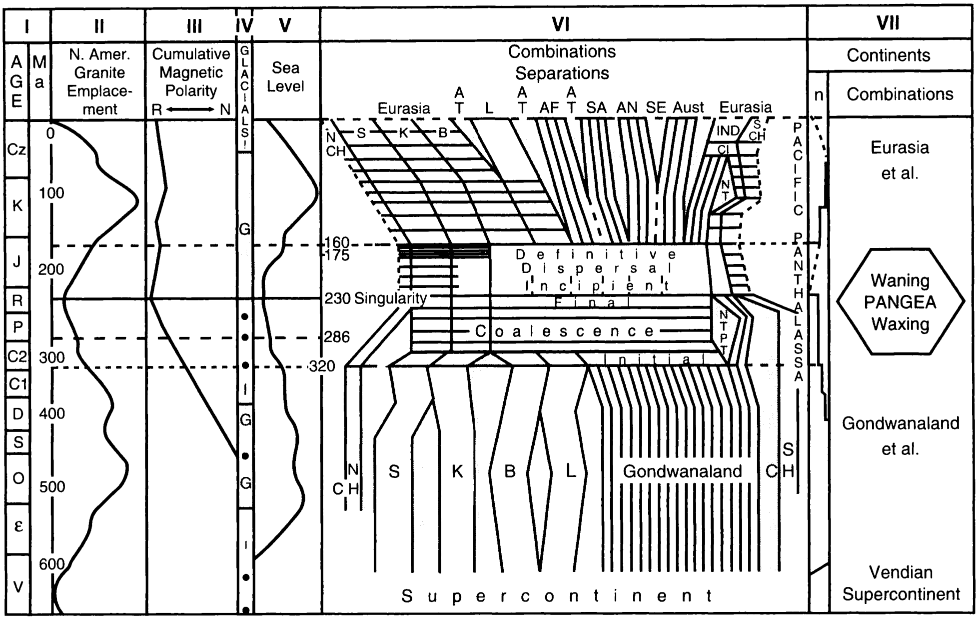


Figure 2.35. Illustration of the plate tectonic cycle (after Veevers, 1989). I. Age before present and geological epochs. II. Rates of emplacement of granitic rocks in North America. III. Cumulative polarity of the Earth's magnetic field; N normal and R reversed. IV. Periods of glaciations. V. Sea level variations. High stands of sea level correspond to high rates of seafloor spreading, which in turn correspond to high rates of subduction and production of granitic rocks. VI. Illustration of coalescence and dispersal of continental blocks. VII. Number of continents n.

sea level and implied low ridge volume. A minimum in continental granitoid production rates in North America (South America is similar) during the time of Pangea has also been noted (Figure 2.35).

The period 500–440 Ma (Ordovician) featured the breakup of the Vendian supercontinent. This was a time of high sea level, extensive island arc volcanism, and large compressional and tensional stresses in the continental crust. The period 380–340 Ma (late Devonian to lower Carboniferous) was a time of continental coalescence and collisions between large continental blocks. Sea level and oceanic spreading rates were decreasing and arc deformation was associated with major collisions during this period. The interval 340–260 Ma (upper Carboniferous to Permian) represented the peak of continental coalescence. There were collisions of large continental blocks and the sweeping up of small blocks. Much of the growing supercontinent was surrounded by subduction zones. Sea level was dropping and there were extensive regions of high continental relief associated with continental collisions. During the period 260–225 Myr before present (late Permian to Triassic) the Pangea supercontinent was at its maximum size. Rates of seafloor spreading and subduction reached their lowest levels, sea level was low, and the oceanic crust was relatively old. During this period clear evidence for subduction is missing along much of the Pangean margin.

The breakup of Pangea began in the early Triassic and continued into the Jurassic and Cretaceous. As emphasized by Veevers (1989), this period was marked by important flood basalts. Larson and Olson (1991) have shown that ocean plateau production occurred during the Cretaceous. As spreading in the ocean basins reinitiated, important subduction zones

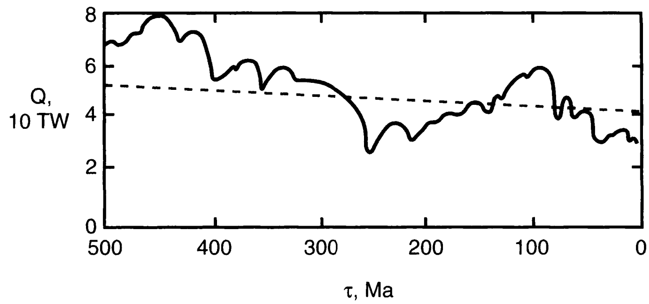


Figure 2.36. Global heat flow Q as a function of time before the present for the last 500 Myr. Solid curve – heat loss inferred from sea level variations. Dashed line – heat loss based on the decay of radioactive elements and secular cooling.

reappeared in oceanic arcs and along continental margins. Continental flood basalts, whose origin could lie in the partial melting of plume heads upwelling near the surface (Richards et al., 1989), were associated with continental breakup (Burke and Dewey, 1973). The positions of many plumes correlate with the position of Pangea at the time of its breakup (Anderson, 1982; Ashwal and Burke, 1989). Magmatism of both oceanic and continental arc type occurred frequently. A number of small terranes can be recognized in the geologic record and deformation associated with very active plates was common. The high stand of sea level during the Cretaceous is well documented and can be associated with excess ridge volume and a large plate tectonic flux, i.e., the area rate of plate formation was high (Hays and Pitman, 1973). This period not only coincides with a peak in ocean plateau formation, it is also coincident with the Cretaceous magnetic polarity superchron, suggesting an association with the core as well (Larson and Olson, 1991).

Fluctuations in sea level over geologic time have been used to deduce variations in the Earth's global heat flow Q (Hallam, 1977, 1992; Turcotte and Burke, 1978), as shown by the solid curve in Figure 2.36. The dashed line is the heat flow expected from the decay of radioactive elements and the secular cooling of the Earth. The plate tectonic cycle, as expressed in sea level variations, is associated with about a 30% fluctuation in global heat flow. Since the subduction of cold oceanic lithosphere is responsible for some 80% or more of the heat loss from the interior of the Earth, variations in the plate flux may influence (or be influenced by) mantle temperature and dynamics as follows (Figure 2.37):

(i) When continents are dispersed, the plate flux is high so that sea level is high, seawater $^{87}Sr/^{86}Sr$ is low, and Andean-type volcanism is extensive. The high plate flux cools the mantle and may impede the formation of mantle plumes by emplacing relatively cold material on the core–mantle boundary, the likely site of plume formation. The net result is a reduction in the mantle plume flux.

(ii) The breakup of continents can be attributed to a high plume flux. With a reduction in this flux the rate of continental breakup decreases. Due to simple kinematics the dispersed continents coalesce to form a supercontinent. A consequence of the resulting continental collisions and the reduction in the number of oceans is a reduction in the plate flux.

(iii) With a single supercontinent and a low plate flux, the mantle heats up due to the decay of radioactive isotopes. The increase in mantle temperature and the warming near the

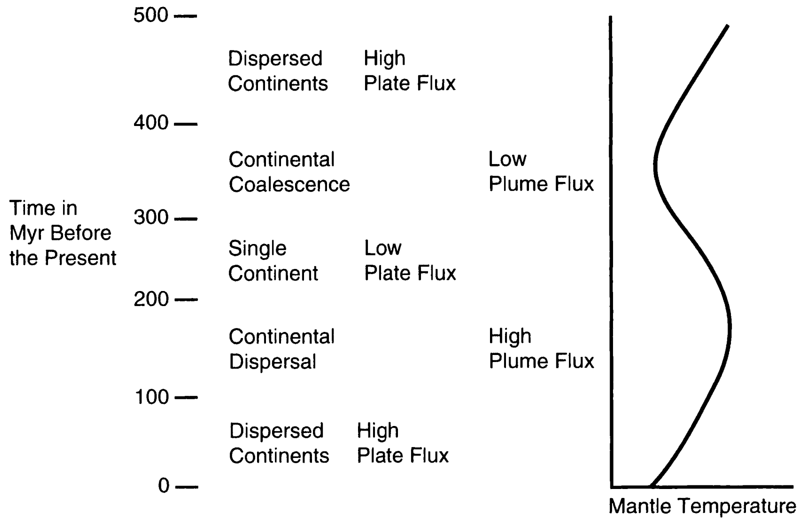


Figure 2.37. A scenario for the cyclic behavior of plate tectonics over the last 500 Myr.

core–mantle boundary leads to an increase in the plume flux and the breakup of the supercontinent (Yale and Carpenter, 1998).

(iv) The cycle repeats.

***Question 2.17:** Is the temporal variability of plate tectonics stochastic or is it driven by episodicity in mantle convection?*

Several plausible explanations for temporal variability in plate tectonics have been advanced, including the one just described. A second explanation is that it is completely stochastic and is associated with continental collisions and randomly generated plumes (Duncan and Turcotte, 1994). A third explanation is that mantle convection exhibits a quasi-periodicity or episodicity perhaps associated with mantle overturns that generate surface orogenic events (Stein and Hofmann, 1994; Condie, 1998). All of these proposed causes remain highly speculative, and there are very few critical observations with which to test them. But at least we can say that temporal variability in plate tectonics is fully expected on theoretical grounds, since the convection in the mantle that drives plate tectonics is unquestionably time-dependent. The nature of time dependence in mantle convection, its underlying causes, and its consequences for the thermal and tectonic history of the Earth are discussed in Chapters 10 and 13.