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General Features of Mantle Plumes

A mantle plume is generally considered to be a blob of relatively hot, low-density mantle that rises because of its buoyancy. The existence of mantle plumes in the Earth was first suggested by J. Tuzo Wilson (1963) as an explanation of oceanic island chains, such as the Hawaiian–Emperor chain, that change progressively in age along the chain. Wilson proposed that a lithospheric plate moves across a fixed hotspot (the mantle plume), volcanism is recorded as a linear array of volcanic seamounts and islands parallel to the direction in which the plate is moving. Morgan (1971) championed the idea of mantle plumes, suggesting that flood basalts formed by melting of plume heads, whereas hotspot volcanic chains were derived from partial melting of plume tails. He also showed that closely spaced hotspots on the same plate had not moved significantly relative to each other and suggested this was evidence that the plumes had come from the core–mantle boundary (Morgan 1972). Morgan noted that some hotspot tracks, like the Mascarene–Chagos–Laccadive track in the Indian Ocean, are traceable to flood basalts and can be used to reconstruct paths of opening ocean basins. Richards, Duncan, and Courtillot (1989) recognized at least 10 flood basalt–hotspot track pairs that formed from mantle plumes in the last 250 Myr.

The first laboratory experiments aimed at understanding mantle plumes better were those of Whitehead and Luther (1975), who showed that plume viscosity has an important effect on the shape of a plume. If a plume has a viscosity greater than its surroundings, it rises as a finger, whereas if it has a lower viscosity, it rises in a mushroom shape with a distinct head and tail. The tail contains a hot fluid that “feeds” the head as it buoyantly rises. Loper and Stacey (1983) developed a theory of flow in plume tails for a case in which the viscosity of a plume is strongly temperature dependent. Because the tail is hot, it has a relatively low viscosity and is quite narrow (≈100 km across). Olson and Singer (1985) developed a theory for the ascent of plume heads that are compositionally distinct from surrounding mantle. They also studied the behavior of plume tails during horizontal shear caused by convective currents. Griffiths and Campbell (1990) were the first to confirm, by experiment and theory, the existence of thermal plume heads and tails and to distinguish between thermal and compositional
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Plume Nomenclature

Numerous terms have been applied to mantle plumes, and there is confusion in the use of these terms. Although general agreement has not yet been reached, it is important to standardize the usage for this book. As noted previously, a mantle plume is a buoyant mass of material in the mantle that rises because of its buoyancy. On reaching the base of the lithosphere, plumes spread laterally. As suggested by the areal extent of some flood basalts, which are derived by partial melting of plumes, plume heads may reach diameters of 500 to 3000 km (Hill et al. 1992). Plume tails, on the other hand, are typically 100 to 200 km in diameter. Large hotspots are the surface manifestation of mantle plumes and are focused zones of melting. They are characterized by high heat flow, variable topographic highs depending on plume depth, and active volcanism. The term superplume is used herein to describe plume heads 1500–3000 km in diameter. Expressed in terms of the volume of plume-derived basalt flows, superplumes give rise to erupted volumes exceeding $0.5 \times 10^6 \text{ km}^3$. The term diapir has been used to describe some mantle plumes. Herein, mantle diapir is used to describe a small mantle plume (<300 km across) that has lost its tail and thus ceases to grow (Herrick 1999). Diapirs may be produced in descending slabs or in the mantle wedge above descending slabs in response to localized thermal gradients. Alternatively, they may form anywhere in the mantle if it is convecting in a hard turbulent regime (Yuen et al. 1993). Another possible production mode is upward “budding” along the tops of superplumes (Sleep 1990).

Subduction of lithospheric slabs into the mantle requires a balancing upward flow. Because slabs are relatively cool, they cool the adjacent mantle as they descend. On the other hand, the mantle between subduction zones, where return flow occurs, is relatively warm and will slowly expand and rise. Larson (1991a) originally used the term superplume to describe the large mantle upwelling in the Pacific basin, and some investigators have continued this usage (Maruyama 1994). However, we will retain the more widely employed term mantle upwelling to describe these large regions of rising warm mantle between subduction zones. If a large lithospheric plate “protects” a large volume of mantle from the cooling effects of subduction, which happens beneath supercontinents and large ocean basins, a large mantle upwelling may be generated beneath the plate. Two such upwellings occur in the mantle today, one beneath Africa
and one in the South Pacific. Such upwellings are typically more than 10,000 km in diameter and may contain many mantle plumes and hotspots.

Plumes may elevate the lithosphere by several hundred meters, producing broad, roughly circular uplifts known as *swells*. In oceanic lithosphere, which is relatively thin, swells may reach diameters of 2000 km with 500–1000 m of relief (Crough 1983; McNutt 1998). Mantle upwellings also raise the lithosphere, producing large swells known as *superswells*, which are more than 10,000 km across (McNutt 1998).

**Internal Structure of the Mantle**

*An Overview*

Before beginning a detailed discussion of mantle plumes, we need to review the internal structure of the mantle as revealed by seismology. Our knowledge of the Earth’s interior comes chiefly from compressional (P-wave) and shear (S-wave) waves that pass through the Earth in response to earthquakes. Seismic wave velocities vary with pressure (depth), temperature, mineralogy, chemical composition, and the amount of melt present. Three first-order seismic discontinuities that reflect changes in composition or mineralogy divide the Earth into crust, mantle, and core (Fig. 1.1): the Mohorovicic discontinuity, or Moho, defines the base of the crust; the large decrease in seismic wave velocity at the base of the mantle (2900 km) defines the core–mantle interface; and at about 5200 km, a small increase in velocity is the inner-core boundary. Smaller but important velocity changes at 50–200, 410, and 660 km provide a basis for further subdivision of the mantle.

![Diagram](image-url)

Figure 1.1. Internal structure of the Earth as represented by compressional ($V_P$) and shear wave velocity ($V_S$) and calculated density ($\rho$). Calculated temperature distributions for layered ($T_L$) and whole-mantle ($T_W$) convection. Also shown are the two major thermal boundary layers: the lithosphere and D". After Condie (1997a).
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The major regions of the Earth are summarized as follows with reference to Figure 1.1:

1. The crust consists of the region above the Moho and ranges in thickness from about 3 km at some ocean ridges to about 70 km in collisional orogens.

2. The lithosphere (50–300 km thick) is the strong outer layer of the Earth, including the crust, that reacts to many stresses as a brittle solid. The lithosphere is a mechanical boundary layer that is broken into plates, which are formed at ocean ridges and descend into the mantle at subduction zones. The asthenosphere, extending from the base of the lithosphere to the 660-km discontinuity, is by comparison a weak layer that deforms by creep. A region of low seismic-wave velocity and high attenuation of seismic-wave energy, a low-velocity zone (LVZ), occurs at the top of the asthenosphere and is from 50- to 100-km thick. Significant lateral variations in density and in seismic-wave velocity are common in the mantle at depths of less than 400 km.

3. The upper mantle extends from the Moho to the 660-km seismic discontinuity. It includes the lower part of the lithosphere, the asthenosphere, and two seismic discontinuities that are caused by important solid-state transformations: olivine to wadsleyite at 410 km, and spinel to perovskite + magnesiowustite at 660 km.

4. The lower mantle extends from the 660-km seismic discontinuity to the core–mantle boundary at 2900 km. Between 200 and 250 km above the core–mantle interface, a flattening of velocity and density gradients occurs in a region known as the D″ layer named after the seismic wave used to define the layer. The lower mantle is also referred to as the mesosphere – a region that is strong but relatively passive in terms of deformational processes.

5. The outer core will not transmit S waves and is interpreted as a liquid composed chiefly of molten iron and nickel.

6. The inner core, which extends from the 5200-km discontinuity to the center of the Earth, transmits S waves (although at very low velocities), suggesting that it is near the melting point.

There are two boundary layers in the Earth with steep thermal gradients: the lithosphere at the surface and the D″ layer just above the core (Fig. 1.1). These layers play an extremely important role in cooling and convection in the Earth. In addition, the steep thermal gradient in the D″ layer may be the site at which most mantle plumes are generated.

Considerable uncertainty exists regarding the temperature distribution in the Earth. It is dependent on such features of the Earth’s history as (1) the initial temperature distribution, (2) the heat contributed by large impacting bodies, (3) the amount of heat generated as a function of both depth and time, (4) convective and conductive heat loss, and (5) the process of core formation. Most estimates of temperature distribution in the Earth are based on one or a combination of two approaches: models of the Earth’s thermal history involving various mechanisms of core formation and models involving redistribution of radioactive heat sources in the Earth by melting and convection processes.

Estimates using various models seem to converge on a temperature at the core–mantle interface of about 4500 ± 500 °C and a temperature at the center of the core of
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6700 to 7000 °C. Two examples of calculated temperature distributions in the Earth are shown in Figure 1.1. Both show significant gradients in temperature in the lithosphere. The layered convection model \((T_L)\) also shows a large temperature change near the 660-km discontinuity because this is the boundary between shallow and deep convection systems in this model. The temperature distribution for whole-mantle convection \((T_w)\), which is preferred by most investigators, shows a steep thermal gradient in the D” layer but no changes at the discontinuities in the upper mantle.

Convection is the dominant mode of heat transfer in the asthenosphere and mesosphere, where an adiabatic gradient is maintained, and thus temperature increases at a very slow rate with increasing depth. On the other hand, conduction is the main way heat is lost from the lithosphere, and temperatures change rapidly with depth and tectonic setting. In this respect, the lithosphere is both a mechanical and a thermal boundary layer in the Earth.

The Lithosphere

The oceanic lithosphere, where thickness is controlled by cooling, can be considered the upper boundary layer with a conductive temperature gradient that overlies a convecting adiabatic interior. Because thickness is dependent on temperature and age, it is sometimes referred to as the thermal lithosphere. Asthenosphere can be converted to oceanic lithosphere simply by cooling. The oceanic lithosphere begins life at ocean ridges as restite left when ocean-ridge basaltic magma is extracted. The progressive thickening of the lithosphere continues until about 70 Myr, and afterwards it remains relatively constant in thickness until subduction. Convective erosion at the base of the oceanic lithosphere may be responsible for maintaining this constant depth.

Unlike oceanic lithosphere, continental lithosphere has a complicated history that probably involves more than one mechanism by which it forms and grows (Condie 1997a). The post-Archean subcontinental lithosphere includes some combination of accreted asthenosphere, remnants of mantle plumes, and remnants of mantle wedges that originally formed above descending plates. Seismic reflection results also suggest that at least some of the continental lithosphere comprises remnants of partially subducted oceanic lithosphere. In northern Scotland, for example, dipping reflectors in the lower lithosphere are thought to represent fragments of now eclogitic oceanic crust (implying deep burial) – a relic of Paleozoic oceanic subduction (Warner et al. 1996). Although basal plume accretion may also have been important in the formation of the thick Archean (>2.5 Ga) lithosphere (Campbell and Griffiths 1992), buoyant subduction must have been important in the Archean as well, and thus oceanic plates may have been plastered beneath the continents, contributing to lithospheric thickening (Condie 1997a).

The Low-Velocity Zone

The seismic low-velocity zone (LVZ) in the upper mantle is characterized by low seismic wave velocities and high electrical conductivity (Condie 1997a). The bottom of the LVZ, sometimes referred to as the Lehmann discontinuity, has been identified
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Figure 1.2. P-wave velocity and specific attenuation factor (Q) distribution in the upper mantle. Q varies inversely with seismic wave attenuation such that low values of Q show the greatest attenuation. After Condie (1997a).

from the study of surface wave and S-wave data in continental areas and occurs at depths of 180–220 km (Fig. 1.2) (Caherty and Jordan 1995). The LVZ plays a major role in plate tectonics, providing a relatively weak, low-viscosity region upon which lithospheric plates move.

Because of the dramatic drop in S-wave velocity and increase in attenuation of seismic energy, interconnected melt along grain boundaries must contribute to producing the LVZ. The probable importance of incipient melting is attested to by the high surface heat flow observed when the LVZ reaches shallow depths, such as beneath ocean ridges and in continental rifts. Experimental results show that incipient melting in the LVZ requires a minor amount of water to depress silicate melting points (Wyllie 1971). With only 0.05–0.1% water in the mantle, partial melting of garnet herzolite occurs in the appropriate depth range for the LVZ. The source of water in the upper mantle may be from the breakdown of minor phases that contain water such as amphiboles and micas. The theory of elastic wave velocities in two-phase materials indicates that only 1% melt is required to produce the lowest S-wave velocities measured in the LVZ. If, however, melt fractions are inter-connected by a network of tubes along grain boundaries, the amount of melting may exceed 5% (Marko 1980). The downward termination of the LVZ probably reflects a combination of the depth at which geotherms pass below the mantle solidus and a rapid decrease in the amount of water available. The width or even the existence of the LVZ depends on the steepness of the geotherms. With steep
geotherms, like those characteristic of ocean ridges and continental rifts, the range of penetration of the mantle solidus is large, and hence the LVZ is relatively wide (100–200 km). The gentle geotherms beneath continental platforms, which show a narrow range of intersection with the hydrated mantle solidus, produce a thin or poorly defined LVZ. Beneath Archean shields, geotherms generally do not intersect the mantle solidus; hence, there is no LVZ (Fig. 1.2).

**The 410-km Discontinuity**

High-pressure experimental studies show that the 410-km seismic discontinuity reflects the breakdown of Mg-rich olivine to the high-pressure Mg–silicate phase wadsleyite at about 14 GPa (Fig. 1.3). Mantle olivine (Fo90) completely transforms to higher density wadsleyite over a less than 300 MPa pressure range at appropriate temperatures for the 410-km discontinuity (Ito and Stixrude 1992). This pressure range is in good agreement with the less than 10-km width of the 410-km discontinuity deduced from seismic data (Vidale et al. 1995). The approximately 6% increase in density observed at this discontinuity suggests that olivine composes 40–60% of the upper mantle as it does in garnet herzolite xenoliths derived from the upper mantle.

High-pressure experimental data indicate that at depths of 350–450 km, both clinopyroxene transforms into a garnet-structured mineral known as majorite, involving a density increase of about 6% (Christensen 1995). This transition has been petrologically observed as pyroxene exsolution laminae in garnet in mantle xenoliths derived from the Archean lithosphere at depths of 300–400 km (Haggerty and Sautter 1990). It is probable that the increase in velocity gradient beginning at 350 km and leading up to the 410-km discontinuity is caused by the majorite transformation.

Experimental data also indicate that wadsleyite transforms to a more dense spinel-structured phase at depths of 500–550 km. This mineral, referred to as Mg–spinel, has

![Figure 1.3. High-pressure phase relations for Mg_2SiO_4 in the mantle. After Christensen (1995).](image-url)
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the same composition as Mg-rich olivine but the crystallographic structure of spinel. The small density change (≈2%) associated with the phase change, however, does not generally produce a resolvable seismic discontinuity.

The 660-km Discontinuity

One of the most important considerations related to the style of mantle convection in the Earth is the nature of the 660-km discontinuity (Fig. 1.2). If descending slabs and mantle plumes cannot penetrate this boundary, two-layer mantle convection is favored in which the 660-km discontinuity represents the base of the upper layer. Large increases in both seismic-wave velocity (5−7%) and density (8%) occur at this boundary. High-frequency seismic waves reflected at the boundary suggest that it has a width of only about 5 km but has up to 20 km of relief over distances of hundreds to thousands of kilometers (Wood 1995).

As with the 410-km discontinuity, it is likely that a phase change in Mg$_2$SiO$_4$ is responsible for the 660-km discontinuity (Christensen 1995). High-pressure experimental results indicate that spinel transforms to a mixture of perovskite and magnesiowustite at a pressure of about 23 GPa, which can account for the seismic velocity and density increases at this boundary if the rock contains 50−60% Mg-spinel. Mg-perovskite and magnesiowustite are extremely high-density minerals and appear to compose most of the lower mantle.

Unlike the shallower phase transitions, the spinel–perovskite transition has a negative Clapeyron slope in P–T space (Fig. 1.3), and thus the transition may impede slabs from sinking into the deep mantle and also make it difficult for mantle plumes to rise into the upper mantle (Davies 1999). The latent heat associated with phase transitions in descending slabs and rising plumes can deflect phase transitions to shallower depths for positive Clapeyron slopes and to greater depths for negative Clapeyron slopes (Liu 1994) (Fig. 1.4). For a positive slope, like the olivine–wadsleyite transition, the elevated region of the denser phase exerts a strong downward pull on the slab or upward pull on a plume, thus helping drive convection. In contrast, for a negative slope, like the spinel–perovskite transition, the low-density phase is depressed, enhancing a slab’s buoyancy and resisting further sinking of the slab. This same reaction may retard a rising

Figure 1.4. Deflection of phase boundaries in subducting slabs (light phase) for reactions with positive and negative Clapeyron slopes.
plume. However, if upward rise of a plume is slow and the temperature is high enough for the phase reaction to proceed, the plume may transform as it rises. Computer models by Davies (1995) suggest that stiff slabs can penetrate the boundary more readily than plume heads, and plume tails are the least able to penetrate it. Some geophysicists have suggested that slabs may locally accumulate at the 660-km discontinuity, culminating in occasional “avalanches” of slabs into the lower mantle (Tackley et al. 1994).

Seismic tomographic images of descending slabs provide an important constraint on the depth of penetration into the mantle. These images indicate that, although some slabs may be delayed at the 660-km discontinuity, all modern slabs eventually sink into the lower mantle. Thus, there is no evidence for layered convection in the Earth in terms of slab distributions in the mantle (van der Hilst et al. 1997).

The Lower Mantle

General Features

High-pressure experimental studies clearly suggest that Mg–perovskite is the dominant phase in the lower mantle. However, it is not clear if the seismic properties of the lower mantle require a change in major element composition at the 660-km phase transition (Wang et al. 1994). Results allow, but do not require, the Fe/Mg ratio of the lower mantle to be greater than that of the upper mantle. If this were the case, the greater density of Fe–perovskite would greatly limit the mass flux across the 660-km discontinuity, and the maintenance of such a chemical difference would favor layered convection.

The isostatic rebound of continents following Pleistocene glaciation, together with gravity data, indicates that the viscosity of the mantle increases with depth by two orders of magnitude and that the largest jump occurs at the 660-km discontinuity (Condie 1997a). This conclusion is in agreement with other geophysical and geochemical observations. For instance, although mantle plumes move upwards relatively fast, it would be impossible for them to survive convective currents in the upper mantle unless they were anchored in a “stiff” lower mantle. Also, only a mantle of relatively high viscosity at depth can account for the small number (two today) of large mantle upwellings. The isolation of geochemical domains in the mantle for billions of years (Chapter 5) can be achieved with a stiff lower mantle that resists mixing.

The D″ Layer

The D″ layer is a region of the mantle within a few hundred kilometers of the core where seismic velocity gradients are anomalously low (Loper and Lay 1995; Montague and Kellogg 2000). Calculations indicate that a relatively small temperature gradient (1–3 °C/km) is necessary to conduct heat from the core into the D″ layer. Because the core diffracts seismic waves, spatial resolution in this layer is poor, and details of its structure are not well known. However, initial seismic results indicate that D″ is a complex region that is vertically and laterally heterogeneous (Kendall and Silver 1996; Lay et al. 1998). Data seem to be equally consistent with a sharp interface between 100 and 300 km above the core–mantle boundary, or small-scale discontinuities that scatter
seismic waves near the boundary (Fig. 1.5), or both. Estimates of the thickness of the D'' layer suggest that it ranges from approximately 100 to 500 km. Despite the poor resolution, large-scale lateral heterogeneities can be recognized in D'' (Lay et al. 1998; Sidorin et al. 1999). For example, regions beneath circum-Pacific subduction zones have anomalously fast P and S waves, which are interpreted by many to represent lithospheric slabs that have sunk to the base of the mantle. Slow velocities in D'' occur beneath the Central Pacific and correlate both with the surface and core–mantle boundary geoid anomalies and a greater concentration of hotspots (Chapter 2).

There are three possible contributions to the complex seismic structures seen in D'': temperature variations, compositional changes, and mineralogical phase changes. Temperature variations appear to be caused chiefly by the sinking of slabs into D'' (a cooling effect that produces relatively fast velocities) and heat released from the core (causing slow velocities). If a significant gradient in viscosity occurs in D'', convection may take place within the boundary layer (Montague and Kellogg 2000). Mixing of molten iron from the core with high-pressure silicates can lead to compositional changes with corresponding velocity changes. Experiments have shown, for instance, that when liquid iron comes in contact with silicate perovskite at high pressures, these substances react vigorously to produce a mixture of Mg–perovskite, a high-pressure silica polymorph, wustite (FeO), and Fe silicide (FeSi) (Wyession et al. 1998). These experiments also suggest that liquid iron in the outer core will seep into D'' by capillary action and affect a region extending for hundreds of meters above the core–mantle boundary. Recent seismic studies reveal the presence of fuzzy zones at the core–mantle interface, which could be due to intense chemical and physical interactions of the core with mantle silicates (Garnero and Jeanloz 2000). Phase changes, such as the possible breakdown of Mg–perovskite to magnesiowustite and silica, provide the best explanation for the sharp velocity increase seen at about 2600 km (Fig. 1.5). Although chemical segregation may