Structure and characteristics of the earth's mantle

Environment, Earth

The interior of Earth, similar to the other terrestrial planets, is divided into layers of different composition. The mantle is a layer between the crust and the outer core. Earth's mantle is a silicate rocky shell with an average thickness of 2, 886 kilometers (1, 793 mi). The mantle makes up about 84% of Earth's volume. It is predominantly solid but in geological time it behaves as a very viscous fluid. The mantle encloses the hot core rich in iron and nickel, which makes up about 15% of Earth's volume. [14] Past episodes of melting and volcanism at the shallower levels of the mantle have produced a thin crust of crystallized melt products near the surface. Information about the structure and composition of the mantle has been obtained from geophysical investigation and from direct geoscientific analyses of Earth mantle-derived xenoliths and mantle that has been exposed by mid-oceanic ridge spreading.

Two main zones are distinguished in the upper mantle: the inner asthenosphere composed of plastic flowing rock of varying thickness, on average about 200 km (120 mi) thick, and the lowermost part of the lithosphere composed of rigid rock about 50 to 120 km (31 to 75 mi) thick. [17A thin crust, the upper part of the lithosphere, surrounds the mantle and is about 5 to 75 km (3. 1 to 46. 6 mi) thick. Recent analysis of hydrous ringwoodite from the mantle suggests that there is between one and three times as much water in the transition zone between the lower and upper mantle than in all the world's oceans combined. In some places under the ocean the mantle is actually exposed on the surface of Earth. There are also a few places on land where mantle rock has been pushed to the surface by tectonic activity, most notably the Tablelands region of Gros Morne National

Park in the Canadian province of Newfoundland and Labrador and Zabargad Island (St. John's Island) in the Red Sea.

Structure

The mantle is divided into sections which are based upon results from seismology. These layers (and their thicknesses/depths) are the following: the upper mantle (starting at the Moho, or base of the crust around 7 to 35 km (4. 3 to 21. 7 mi) downward to 410 km (250 mi)), the transition zone (410–660 km or 250–410 mi), the lower mantle (660–2, 891 km or 410–1, 796 mi), and anomalous core–mantle boundary with a variable thickness (on average ~200 km (120 mi) thick). The top of the mantle is defined by a sudden increase in seismic velocity, which was first noted by Andrija Mohorovičić in 1909; this boundary is now referred to as the Mohorovičić discontinuity or " Moho". The uppermost mantle plus overlying crust are relatively rigid and form the lithosphere, an irregular layer with a maximum thickness of perhaps 200 km (120 mi). Below the lithosphere the upper mantle becomes notably more plastic. In some regions below the lithosphere, the seismic shear velocity is reduced; this so-called low-velocity zone (LVZ) extends down to a depth of several hundred km. Inge Lehmann discovered a seismic discontinuity at about 220 km (140 mi) depth; although this discontinuity has been found in other studies, it is not known whether the discontinuity is ubiquitous. The transition zone is an area of great complexity; it physically separates the upper and lower mantle. Very little is known about the lower mantle apart from that it appears to be relatively seismically homogeneous. The D" layer at the core–mantle boundary separates the mantle from the core. In 2015, research using gravitational data from GRACE satellites and the long wavelength nonhydrostatic geoid indicated viscosity increases by a factor of ten to 150 about 1, 000 kilometers (620 mi) below earth's surface; separate research also indicates sinking tectonic plates stall at this depth, leading Robert van der Hilst to speculate " In terms of structure and dynamics, 1, 000 kilometers could be more important" (than the currently accepted 660 km depth upper – lower division). The lower mantle also contains some discontinuous zones, called " thermochemical piles" which have been interpreted as either thermally differentiated, upwellings bringing warmer material towards the surface, or as chemically differentiated material. A principal source of the heat that drives plate tectonics is the radioactive decay of uranium, thorium, and potassium in Earth's crust and mantle.

Characteristics

The mantle differs substantially from the crust in its mechanical properties as the direct consequence of the difference in composition (expressed as different mineralogy). The distinction between crust and mantle is based on chemistry, rock types, rheology and seismic characteristics. The crust is a solidification product of mantle derived melts, expressed as various degrees of partial melting products during geologic time. Partial melting of mantle material is believed to cause incompatible elements to separate from the mantle, with less dense material floating upward through pore spaces, cracks, or fissures that would subsequently cool and solidify at the surface. Typical mantle rocks have a higher magnesium to iron ratio and a smaller proportion of silicon and aluminum than the crust. This behavior is also predicted by experiments that partly melt rocks thought to be representative of Earth’s mantle.

Mapping the interior of the Earthwith earthquake waves. Mantle rocks shallower than about 410 km (250 mi) depth consist mostly of olivine, pyroxenes, spinel-structure minerals, and garnet; typical rock types are thought to be peridotite, dunite (olivine-rich peridotite), and eclogite. Between about 400 km (250 mi) and 650 km (400 mi) depth, olivine is not stable and is replaced by high pressure polymorphs with approximately the same composition: one polymorph is wadsleyite (also called beta-spinel type), and the other is ringwoodite (a mineral with the gamma-spinel structure). Below about 650 km (400 mi), all of the minerals of the upper mantle begin to become unstable. The most abundant minerals present, the silicate perovskites, have structures (but not compositions) like that of the mineral perovskite followed by the magnesium/iron oxide ferropericlase. The changes in mineralogy at about 400 and 650 km (250 and 400 mi) yield distinctive signatures in seismic records of the Earth’s interior, and like the moho, are readily detected using seismic waves. These changes in mineralogy may influence mantle convection, as they result in density changes and they may absorb or release latent heat as well as depress or elevate the depth of the polymorphic phase transitions for regions of different temperatures. The changes in mineralogy with depth have been investigated by laboratory experiments that duplicate high mantle pressures, such as those using the diamond anvil.

The inner core is solid, the outer core is liquid, and the mantle solid/plastic. This is because of the relative melting points of the different layers (nickel–iron core, silicate crust and mantle) and the increase in temperature and pressure as depth increases. At the surface both nickel–iron alloys and silicates are sufficiently cool to be solid. In the upper mantle, the silicates are generally solid (localized regions with small amounts of melt exist); however, as the upper mantle is both hot and under relatively little pressure, the rock in the upper mantle has a relatively low viscosity. In contrast, the lower mantle is under tremendous pressure and therefore has a higher viscosity than the upper mantle. The metallic nickel–iron outer core is liquid because of the high temperature, despite the high pressure. As the pressure increases, the nickel–iron inner core becomes solid because the melting point of iron increases dramatically at these high pressures.

Temperature

In the mantle, temperatures range between 500 to 900 °C (932 to 1, 652 °F) at the upper boundary with the crust; to over 4, 000 °C (7, 230 °F) at the boundary with the core. Although the higher temperatures far exceed the melting points of the mantle rocks at the surface (about 1200 °C for representative peridotite), the mantle is almost exclusively solid. The enormous lithostatic pressure exerted on the mantle prevents melting, because the temperature at which melting begins (the solidus) increases with pressure.

Mantle convection

Colors closer to red are hot areas and colors closer to blue are cold areas. In this figure, heat received at the core–mantle boundary results in thermal expansion of the material at the bottom of the model, reducing its density and causing it to send plumes of hot material upwards. Likewise, cooling of material at the surface results in its sinking.

Because of the temperature difference between the Earth’s surface and outer core and the ability of the crystalline rocks at high pressure and temperature to undergo slow, creeping, viscous-like deformation over millions of years, there is a convective material circulation in the mantle. Hot material upwells, while cooler (and heavier) material sinks downward. Downward motion of material occurs at convergent plate boundaries called subduction zones. Locations on the surface that lie over plumes are predicted to have high elevation (because of the buoyancy of the hotter, less-dense plume beneath) and to exhibit hot spot volcanism. The volcanism often attributed to deep mantle plumes is alternatively explained by passive extension of the crust, permitting magma to leak to the surface (the “ Plate” hypothesis). The convection of the Earth’s mantle is a chaotic process (in the sense of fluid dynamics), which is thought to be an integral part of the motion of plates. Plate motion should not be confused with continental drift which applies purely to the movement of the crustal components of the continents. The movements of the lithosphere and the underlying mantle are coupled since descending lithosphere is an essential component of convection in the mantle. The observed continental drift is a complicated relationship between the forces causing oceanic lithosphere to sink and the movements within Earth’s mantle.

Although there is a tendency to larger viscosity at greater depth, this relation is far from linear and shows layers with dramatically decreased viscosity, in particular in the upper mantle and at the boundary with the core. [39] The mantle within about 200 km (120 mi) above the core–mantle boundary appears to have distinctly different seismic properties than the mantle at slightly shallower depths; this unusual mantle region just above the core is called D″ (“ D double-prime”), a nomenclature introduced over 50 years ago by the geophysicist Keith Bullen. D″ may consist of material from subducted slabs that descended and came to rest at the core–mantle boundary and/or from a new mineral polymorph discovered in perovskite called post-perovskite.

Earthquakes at shallow depths are a result of stick-slip faulting; however, below about 50 km (31 mi) the hot, high pressure conditions ought to inhibit further seismicity. The mantle is considered to be viscous and incapable of brittle faulting. However, in subduction zones, earthquakes are observed down to 670 km (420 mi). A number of mechanisms have been proposed to explain this phenomenon, including dehydration, thermal runaway, and phase change. The geothermal gradient can be lowered where cool material from the surface sinks downward, increasing the strength of the surrounding mantle, and allowing earthquakes to occur down to a depth of 400 km (250 mi) and 670 km (420 mi). The pressure at the bottom of the mantle is ~136 GPa (1. 4 million atm). Pressure increases as depth increases, since the material beneath has to support the weight of all the material above it. The entire mantle, however, is thought to deform like a fluid on long timescales, with permanent plastic deformation accommodated by the movement of point, line, and/or planar defects through the solid crystals comprising the mantle. Estimates for the viscosity of the upper mantle range between 1019 and 1024 Pas, depending on depth, temperature, composition, state of stress, and numerous other factors. Thus, the upper mantle can only flow very slowly. However, when large forces are applied to the uppermost mantle it can become weaker, and this effect is thought to be important in allowing the formation of tectonic plate boundaries.

Exploration

Exploration of the mantle is generally conducted at the seabed rather than on land because of the relative thinness of the oceanic crust as compared to the significantly thicker continental crust. The first attempt at mantle exploration, known as Project Mohole, was abandoned in 1966 after repeated failures and cost over-runs. The deepest penetration was approximately 180 m (590 ft). In 2005 an oceanic borehole reached 1, 416 metres (4, 646 ft) below the sea floor from the ocean drilling vessel JOIDES Resolution.

On 5 March 2007, a team of scientists on board the RRS James Cook embarked on a voyage to an area of the Atlantic seafloor where the mantle lies exposed without any crust covering, midway between the Cape Verde Islands and the Caribbean Sea. The exposed site lies approximately three kilometers beneath the ocean surface and covers thousands of square kilometers. A relatively difficult attempt to retrieve samples from the Earth's mantle was scheduled for later in 2007. The Chikyu Hakken mission attempted to use the Japanese vessel Chikyū to drill up to 7, 000 m (23, 000 ft) below the seabed. This is nearly three times as deep as preceding oceanic drillings.

A novel method of exploring the uppermost few hundred kilometers of the Earth was recently proposed, consisting of a small, dense, heat-generating probe which melts its way down through the crust and mantle while its position and progress are tracked by acoustic signals generated in the rocks. The probe consists of an outer sphere of tungsten about one meter in diameter with a cobalt-60 interior acting as a radioactive heat source. It was calculated that such a probe will reach the oceanic Moho in less than 6 months and attain minimum depths of well over 100 km (62 mi) in a few decades beneath both oceanic and continental lithosphere. Exploration can also be aided through computer simulations of the evolution of the mantle. In 2009, a supercomputer application provided new insight into the distribution of mineral deposits, especially isotopes of iron, from when the mantle developed 4. 5 billion years ago.