

MANTLE AND CORE OF THE EARTH

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Summary

Earth can be divided into several zones. Zone B (at depths less than 400 km) is a region of pronounced lateral heterogeneity and seismic anisotropy. Lateral heterogeneity of all properties is clearly related to surface tectonics. The heterogeneity is of thermal and compositional origins. Anisotropy is caused by alignment of olivine crystals, and corresponds either to ancient deformations (in the lithosphere), or to present-day deformations (in the asthenosphere). Properties of the transition zone at depths between 400 km and 700 km are explained by a series of phase transformations. Phase transformations in the down-going slabs explain structure and seismicity of subduction zones.

The core–mantle boundary zone is the most complicated and heterogeneous in the lower mantle. Large-scale low-velocity regions are found in this zone beneath the central Pacific and Africa. High-velocity anomalies form a ring around the Pacific. Small-scale scatterers of seismic waves are found in many regions. In some regions there are indications of a thin, ultra-low velocity layer immediately above the core–mantle boundary. There are observations of seismic anisotropy in the core–mantle boundary zone, with SH waves traveling faster than SV. Both the upper and the lower mantle are in a state of convection, but the overall flow pattern still is a matter of dispute. The outer core is featureless, but a complicated structure is present in the inner core. This structure includes strong anelastic attenuation in the upper layer of the inner core, and laterally variable seismic anisotropy.

1. Introduction

Earth consists of a solid crust, a solid mantle, and a core, the outer part of which is liquid. The central part of the core (inner core) with a radius of 1221 km is solid, or a mixture of solid and liquid. The depth of the upper boundary of the mantle (the Moho discontinuity) varies between about 80 km (beneath Tibet) and about 10 km in the oceans. The lower boundary of the mantle is at a depth near 2890 km, and thus the mantle represents about 83 % of the volume of Earth. By comparison, the volume of the crust is only 1 %.

Processes in the mantle and the core are responsible for many phenomena observed in the crust and at Earth's surface. For example, Earth's magnetic field is an effect of convection in the outer core. Large-scale deformations in the crust are related to convection in the mantle. Bullen in 1942 divided the mantle into several shells: B (from the Moho to 413 km), C (from 413 km to 984 km), D' (from 984 km to 2700 km), and D" (from 2700 km to the core–mantle boundary). In spite of the discoveries made after 1942, this division still is useful. Zone B is remarkable for its lateral heterogeneity, which is clearly related to surface tectonics. The layer between 413 km and 660 km in zone C is known as the transition zone, within which velocity and density gradients are much higher than in the rest of the mantle. Zone B and the transition zone taken together are termed the upper mantle. The rest is termed the lower mantle. The region D" of the lower mantle has an extremely complicated and laterally heterogeneous structure, which can be compared in complexity with the crust and the uppermost mantle.

This article gives a review of the classic and more recent data pertaining to the properties of the mantle and the core. The most detailed knowledge of the structure of the deep interior of the earth is provided by seismological observations, and the section that follows contains a short overview of seismic methods. Radial variations of properties of Earth are discussed in Section 3. In the other sections, attention is shifted to lateral variations and fine details.

2. Seismic Methods

2.1 Seismic Waves

Earthquakes generate elastic (compression P and shear S) body waves, which travel through the Earth along the rays. P and S waves are polarized along and normal to the ray, respectively. P waves travel about twice as fast as S waves. Under the assumption of isotropy (that is, assuming that velocity is independent of the propagation direction) the velocities V_P and V_S are related to the density ρ , adiabatic incompressibility or bulk modulus K , and rigidity or shear modulus μ as:

$$V_P = ((K + 4\mu/3)/\rho)^{1/2} \quad (1)$$

$$V_S = (\mu/\rho)^{1/2} \quad (2)$$

The seismic parameter Φ is defined as:

$$\Phi = V_P^2 - (3/4)V_S^2 \quad (3)$$

At high pressures and temperatures, thermodynamic estimates for K are more reliable than for μ , and the seismic parameter is often used instead of V_P and V_S , when seismic data are compared with the data of high-pressure physics.

The Poisson ratio, σ , is defined as:

$$\sigma = (1/2)(K - (2/3)\mu)/(K + (2/3)\mu) \quad (4)$$

The Poisson ratio is very sensitive to composition and temperature. The parameter η is defined as:

$$\eta = (\Phi/\rho g)(d\rho/dz) \quad (5)$$

where z is depth, and g is acceleration due to gravity, which is almost constant through the mantle (around $10 \text{ m}^2\text{s}^{-1}$). The deviation of η from unity indicates that the variations of velocities and density with depth deviate from those that are expected for adiabatic self-compression of a homogeneous material.

When the body wave meets a discontinuity in elastic parameters it generates reflected and transmitted waves. The reflected or transmitted wave can be in the same mode (P or S) or mode-converted. As a result of interaction of primary waves with discontinuities, the number of seismic arrivals proliferates, as does the number of observables for constraining the parameters of Earth's interior. The most important data related to the structure of Earth's interior are travel times of body waves as functions of distance from the source. The travel times can be inverted for V_P and V_S as functions of depth. Further constraints on the properties of Earth's interior can be obtained by considering the amplitudes and waveforms of seismic phases.

Unlike the body waves, surface waves propagate parallel to Earth's surface. Travel times of surface waves depend on their frequency of vibration. This dependence can be used to infer V_S as a function of depth. There are two kinds of surface waves, which

differ by velocities and polarization. Love waves are polarized horizontally in the direction normal to the propagation direction. Motion in Rayleigh waves is elliptic in the vertical plane containing the source and the receiver. At very long periods (hundreds and thousands of seconds) the Earth resonates as a whole at certain discrete periods. These periods can be inverted for the velocities and density inside the Earth.

Early seismic observations allowed us to construct travel-time tables of the main body-wave phases, and to derive the corresponding one-dimensional (1-D) velocity models of the mantle and the core. Here, the term “1-D model” means a model that depends only on depth. Subsequent observations have shown that the actual velocities in specific regions can deviate from velocities in “average” 1-D models by up to several percent. In particular, large-scale lateral heterogeneity of the upper mantle was documented by using the travel times of long-period (50–200 s) surface waves. The free vibrations of the Earth, which were reliably recorded for the first time in 1960, have also shown significant departures of the observed periods from those predicted by the 1-D model. Besides lateral heterogeneity, two other properties of the Earth’s interior emerge from seismic data: anisotropy and anelasticity.

2.2 Anisotropy and Anelasticity

Seismic anisotropy is the dependence of seismic velocities on the direction of propagation (for both P and S waves) and on polarization (for S and surface waves). The simplest (hexagonal or axially symmetric) anisotropy can be described by five elastic constants (instead of two in an isotropic solid), and two angles, which define the direction of the axis of symmetry. In the general case the number of elastic constants is 21. There are two principal causes of anisotropy: lattice preferred orientation (LPO) and shape preferred orientation (SPO). Elastic anisotropy is a property of crystals of all rock-forming minerals, and LPO is created by crystal alignment. SPO is created by aligned inclusions or fine layering. Wave fields in anisotropic solids are far more complicated than in isotropic ones. In particular, in anisotropic solid there are two quasi-shear waves, which are polarized at 90° to each other and propagate with different velocities.

Earth is not perfectly elastic. Internal friction in rocks is the reason for damping of traveling seismic waves with distance (and of free vibrations of Earth with time). Traveling seismic waves are also damped by geometrical spreading and scattering. Internal friction transforms energy of strain into heat: friction includes such processes as movement of dislocations in crystals and sliding crystals along grain boundaries. Anelasticity is measured by the loss of strain energy E per cycle of oscillation:

$$\delta E/E = 2\pi/Q \quad (6)$$

where Q is a quality factor. A lower value of Q corresponds to stronger damping. Anelastic attenuation in the mantle is much stronger for S waves than for P waves. At frequencies lower than 1 Hz, Q in the mantle is almost independent of frequency. Processes of internal friction are activated by temperature, and lower values of Q correspond to higher temperature. Partial melting or presence of volatiles may result in very low Q . Anelastic attenuation makes waves slower at longer periods.

2.3 Seismic Tomography

At present, mapping 3-D velocity heterogeneity of Earth's interior is one of the principal objectives of seismology. 3-D deviations of velocity from standard values in the mantle are small (up to several percent), but they may correspond to much stronger variations of other properties, like temperature or viscosity. An anomaly of seismic velocity of a few percent corresponds to an anomaly of temperature of a few hundred degrees Celsius. A temperature change of 100 °C changes the viscosity of the mantle by an order of magnitude. These variations have enormous effect upon Earth's dynamics.

The various techniques used to construct 3-D velocity models are termed seismic tomography. The general idea is to choose a starting 1-D velocity model, and to find small 3-D corrections that minimize discrepancies between the observed travel times for many wavepaths across the same region and the predicted travel times. Since the corrections are small, they are related by linear equations to the corresponding perturbations of travel times.

For practical tomography, the earth's medium should be discretized. In one approach, Earth is divided into a large number of cells, and the solution of the corresponding system of linear equations provides a correction for every cell. In the regions that are well sampled by the waves, the cells can be made very small, and can provide higher resolution than elsewhere. This approach is especially advantageous in high-resolution regional studies. In another parameterization, the spatial variations of velocity within the Earth are expanded in a series of orthogonal basis functions, and the correction to the starting model is expressed as a smooth function of coordinates. The horizontal variations are approximated by spherical harmonics. Resolution of the resulting model depends on the degree at which the series is truncated. Selection of the cut-off number depends on the coverage provided by the data. This approach is optimum for mapping large-scale heterogeneities.

In other modifications of seismic tomography the corrections are obtained by fitting not only the travel times, but also the waveforms. Similar principles can be used for retrieving variations of the quality factor Q .

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Biographical Sketch

L. P. Vinnik, was born in 1935 at Smolensk (USSR), and is married with one son. He graduated from school in 1952, and entered the geological faculty of Moscow University. He graduated from the university in 1957. In 1957–1959 he wintered in the Antarctic as a scientist of the Soviet Antarctic expedition. From 1959, he was a scientist of the Institute of Physics of the Earth (IPE) of the Soviet Academy of Sciences, Moscow. He was a candidate of physics and mathematics (PhD, 1966), and doctor of physics and mathematics. All degrees have been obtained from the IPE. At present, he is professor, director of laboratory, and head of department of the interior structure of the IPE, a Fellow of the American Geophysical Union, and member of the Academia Europaea. He was awarded the Alexander-von-Humboldt prize (1991, Germany) and the B. B. Golitsyn prize of the Russian Academy of Sciences (1997).