PHYSICAL PROPERTIES OF ROCKS

Because most of the earth is not accessible for direct observation, it is important to understand as much as possible about the physical properties of rocks. In many geological settings or studies (e.g., those concerning oil exploration, crustal structure, mantle convection), the only information we have comes from indirect observations of gravity and magnetic anomalies and/or seismic velocities. In order to make an assessment of the composition or structure of the subsurface for inaccessible regions, we need to know as much as possible about the physical properties of the rocks that compose the subsurface. The rock properties most often studied include the density, elastic constants (bulk modulus, shear modulus, Poisson's ratio), thermal conductivity, coefficient of thermal expansion, magnetic susceptibility, electrical conductivity, electrical resistivity, and viscosity. A detailed discussion of all these properties is beyond the scope of this entry, so we will focus on the elastic constants, which can be directly related to the seismic velocity, density, coefficient of thermal expansion, and viscosity.

Elastic Constant and Seismic Velocities

An isotropic solid is one in which the response of the solid to an imposed stress is independent of the orientation of the solid. Only cubic minerals are truly isotropic. Because of the nature of the crystal structure of minerals, almost all minerals are slightly anisotropic, with as much as a 10 percent difference in their elastic constant depending on the orientation. However, an aggregate of crystals in various random orientations is often assumed to be isotropic. We begin our discussion of elastic constants with the isotropic case.

For an isotropic, homogeneous solid there is a linear relation between stress and strain, under the limit of infinitesimal strains. The limit of infinitesimal strains is a reasonable approximation for the deformation caused by seismic waves as they travel through the earth. For an isotropic system, the number of independent elastic constants reduces to two. We can express the relationship between stress (σ) and strain (ε) as

$$\sigma_{ij} = \lambda \delta_{ij} \sum_{k} \varepsilon_{kk} + 2\mu\varepsilon_{ij} \tag{1}$$

where δ_{ij} is equal to 1 if $i \neq j$ and to zero if $i \neq j$, and λ and μ are the two independent Lamé constants. The elastic properties of an isotropic material can be described by elastic moduli: the shear modulus, μ, and the bulk modulus or incompressibility, K, defined as

$$K = \frac{3\lambda + 2\mu}{3} \tag{2}$$

We can measure the bulk modulus with the following experiment. Given a uniform cube of an isotropic material with a volume of V, at a starting pressure P, we increase the pressure to a value of $P + \Delta P$ and measure the change in volume ΔV. The bulk modulus is given by

$$K = -V \cdot \frac{\Delta P}{\Delta V} = \frac{dP}{d \ln V} \tag{3}$$

Another useful parameter is Poisson's ratio, ν. We can measure Poisson's ratio with the following simple experiment. We take a long, cylindrical rod and subject it to a uniform stress along its axis of symmetry. Poisson's ratio is defined as minus the ratio of the strain normal to the stress axis to the strain along the stress axis (i.e., ratio of thinning to elongation or thickening to contraction):

$$\nu = -\frac{\varepsilon_{22}}{\varepsilon_{11}} \tag{4}$$

Poisson's ratio is useful in earth science because it can be expressed as a function of the ratio v_p/v_s of the velocities of pressure (v_p) and shear (v_s) seismic waves. We can write the seismic velocities in terms of the elastic moduli or the Lamé parameters:

$$v_p = \left(\frac{\lambda + 2\mu}{\rho}\right)^{1/2} = \left(\frac{K + 4\mu/3}{\rho}\right)^{1/2} \tag{5}$$

$$v_s = \left(\frac{\mu}{\rho}\right)^{1/2} \tag{6}$$

Hence,

$$\nu = \frac{(v_p/v_s)^2 - 2}{2[(v_p/v_s)^2 - 1]} \tag{7}$$

In the crust, we often find $v_p = \sqrt{3}\, v_s$, which corresponds to $\nu = 0.25$. In this case, $\lambda = \mu$, so there is only one independent elastic modulus. Such a material is called a Cauchy solid.

Since the mid–1960s, the seismic velocity of a larger number of rocks and minerals has been measured. These have provided a basic understanding of many factors that influence seismic velocity and attenuation of rocks believed to be abundant constituents of the continental lithosphere.

The effect of confining pressure on velocities has been reported in a number of investigations. An example of data for a typical crystalline rock (Twin Sisters Dunite, an olivine-rich igneous rock) is shown in Figure 1. The characteristic shape of the curve of velocity versus pressure is attributed to the closure of microcracks. As can be seen from Figure 1, much of the closure takes place over the first 100 megapascals (MPa). Velocities measured in crystalline rocks at pressures up to 3,000 MPa demonstrate that changes in velocity with pressure do not approach those of single crystals until the confining pressure is above 1,000 MPa. Even at these high pressures, solid contact between the mineral components is probably only approximate because some porosity has originated from anistropic thermal contraction of the minerals.

Fewer data are available on the influence of temperature of rock velocities. It has been well known that the application of temperature to a rock at atmospheric pressure results in the creation of cracks that often permanently damage the rock. Thus, reliable measurements of the temperature derivatives of velocity must be obtained at confining pressures high enough to prevent crack formation. In general, pressures of 200 MPa are sufficient for temperature measurements to 300°C.

A wide variety of techniques have been employed to measure the influence of temperature on rock velocities. An example of data showing the influence of temperature on velocities is shown in

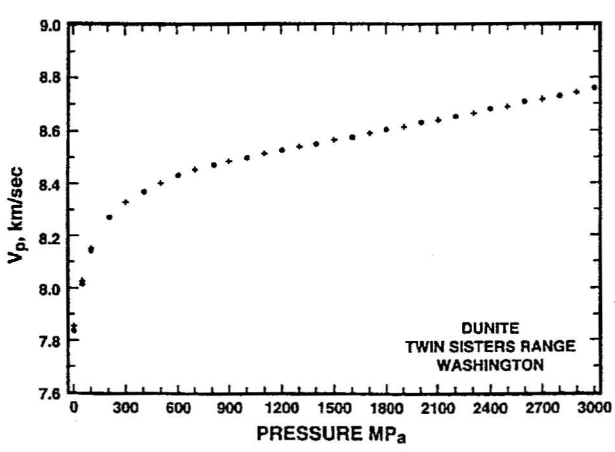

Figure 1. Laboratory measurement of seismic P velocity as a function of pressure for Twin Sisters Dunite (olivine-rich igneous rock).

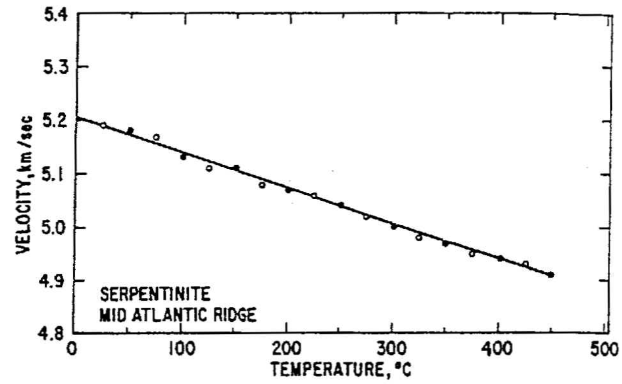

Figure 2. Laboratory measurement of seismic velocity as a function of temperature for serpentinite (metamorphic basalt or gabbro) from the Mid-Atlantic Ridge.

Figure 2. Increasing temperature decreases velocities, whereas increasing pressure increases velocities. Thus, in a homogeneous crustal region, velocity gradients depend primarily on the geothermal gradient. The change of velocity with depth is given by

$$\frac{dV}{dz} = \left(\frac{\partial V}{\partial P}\right)_T \frac{dP}{dz} + \left(\frac{\partial V}{\partial T}\right)_P \frac{dT}{dz} \tag{8}$$

where z is depth, T is temperature, and P is pressure. For regions with normal geothermal gradients (25–40°C/km), the change in compressional velocity with depth dV_p/dz is close to zero. However, in the high heat-flow regions, crustal velocity reversals are expected if compositional changes with depth are minimal.

Amplitudes of seismic waves decrease with increasing distance from their source. This property is called seismic wave attenuation. Seismic wave attenuation has great potential as a tool to yield a better understanding of the anelastic properties, and hence the physical state, of rocks in the earth's interior.

The three parameters most often reported as the attenuation are the seismic quality factor Q, also referred to as the specific attenuation Q^{-1}, the attenuation coefficient, α, and the logarithmic decrement, δ. These are related for low-loss materials ($Q > 10$) by

$$\frac{1}{Q} = \frac{\alpha V}{\pi f} = \frac{\delta}{\pi} \tag{9}$$

where V is the phase velocity and f is the frequency. In both the field and laboratory, difficulties arise in separating the intrinsic dissipation of the rock, that is, processes by which seismic energy is converted into heat, from geometric spreading, transmission losses, scattering, and other factors. Nevertheless, the utilization of laboratory attenuation measurements to tie seismic data to the anelastic properties of rocks is promising, and the refinement of laboratory techniques and the theory concerning the mechanisms involved has yielded and will continue to supply valuable insights into the structure and composition of the continental crust and upper mantle.

All investigations have found that Q increases with increasing confining pressure. Laboratory measurements show a sharp increase in Q at low pressures, which then levels off at high pressures, a response similar to that observed for velocities. The form of the Q versus P curve is generally attributed, therefore, to the closure of microcracks. As with velocity measurements, few researchers have studied attenuation as a function of temperature for rocks of the lithosphere. At temperatures below the boiling point of the rock's volatiles, Q appears to be temperature-independent, and above this Q increases, indicating outgassing of pore fluids and/or thermal cracking. At the onset of partial melting, Q decreases.

Seismic velocities are the most often used properties of rocks. Seismologists have produced one-dimensional (radial) models of seismic velocity from the surface to the center of the earth (Figure 3). While seismic velocities generally increase gradually as a function of pressure (depth), an abrupt jump in seismic velocity over a small pressure (depth) range usually indicates a change in chemical composition or a change in the solid phase of the material. For example, the jump in v_p and v_s at approximately 400-km depth corresponds with the pressure and temperatures of the olivine [(Mg, Fe)$_2$SiO$_4$] to wadslayite phase change observed in the laboratory. Seismic velocity models place an important constraint on the composition of the interior of the earth. Because the seismic velocities of minerals such as olivine, pyroxene, garnet, and perovskite behave differently as a function of depth, one can attempt to match the seismic velocity curve with simple models of mantle composition. At 2,900-km depth, there is an abrupt change in both density and seismic velocity. This marks the chemical boundary between the silicate mantle and the liquid iron outer core.

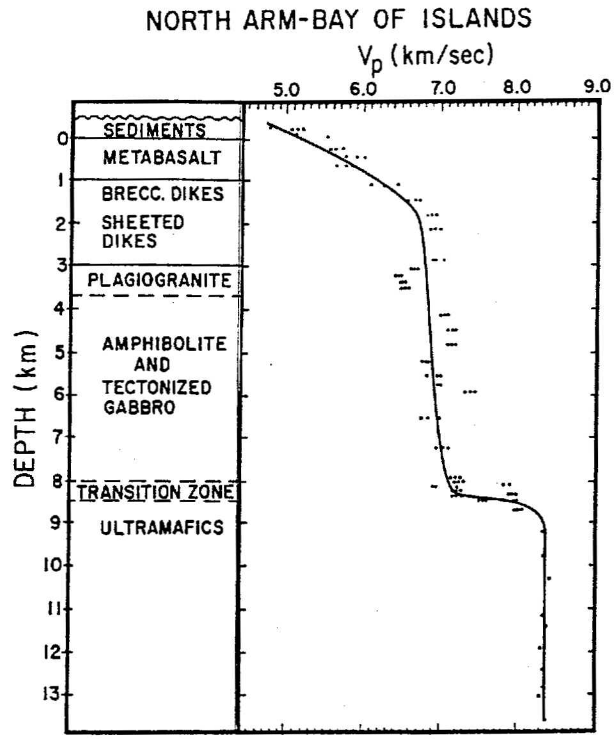

Figure 3. Seismic velocities as a function of depth from the Preliminary Reference Earth Model (PREM).

In the 1990s, seismic velocities have played an important role in understanding the structure and composition of the earth's crust. An excellent example is offered by combining oceanic crustal seismic studies with laboratory measurements of the velocity structure of ophiolites. Ophiolites, on land exposures of oceanic crust and upper mantle, contain a stratified sequence of rocks which, from top to bottom, consists of marine sediments, pillow basalts, dikes, gabbros, and peridotites. The laboratory-determined velocity structure of ophiolites (see Plate 32) matches very well with field measurements of velocities in oceanic basins. The initial rapid increase of velocity extending to depths of 3 km originates from a decrease in porosity with depth. At depths between 4 and 7 km, velocities are fairly constant. This region contains mainly gabbro and metagabbroic rocks. The rapid increase in velocity encountered at 7 km is similar to field observations of velocity changes at the Mohorovičič discontinuity. The ultramafic sections of ophiolites show 6–8 percent anisotropy originating from preferred orientation of olivine and pyroxene. This anisotropy correlates well with upper mantle seismic anisotropy measured with marine seismic surveys.

Velocities in single crystals of the common rock-forming minerals vary significantly with propagation direction. In general, for a given propagation direction in anisotropic media such as single crystals, there are three waves, one compressional and two shear. Their vibration directions form an orthogonal set, which usually are not parallel or perpendicular to the propagation direction. The propagation of waves is related to the single-crystal elastic constants through the Christoffel equation, which gives the three velocities for each direction as roots of a cubic equation. Details of wave propagation are related to the crystal symmetry. Most metamorphic rocks and some cumulate igneous rocks have preferred mineral orientations that are usually related to cleavage, foliation, or banding. It follows that many rocks are seismically anisotropic in a manner similar to single crystals. Compressional wave velocities vary with propagation direction and two shear waves travel in a given direction through the rock with different velocities. This latter property of anisotropic rocks, termed shear wave splitting, has recently been observed in several crustal and upper mantle regions.

Density and the Coefficient of Thermal Expansion

The density of rock is important for understanding gravity anomalies. Density depends on the chemical composition of the rock, the mineral structure, and the void space between minerals. Gravity anomalies are the result of mass anomalies at the surface of the earth (for example, changes in elevation) as well as mass anomalies within the earth (for example, mineral deposits or sedimentary basins). The relationship between a mass anomaly, dm, and the resulting gravity anomaly, dg, is

$$dg = G\,dm/r^2 = G d\rho\, V/r^2 \qquad (10)$$

where V is the volume of the mass anomaly and $d\rho$ is its density difference from the surrounding rocks, G is the universal gravitational constant (6.67×10^{-11} $m^3/kg\ s^2$), and r is the distance between the mass anomaly and the point where gravity is being measured. If we knew the distribution of mass anomalies within the earth exactly, we could integrate equation (10) over the whole earth and calculate the observed gravity everywhere. In routine gravity surveys, it is not uncommon to measure anomalies as small as 0.00001 percent of the average earth gravity to detect subsurface structure. From equation (10) it is clear that there is a trade-off between the size of the body (V) and the density difference ($d\rho$). Thus, the better we know the density of the subsurface material, the more accurately we can estimate the size of the body of interest.

The influence of temperature on density is also important in understanding crustal and mantle dynamics. For all earth materials, the density of a rock or mineral increases with increasing pressure and the density decreases with increasing temperature. The relationship between density and temperature is measured by the coefficient of thermal expansion, α, which is defined as

$$\rho(T) = \rho_0(1 - \alpha T) \qquad (11)$$

The coefficient of thermal expansion for the mineral olivine is approximately $3 \times 10^{-5}/°C$. Thus, a temperature change of 1,000° changes the density of the material by 3 percent. This density difference provides the force that drives mantle convection and, ultimately, plate tectonics. Density is also affected by pressure (mainly at mantle depths), changes in phase, and changes in composition. Thus, convection in the mantle is more complicated than a simple, uniform fluid as in most tank experiments of convection in the laboratory.

Viscosity

While earthquakes provide evidence that the earth behaves elastically on short timescales, on the timescales of postglacial rebound or mantle convection, the earth behaves like a viscous fluid. The viscosity of the mantle is one of the most important properties for understanding mantle flow, but it is also one of the most poorly constrained properties.

Laboratory measurements of deformation indicate that the viscosity of upper mantle minerals such as olivine is a strong function of temperature, pressure, grain size, and stress (Karato and Wu, 1993). For a temperature increase of 100 K, the viscosity decreases by an order of magnitude at constant stress. An increase of deviatoric stress by a factor of two decreases the viscosity by an order of magnitude. Other factors, such as partial pressure of oxygen and water content, may also have important effects, but are less well studied.

Because of the large difference between time and space scales in the laboratory and the mantle, estimates of mantle viscosity based on modeling large-scale geophysical observations (e.g., postglacial rebound) play an important role in our understanding of mantle viscosity structure; however, viscosity models deduced from these observations are not unique and require additional assumptions. For example, in modeling postglacial rebound, the thickness and extent of the ice sheet through time cannot be determined directly. The uncertainty in the ice sheet model adds to the complexity of the problem. In addition, the theoretical models are often greatly simplified to keep them mathematically tractable.

From these studies, several classes of viscosity models appear, one with essentially a uniform viscosity throughout the mantle and one with a low viscosity channel beneath the lithosphere and a higher viscosity in the lower mantle. At present, it seems that the observational constraints are not strong enough to exclude one of these models. Perhaps more worrysome, however, is that even less is known about the effect of lateral viscosity variations on surface observables.

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