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SEISMIC PROPERTIES OF ROCKS

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Synonyms

Rock P and S velocities

Definitions

Compressional (P) waves. Seismic waves in which the rock particles vibrate parallel to the direction of wave propagation.
Shear (S) waves. Seismic waves in which the rock particles vibrate perpendicular to the direction of wave propagation.
Poisson's ratio (σ). The ratio of the lateral unit strain to the longitudinal unit strain in a body that has been stressed longitudinally within its elastic limit ($2\sigma = (R^2 - 2)/(R^2 - 1)$ where $R = Vp/Vs$).
Transversely isotropic. Anisotropic solids with a single symmetry axis. Rock symmetry axes are usually normal to foliation, cleavage, or layering.

Introduction

Although many disciplines have contributed significantly to our knowledge of the Earth's interior, none has a resolution comparable to seismics. For nearly 6 decades seismic studies have provided geophysicists with worldwide information on crustal and upper mantle compressional (P) and shear (S) wave velocities. Significant data have recently become available on velocity gradients, velocity reversals, compressional and shear wave velocity ratios, and anisotropy in the form of azimuthal variations of compressional wave velocities, as well as shear wave splitting. Reflections within the crust and mantle originate from contrasts of acoustic impedances, defined as products of velocity and density. The interpretation of this seismic data requires detailed knowledge of rock velocities provided by laboratory techniques to a precision at least comparable with that of seismic measurements. In particular, to infer composition of the Earth's upper 30–50 km, the "crust," requires studies of the elasticity of rocks at conditions approaching those that exist at these depths. Of fundamental importance is the presence of mean compressive stress and temperature increasing with depth and on the average reaching about 1 GPa and 500 °C at the base of the crust. Because of this, the most relevant velocity measurements for identifying probable rock types within the crust have been measurements at elevated pressures and temperatures. These measurements often allow the seismologist to infer mineralogy, porosity, the nature of fluids occupying pore spaces, temperature at depth, crack orientations, etc.

Francis Birch (Figure 1) was the pioneer in the study of rock velocities. In addition to his laboratory work on physical properties of rocks and minerals at high pressures and temperatures, he was well known for his studies of heat flow and theoretical work on the composition of the Earth's interior. Two of his benchmark papers on compressional wave velocities in rocks (Birch, 1960, 1961) set the stage for modern experimental studies of rock elasticity and have been frequently cited during the past 5 decades. These papers for the first time provided information on compressional wave velocities for many common rock types, as well as major findings on their anisotropies and relations to density. It is interesting to note that these measurements were carried out to pressures of 1 GPa, a pressure at which even today only a limited number of laboratories have been able to generate for modern rock seismic velocity measurements.

Measurement techniques

Rock velocities are usually measured in the laboratory using the pulse transmission technique. The transit time

Harsh K. Gupta (ed.), *Encyclopedia of Solid Earth Geophysics*, DOI 10.1007/978-90-481-8702-7,

of either a compressional or shear wave is measured along the axis of a cylindrical rock specimen of known length. The cores are usually taken from rock samples using a 2.54 cm inner diameter diamond coring bit. The cores are trimmed and ground flat and parallel on a diamond grinding disk. The volume of each core is obtained from the length and diameter. The cores are weighed and densities are calculated from their masses and dimensions. The cores are then fitted with a copper jacket to prevent penetration of high-pressure oil into the rock samples. For measurements at high temperatures, where gas is the pressure medium, the samples are usually encased in stainless steel. Transducers are placed on the ends of the rock core (Figure 2). Compressional and shear waves are often generated by means of lead zirconate titanate (PZT) and AC cut quartz transducers with resonant frequencies of 1 MHz. The sending transducer converts the input, an electrical pulse of 50–500 V and 0.1–10 μs width, to a mechanical signal, which is transmitted through the rock. The receiving transducer changes the wave to an electrical pulse, which is amplified and displayed on an oscilloscope screen (Figure 3). Once the system is calibrated for time delays, the travel time through the specimen is determined directly by a computer or with the use of a mercury delay line. The major advantage of the delay line is that it increases the precision, especially for signals with slow rise times, because the gradual onset of the first arrival from the sample is approximated by the delay line. The velocity is the ratio of the length of the specimen to the travel time of the compressional or shear wave. The total error limits for Vp and Vs are estimated to be less than 1%. Interfacing the pressure system with a computer for data acquisition and storage permits automatic calculations of velocities as successive readings are taken. Using a least-squares routine, the computer fits a curve to the data points and calculates velocities for selected pressures. A velocity versus pressure curve is plotted along with recorded data points. Sample length, density, measured pressure velocity pairs, traces of the waveforms at selected pressures, the curve fit equations, and calculated pressure velocity pairs are recorded and stored digitally.

Hydrostatic pressure generating systems capable of producing true hydrostatic pressures as high as 3 GPa, equivalent to a depth of approximately 100 km, have been used for rock velocity measurements. Low viscosity synthetic petroleum and argon for high temperature measurements are frequently used as pressure media. An alternate technique for obtaining velocities under quasi-hydrostatic conditions has used a triaxial press with cubic samples. Transducers are placed on the six pistons and corrections are made for travel times through the pistons. Rock velocities obtained using this technique have provided valuable information on the effect of temperature on velocity, but have been limited to pressures of 0.6 GPa.

The behavior of a rocks velocity as a function of pressure is primarily dependent upon mineralogy and porosity. Many igneous and metamorphic rocks have porosities of the order of a few tenths of 1%, which are present as thin openings between grain boundaries. As pressure is applied to the rock, the cracks close and velocities increase. Once the cracks close any increase in velocity with increasing pressure is related to the intrinsic effects of pressure on the mineral velocities. This is illustrated in Figure 4 for a garnet granulite. Velocities first increase rapidly over the first 100 MPa as cracks close and then increase slowly as pressure is increased. Also the velocity determined at a given pressure depends upon whether the pressure is approached from lower or higher pressure (Figure 4). This hysteresis is usually quite small if sufficient time is taken for measurements between pressure increments.

A considerable number of investigations have also focused on the influence of temperature on rock velocities. These studies have used either resonance techniques or more frequently the pulse transmission method. Early studies demonstrated that the application of temperature to rock at atmospheric pressure results in the creation of cracks that often permanently damage the rock and dramatically lower velocities. Thus reliable measurements of the temperature derivatives of velocities are obtained only at confining pressures high enough to prevent crack formation. At elevated confining pressures $\delta Vp/\delta T$ for common rocks often ranges from -0.3×10^{-3} to -0.6×10^{-3} km/s/°C and $\delta Vs/\delta T$ varies between -0.2×10^{-3} and -0.4×10^{-3} km/s/°C.

Rock velocities

Seismic velocities have been measured for practically all igneous and metamorphic rock types believed to be important constituents of the lithosphere. Because rock classification schemes allow for considerable variations in mineralogy for a given rock type, many rocks have wide ranges in elastic properties. However, some lithologies, such as the monomineralic rocks hornblendite and dunite with little or no alteration have fairly well-defined velocities. For detailed summaries of rock seismic properties the reader is referred to the compilations of Birch (1960), Christensen (1982, 1996), Gerbande (1982), Holbrook et al. (1992), Rudnick and Fountain (1995), and Mavko et al. (1998).

Table 1 contains average velocities, in the order of increasing compressional wave velocity, for several common igneous and metamorphic rocks. Volcanic rocks usually have lower velocities than their plutonic equivalents. This is due to the presence of glass, abundant alteration products and vesicles in volcanic rocks, all of which have lower velocities. In general, for a given composition, velocity increases with increasing metamorphic grade. For example, mica and quartz bearing schists have higher velocities than slates and phyllites. Low-grade metamorphosed basalts have lower velocities than higher grade amphibolite and mafic granulite. Eclogites have the highest velocity of mafic rocks. Note that shear velocities are relatively high in quartzites and low in serpentinites.

Early attempts to infer crustal composition by comparing laboratory and field derived velocities relied primarily on compressional wave velocities. However correlations between compressional wave velocity and composition are limited due to the similar velocities of many common crustal rock types. Because of this nonuniqueness of compressional wave velocity laboratory and field data comparisons, many recent studies have focused on investigations of crustal composition using both compressional and shear wave velocities. In these studies the ratio Vp/Vs or Poisson's ratio (σ) calculated from Vp/Vs have resolved some of the ambiguities.

The values of Vp/Vs and σ, assuming isotropic elasticity, are given in Table 1 for several common igneous and metamorphic rocks at 1 GPa. This high-pressure eliminates cracks so the values only reflect mineralogy. The relatively low σ for quartzites (0.10) agrees well with isotropic aggregate calculations based on the elastic constants of single crystal quartz. Anorthosites, on the other hand, have relatively high Poisson's ratios ($\sim$0.31). As expected, values for the granites and granitic gneisses, consisting primarily of quartz and feldspar, fall in between those of quartzites and anorthosites and are relatively low. Thus, crustal regions where field measured values of $\sigma \leq 0.25$ are observed, are likely quartz-rich. Serpentinites containing lizardite, the variety of serpentine stable at crustal PT conditions, have extremely high values of Poisson's ratio (0.36), whereas unaltered dunites and peridotites have Poisson's ratios in the range of 0.25–0.26. Partially serpentinized dunites and peridotites have Poisson's ratios that fall between these limiting values. Laboratory measurements have established a well-defined relationship between Poisson's ratio, percent serpentinization and density. Changes in Poisson's ratio with progressive metamorphism of mafic igneous and pelitic rocks are considerably more complicated than the above examples (Christensen, 1996).

Velocity anisotropy

Most crustal and upper mantle rocks show some degree of velocity anisotropy, which can originate from several processes. Laminar flow within magma and lava will orient elongate crystals such as feldspars along flow directions. Tabular sediments may settle preferentially and anisotropy may be enhanced by sediment compaction. Plastic flow and recrystallization during metamorphism often produce strong mineral orientations parallel to foliation and banding. In shallow crustal rocks oriented cracks producing anisotropy often originate from differential principle stresses. Anisotropy observed in laboratory measurements at pressures above approximately 100 MPa, where cracks are closed, often originate from preferred orientations of highly anisotropic minerals such as micas, amphiboles, pyroxenes, and olivine.

In general, for a given propagation direction in an anisotropic rock there are three waves, one compressional and two shear. Their vibration directions form an orthogonal set, which usually are not parallel or perpendicular to their propagation direction. Compressional and shear 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, was first recognized in laboratory studies and has been observed by field studies in several crustal and upper mantle regions.

Anisotropy is usually studied in the laboratory by taking cores from a rock in different directions. Following the early investigations of Birch (1960, 1961) it is common practice to take three mutually perpendicular cores for velocity measurements. If the rock has a planar structure, such as a foliation, cleavage, or bedding, two cores are oriented with their axes within the planar structure. One of these cores is oriented parallel to a lineation, if present. In general, three velocities are measured per core: the compressional wave velocity, the velocity of the shear wave vibrating parallel to the layering, and the velocity of the shear wave vibrating in a plane perpendicular to layering. For cores taken perpendicular to layering, shear wave velocities are measured with vibration directions parallel to the axes of the cores taken in the layering.

It has been demonstrated that the above procedure provides information on maximum compressional wave anisotropy and maximum shear wave splitting. In general, the highest compressional wave velocities propagate in the plane of the foliation and parallel to lineations. Maximum shear wave splitting is observed for propagation within the planar structures. At near-normal incidence there is often ~~a lack of~~ minimal shear wave splitting. Au3

Velocity measurements using multiple cores have provided detailed information on anisotropic velocities for non-axial propagation (e.g., Johnston and Christensen, 1995; Christensen and Okaya, 2007).The number of measurements necessary to completely describe wave propagation depends on the symmetry of the rock fabric. In the most general anisotropic elastic solid, 21 independent constants are required to describe the equations of motion. Examples of materials with this type of behavior are minerals possessing triclinic symmetry such as kyanite and some feldspars. For many rocks, the existence of symmetry elements in the elastic properties leads to the vanishing of some elastic constants along with simple algebraic relations between others (e.g., Auld, 1990). Some crustal and upper mantle metamorphic rocks behave as elastic solids with orthorhombic symmetry, which require nine independent constants to describe the elastic tensors. Shales and many low to medium grade metamorphic rocks often have well-developed bedding or foliation and behave as transversely isotropic elastic solids (hexagonal symmetry with the symmetry axis normal to bedding or foliation). Transversely isotropic solids have five independent elastic constants, which can be calculated from five independent velocity measurements (two compressional wave velocities, one quasi-compressional wave velocity, and two shear wave velocities) and density. In an isotropic solid, only two independent constants are required for a complete description of elastic behavior and wave velocities are independent of propagation direction.

To describe three-dimensional wave propagation in anisotropic rocks, phase velocity surfaces can be calculated using the Kelvin-Christoffel equations (e.g., Auld, 1990) and elastic constants can be determined from velocity and density measurements. For transversely isotropic rocks, these surfaces describe variations in velocity as a function of angle to the bedding or foliation (Figure 5). Three velocity surfaces are calculated, one for the quasi-compressional wave, one for the shear wave vibrating parallel to the planar structure, and one for the quasi-shear wave vibrating in a plane perpendicular to the foliation or bedding. For propagation parallel and perpendicular to the foliation all wave modes are pure. The velocities shown in Figure 5 were calculated from velocity measurements at 600 MPa for a transversely isotropic quartz -mica schist from South Island, New Zealand. They show several important features about elastic wave propagation in this rock, which are typical of many foliated rocks. First, compressional wave velocities do not increase significantly until propagation directions greater than about 45° from foliation normal are reached. At larger angles, compressional wave velocity increases rapidly and reaches a maximum for propagation parallel to the foliation. Shear wave singularities (directions in which two shear waves have equal velocities) occur for propagation parallel to and at approximately 50° to the symmetry axis. Shear wave splitting occurs for all other propagation directions and reaches to maximum of 90° from the normal to the foliation.

Summary

Beginning with the compressional wave velocity measurements of Birch (1960, 1961), much has been learned about elastic wave propagation in rocks. Compressional wave velocities are now available for most common rock types at pressures existing in the continental crust and uppermost mantle. Additional important contributions include laboratory measurements of shear wave velocities, velocity ratios, and the influence of temperature on velocities. Estimates of crustal composition from compressional wave velocities are nonunique, but this ambiguity is often reduced by complimentary shear wave velocity (and Poisson's ratio) observations. Recent field studies have found that seismic anisotropy is an important crustal feature. Thus systematic laboratory studies of compressional wave anisotropy and shear wave splitting will be critical in understanding crustal composition and deformation, just as they have been in investigations of the upper mantle.

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Cross-references

Deep Seismic Reflection and Refraction Profiling
Seismic Reflectivity Method
Seismic Anisotropy
Seismic Imaging
Seismic Travel Time Tomography
Vertical Seismic Profiling

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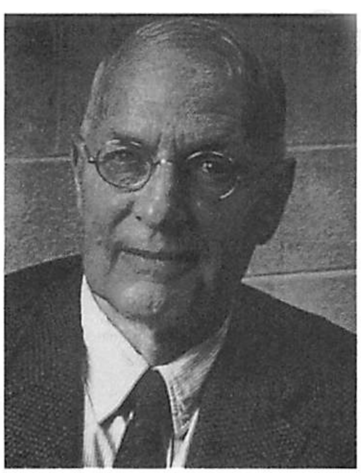
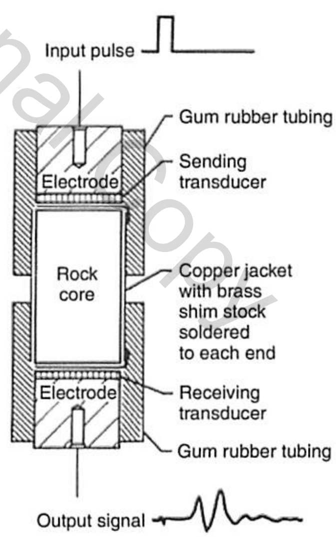
t1.1 **Seismic Properties of Rocks, Table 1** Average compressional (Vp) and shear (Vs) wave velocities, velocity ratios (Vp/Vs), and Poisson's ratios (σ) at 1 GPa for common rock types Christensen (1996)

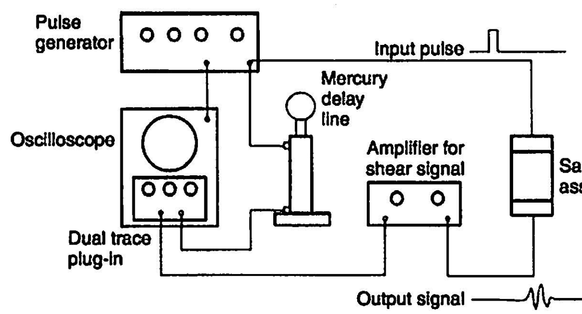
	Rock	Vp (km/s)	Vs (km/s)	Vp/Vs	σ
t1.2					
t1.3	Serpentinite	5.607	2.606	2.152	0.36
t1.4	Andesite	5.940	3.177	1.870	0.30
t1.5	Quartzite	6.091	4.054	1.502	0.10
t1.6	Basalt	6.118	3.291	1.859	0.30
t1.7	Granitic gneiss	6.271	3.627	1.729	0.25
t1.8	Granite-Granodiorite	6.372	3.726	1.710	0.24
t1.9	Tonalite gneiss	6.366	3.636	1.751	0.26
t1.10	Slate	6.379	3.432	1.858	0.30
t1.11	Phyllite	6.398	3.608	1.774	0.27
t1.12	Mica quartz schist	6.523	3.654	1.785	0.27
t1.13	Zeolite facies basalt	6.530	3.493	1.869	0.30
t1.14	Diorite	6.675	3.756	1.777	0.27
t1.15	Diabase	6.814	3.766	1.809	0.28
t1.16	Greenschist facies basalt	6.983	3.955	1.766	0.26
t1.17	Marble	6.985	3.794	1.841	0.29
t1.18	Mafic granulite	7.000	3.849	1.818	0.28
t1.19	Amphibolite	7.046	3.987	1.767	0.26
t1.20	Anorthosite	7.124	3.717	1.917	0.31
t1.21	Gabbro	7.299	3.929	1.858	0.30
t1.22	Pyroxenite	7.935	4.519	1.756	0.26
t1.23	Eclogite	8.198	4.594	1.785	0.27
t1.24	Dunite	8.399	4.783	1.756	0.26



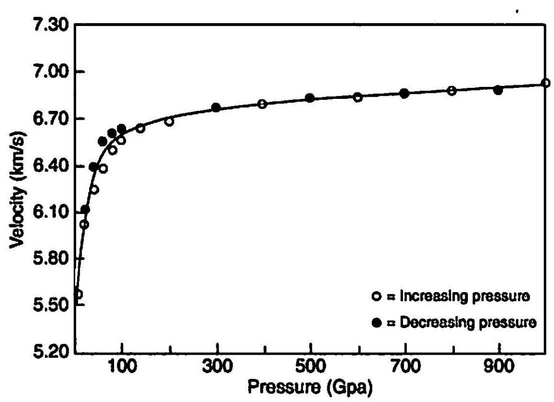
Seismic Properties of Rocks, Figure 1 Francis Birch (1903–1992), a pioneer in rock physics research.



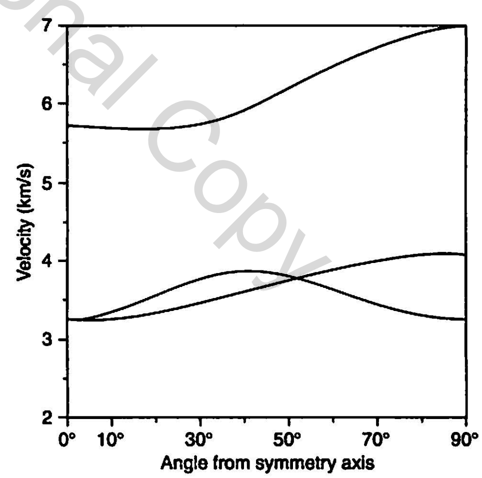
Seismic Properties of Rocks, Figure 2 Transducer and rock core assembly for velocity measurements at elevated pressures.



Seismic Properties of Rocks, Figure 3 Electronics for velocity measurements using a mercury delay line.



Seismic Properties of Rocks, Figure 4 Compressional wave velocity measurements as a function of confining pressure for a mafic granulite.



Seismic Properties of Rocks, Figure 5 Compressional wave anisotropy (*upper curve*) and shear wave splitting (*lower curves*) for a transversely isotropic quartz-mica schist.