Chapter 7

Laboratory techniques for determining seismic velocities and attenuations, with applications to the continental lithosphere

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ABSTRACT

The interpretation of seismic data is dependent on laboratory investigations of the elastic and anelastic properties of rocks. Important constraints on the composition of the continental lithosphere have been provided by comparing laboratory and field determined seismic velocities. Much less is known about the nature of attenuation in the crust and upper mantle. Furthermore, laboratory attenuations have not been studied as extensively as velocities. Nevertheless, the utilization of laboratory attenuation measurements to tie seismic data to the anelastic properties of rocks is promising. Laboratory velocity measurements are usually obtained with a pulse transmission technique, whereas measurements of seismic attenuation in the laboratory are determined by resonance techniques, ultrasonic pulse propagation, stress-strain hysteresis loop analysis, and torsion-pendulum oscillations. Velocities in rocks increase with increasing pressure, whereas attenuations decrease. The greatest changes, which occur over the first 100 MPa, are attributed to the closure of microcracks. Velocities decrease with increasing temperature. At temperatures below the boiling point of a rock's volatiles, attenuation appears to be temperature-independent, and above this, attenuation decreases. Increasing pore pressure, which lowers compressional and shear-wave velocities in sedimentary and crystalline rocks, produces a marked increase in Poisson's ratio. The influence of pore pressure on attenuation in crystalline rocks is at present poorly understood. At high pressures, velocities are primarily a function of mineralogy, whereas mineralogy may only be of secondary importance for attenuation. Anisotropy, which is common in velocities of many crustal and upper-mantle rocks, may also be an important property of rock attenuation. Attenuation may vary significantly with frequency for saturated rocks, but appears to be frequency-independent for dry rocks. The effect of frequency on velocity is minimal.

INTRODUCTION

Although many disciplines have contributed significantly to our knowledge of the nature of the continental crust and upper mantle, none has provided resolution comparable to seismologic studies. Since 1910, when A. Mohorovičić first presented evidence for a major seismic discontinuity in the Balkans at a depth of approximately 30 km, crustal thickness has been largely determined by seismic refraction investigations. These studies have also provided geophysicists with worldwide information on compressional wave velocities at various crustal depths, as well as upper-mantle velocities. Recently, significant data have become available on crustal velocity gradients, velocity reversals, shearwave velocities, and anisotropy in the form of azimuthal variations and shear-wave splitting. Studies of the attenuation of seismic waves give additional insight into the nature of the continental lithosphere. In addition, large-scale reflection studies, which currently represent major geophysical efforts in many

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countries, continue to provide new and exciting information on crustal structure and the nature of the Mohorovičić discontinuity.

Crustal and upper-mantle velocities and their attenuations are used to infer mineralogy, porosity, the nature of fluids occupying pore spaces, temperature at depth, and present or paleolithospheric stress reflected by mineral and crack orientation. Reflections within the continental crust and upper mantle originate from contrasts of acoustic impedances, defined as products of velocity and density. The interpretation of this seismic data is dependent on detailed knowledge of the elastic and anelastic properties of rocks provided by laboratory investigations. This chapter briefly reviews some of the techniques and major findings of laboratory seismology that apply to the exploration of the continental lithosphere.

VELOCITY MEASUREMENTS

In 1960, Francis Birch published the first comprehensive study of laboratory seismic velocities in crystalline rocks. Subsequent work by Birch and his research group (e.g., Birch, 1961, 1972; Simmons, 1964; Christensen, 1965, 1966) provided a foundation for future studies of rock elasticity pertinent to lithospheric velocities. At present, several laboratories throughout the world are actively engaged in studies of the seismic properties of rocks.

Although a more elaborate pulse matching scheme described by Mattaboni and Schreiber (1967) has been used, the technique for velocity measurements commonly employed in these laboratories is similar to or slightly modified from the pulse



Figure 1. Rock sample assembly for high-pressure velocity measurements (Christensen, 1985).

transmission method described by Birch (1960). This consists of determining compressional or shear-wave traveltimes through cylindrical rock specimens. Transducers are placed on the ends of the rock core. The sending transducer converts the input, an electrical pulse of 50 to 500V and 0.1 to 10 μ sec 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 (Fig. 1). Once the system is calibrated for time delays, the traveltime through the specimen is determined directly on the oscilloscope or with the use of a mercury delay line (Fig. 2). 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 mercury delay line (see Christensen, 1985, Fig. 4) consists of a fixed and a moveable transducer mounted in a column of mercury. The signal from the delay line is displayed as a second trace on the oscilloscope. The distance between the two transducers is adjusted so that the delay line signal is superimposed on the signal from the sample. With proper calibration, the time of flight of the signal through the sample is then equivalent to that through the delay line; the rock velocity is simply calculated from the length of the mercury separating the transducers, the length of the sample, and the velocity of mercury.

Generally the rock specimens are cores 2.54 cm in diameter and 4 to 6 cm long. Bulk densities are calculated from dimensions and weights. For measurements at elevated pressures, the cores are jacketed with copper foil and rubber tubing to prevent highpressure oil from entering microcracks and pores (Fig. 1). The core ends are either coated with silver conducting paint, or a strip of brass is spot-soldered to the copper jacket at each end to provide an electrical ground for the transducers. For measurements at high temperatures and pressures where gas is the pressure medium, the samples are usually encased in stainless steel. Velocity measurements as functions of confining and pore pressure require an additional pressure generating system and a more elaborate sample jacketing procedure, since the pore fluid must be isolated from the confining pressure fluid.

ATTENUATION MEASUREMENTS

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 continental lithosphere. Because of this potential, an expanding body of laboratory work has concentrated on bringing to fruition the diagnostic capabilities of attenuation measurements.

Unlike velocity determinations, there is no common method used for measuring seismic attenuation in the laboratory. There is not even a standard definition of attenuation (O'Connell and Budiansky, 1978). 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



Figure 2. Electronics for velocity measurements (Christensen, 1985).



Figure 3. An example of a resonant bar apparatus (Born, 1941).

logarithmic decrement δ . These are related for low-loss materials (Q > 10) by

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

where V is the phase velocity and f is the frequency (Johnston and Toksoz, 1981). In both the field and laboratory, difficulties arise in separating the intrinsic dissipation of the rock, i.e., 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.

The methods used in determining the attenuation can be separated into three categories: resonance techniques, ultrasonic pulse propagation, and low-frequency methods. Each method is evaluated in terms of the ease with which accurate results are obtained as functions of pressure (confining and pore), temperature, strain amplitude, and frequency.

The resonant bar method has had the widest application. Its use in measuring velocity and attenuation extends back to the

earliest days of laboratory seismology. A diagram of the equipment used by Born (1941) is shown in Figure 3. The technique consists of fashioning a long, thin rod out of the material, clamping or suspending it at a vibrational node (typically the midpoint), and vibrating it at one of its resonant modes f_n . Extensional, torsional, and flexural motion may be employed, and the velocity for the first two is given by

$$V = \frac{2f_{n}L}{n} \quad n = 1, 2, ...,$$
(2)

where L is the length of the sample and n is the mode. Analogous to the Q of an electronic resonant circuit, the seismic Q is defined as

$$Q = \frac{f_n}{\Delta f},$$
 (3)

where Δf is the frequency spread between the 3 dB points below the resonance peak f_n . The attenuation may also be calculated from the decay of resonance after the driving force has been removed, which is characterized by the time constant of resonant decay τ (similar to that of an R-C circuit), and Q is then determined from

(

$$Q = \tau \pi f_n. \tag{4}$$

Practical considerations abound in the application of the resonant bar technique. The ratio of length to radius for the bar should be at least 10 to 1 (Tittmann, 1977); thus a 2.5-cm sample radius necessitates a length of at least 0.25 m, and a bar closer to 1 m in length is preferable. This also means that the sample must be well consolidated, have grain sizes smaller than the radius, and be relatively free of fractures. Such constraints impose a bias on the measurements. Application of confining and possibly pore pressure requires jacketing the sample, and this will change the resonant frequency of the rod. Damping of the resonance by the confining fluid lowers Q in the extensional and flexural cases (not significantly for torsional [Birch and Bancroft, 1938]); for this reason, a gas (such as helium because its long sonic wavelengths avoid standing waves in the pressure vessel [Gardner and others, 1964]) is often used as the confining pressure medium. An abundance of transducer configurations for exciting and measuring the vibrations have been used, with piezoelectric, electromagnetic, and electrostatic transducers being applied in many novel ways. Losses in the transducers may be significant. Mounting transducers on the ends of the sample affect the resonance peak, as will the end caps required for measurements at elevated pressures. This shift is purposely employed by Tittmann (1977) to end load the rod in the flexural mode with copper blocks, thereby lowering the resonant frequency. Most if not all of the above problems can either be correct or made negligible in determining the intrinsic Q of the sample.

The vibrational spectra of spherical rock specimens have also been employed to determine Q. Used by Birch (1975) and Mason and others (1978) on rocks, this technique has the distinct advantage of providing a wide spectrum of vibrational modes, as opposed to the limited number of modes available in any given rod configuration. The method consists of placing a sphere between a sending and receiving transducer and sweeping a variable-frequency oscillator very slowly through the frequency range. In his measurements, for example, Birch (1975) covered frequencies between about 25 and 250 kHz. Peaks corresponding to various modes are observed, and Q is calculated from equation (3). Tone bursts may also be applied with the subsequent decay of resonance giving Q (equation (4)). The spherical resonance method is tedious to implement because: (1) spheres of rock are difficult to fashion and are usually not isotropic, (2) support of the sphere at a vibrational node is attainable only for torsional modes, and (3) it is troublesome to apply pressure and temperature beyond room conditions. Birch also found that this method gives erratic results due to noise and asymmetrical peaks.

Resonance methods can routinely accommodate variable frequency attenuation measurements. They have been used in the 0.1- to 100-kHz frequency range, with most of the data occurring between 1 and 20 kHz. Variable strain has been analyzed directly (Winkler and Nur, 1982; Murphy, 1982) and indirectly (Johnston and Toksoz, 1980b). Temperature dependence has been measured under room conditions (Kissell, 1972; Attewell and Brentnall, 1964; Gordon and Davis, 1968) and at elevated pressures (pore and confining) (Jones and Nur, 1983). Measuring attenuation with resonance methods under pressure is fairly involved; it has been carried out by Gardner and others (1964) and Katahara and others (1982), for example.

Ultrasonic pulse propagation has been used only recently for attenuation measurements in rocks, although its use in nondestructive evaluation attenuation measurements in materials research extends back to the early 1950s (Roderick and Truell, 1952). Through-transmission uses separate source and receiver piezoelectric transducers on opposite ends of a cylindrical rock sample, similar to that described above for velocity measurements. Pulse-echo uses a single transducer for transmitting and detecting the signal, with the one or more echos observed on the oscilloscope having been reflected from the rock core face opposite the transducer. Unlike the resonance methods, no peak is observed from which to determine Q; rather, it is assumed that the amplitude reduction of the pulse with distance A(x) can be expressed in terms of a damped exponential

$$A(x) = A_0 e^{-\alpha x}$$
(5)

where A_0 is the initial amplitude and α is the attenuation coefficient; α is therefore determined in wave-propagation measurements, and equation (1) relates this to a Q value.

An advantage of wave-propagation methods is the comparative ease with which high-pressure measurements may be performed (Johnston and Toksoz, 1980a; Winkler, 1983, 1985; Ramana and Rao, 1974). Few variable strain and temperature attenuation measurements have been performed using pulse propagation. Measuring attenuation as a function of frequency is difficult for most ultrasonic pulse methods; O values are typically reported at a single frequency, and when they are not, they are strictly band-limited. Furthermore, the frequencies used are all above 100 kHz and are normally about 1 MHz, placing them well outside the range of seismic field measurements. Only the technique of Winkler and Plona (1982) is designed to deliver Q as a function of frequency. Some or all of the following assumptions are made when ultrasonic wave propagation is used: Q is either frequency independent or slowly varying over the frequency range of interest, diffraction, beam spreading, coupling and transducer losses, and scattering can be corrected for or made negligible, and the ends of the sample are flat and parallel.

Through-transmission has been used in conjunction with spectral ratios (Toksoz and others, 1979; Sears and Bonner, 1981) and pulse broadening (Gladwin and Stacey, 1974; Ramana and Rao, 1974) to determine the attenuation of rocks. The spectral ratio technique is similar to the attenuation estimation procedure used in field seismology. When the output signal amplitude from sample, A_{sample} , is compared to a reference standard, A_{ref} , with an extremely high Q (such as aluminum, $Q > 10^5$), a linear equation is derived; plotting ln (A_{ref}/A_{sample}) vs. frequency yields a straight line whose slope equals $\pi L/(QV)$, where L is the sample length and V is the phase velocity. In addition to the above assumptions, this procedure assumes that the reference and sample assemblies are identical and that $Q_{ref} >>> Q_{sample}$.



Figure 4. Pulse-echo sample assembly (Winkler and Plona, 1982).

Based on the constant Q assumption, Q may be derived from pulse broadening with increasing traveltime. Only the first arrival and the subsequent peak are needed for the Q calculation, thus making it easy to use on almost any wavetrain. A precise pick of the wave arrival is one of the primary difficulties in applying this method, and it has not seen extensive use.

Pulse-echo experimental procedures have implemented spectral analysis, as well as the exponential decay of multiple reverberations. Figure 4 shows the sample assembly employed by Winkler and Plona (1982) for spectral ratio attenuation determination. Two wavelets, the first reflected from the sample's front face and the second from the back, are separately windowed, and the phase and amplitude spectra are calculated for each. The phase velocity $V(\omega)$ is given by:

$$\mathbf{V}(\boldsymbol{\omega}) = \frac{2\boldsymbol{\omega}\mathbf{L}}{\Delta\boldsymbol{\phi}},\tag{6}$$

where L is the sample length and $\Delta \phi$ is the phase difference between the two pulses at frequency ω . The attenuation coefficient α , in decibels per centimeter, is calculated using

$$\alpha(\omega) = \frac{8.686}{L} \ln \left[\frac{A_1(\omega)}{A_2(\omega)} \left(1 - R_{12}^2(\omega) \right) \right]$$
(7)

where A is the amplitude, R is the reflection coefficient, and the subscripts on A indicate the first or second wavelet. The primary advantage of this method is that transducer problems are effectively suppressed, since the same one is both the source and the re-

ceiver. In addition, Q may be measured as a function of frequency, and the coupling buffer provides the reference (high Q) material. Problems include the need to find the reflection coefficient as a function of frequency $R(\omega)$, choosing an adequate buffer material, and the application of diffraction corrections. No use of this technique on igneous or metamorphic rocks has yet been reported. The amplitude reduction of echos entails fitting a damped exponential $A_0e^{-\alpha x}$ to the peaks of the decaying multiples, thereby determining α (Peselnick and Zeitz, 1959). An example of the echos obtained by Mason and Kuo (1971) is shown in Figure 5. This technique requires high Q and/or short samples since a number of multiples are needed for a good estimate of the attenuation.

Low-frequency methods seek to place attenuation measurements within the frequency range of field seismology. Two general approaches have been taken to achieve this: stress-strain hysteresis loop analysis, and the torsion pendulum. Stress-strain curve analysis uses cyclic loading and unloading of a rock sample to produce a stress vs. strain hysteresis loop, and Q is determined from the enery lost per cycle (Usher, 1962; Gordon and Davis, 1968: McKavanagh and Stacey, 1974; Brennan, 1981). The torsion pendulum requires fashioning a long, thin, cylindrical rod of rock, attaching a mass with a high moment of inertia to the lower end of the rod, and suspending it vertically from the upper end. Torsional oscillations are induced, and after the driving force is shut off, the amplitude decay of free vibrations gives the attenuation (Peselnick and Outerbridge, 1961; Woirgard and others, 1977; Murphy, 1982, 1984). A variant of this is the flexion pendulum of Woirgard and Guegen (1978) in which a vertical rod of rock is clamped at the bottom and weighted at the top, with flexural vibrations being used to determine Q.

These techniques yield attenuation data at frequencies comparable to those used in the field (0.001 to 1 Hz for stress-strain;



Figure 5. Pulse-echo decaying multiples. Time scale is 5 μ sec per division (Mason and Kuo, 1971).



Figure 6. Compressional velocity vs. pressure for two specimens of granite (Birch, 1960). Velocity error bar of 1 percent is shown.

1 to 500 Hz for the torsion pendulum). Variable strain is easily incorporated, and use in a pressure vessel is discussed by Gordon and Davis (1968). Temperatures to 1,100°C were applied by Woirgard and Guegen (1978) to the flexion pendulum. In general, though, pressure-temperature (P-T) conditions in the earth are difficult to simulate with the low-frequency methods. Brennan and Stacey (1977) and others have shown that linearity for applied stresses (principle of superposition) holds only when the strain amplitude is less than 10^{-6} , and thus the application and detection of ultra-low stress and strain, respectively, is another source of problems.

DATA PERTINENT TO CRUSTAL AND UPPER-MANTLE COMPOSITION

Over the past three decades a large number of laboratory measurements have been reported (for a summary of velocity data, see Christensen, 1982). These have provided a basic understanding of many factors that influence velocity and attenuation of rocks believed to be abundant constituents of the continental lithosphere. The effects of some of these parameters are briefly reviewed in this section.

Pressure and temperature

The effect of confining pressure on velocities has been reported in a number of investigations. An example of data for a typical crystalline rock is shown in Figure 6. The characteristic shape of the curve of velocity vs. pressure is attributed to the closure of microcracks. As can be seen from Figure 6, much of the closure takes place over the first 100 MPa. Velocities measured in crystalline rocks at pressures 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

TABLE 1. COMPRESSIONAL VELOCITY PRESSURE AND TEMPERATURE DERIVATIVES*

Rock	dV _p /∂T 10 ⁻³ km s ⁻¹ °C ⁻¹ at 200 MPa	dV _p /∂P km s ⁻¹ GPa ⁻¹ at 500-800 MPa		
Serpentinite, Mid-Atlantic Ridge	-0.68	0.45		
Granite, Cape Ann, Massachuse	etts -0.39	0.25		
Quartzite, Baraboo, Wisconsin	-0.45	0.30		
Granulite, New Jersey Highlands	s -0.4 9	0.20		
Granulite, Saranac Lake, New Y	ork -0.51	0.21		
Basalt, East Pacific Rise	-0.39	0.16		
Gabbro, Mid-Atlantic Ridge	-0.57	0.20		
Granulite, Adirondacks, New Yor	r k -0 .60	0.28		
Amphibolite, Indian Ocean	-0.55	0.28		
Anorthosite, Lake St. John, Que	bec -0.41	0.20		
Granulite, Valle d'Ossola, Italy	-0.52	0.25		
Dunite, Twin Sisters, Washington	n -0.56	0.23		
Eclogite, Nove' Dvory,				
Czechslovakia	-0.53	0.19		
*Christensen, 1979.	, <u>, , , , , , , , , , , , , , , , , , </u>			

has originated from anistropic thermal contraction of the minerals. The pressure derivatives of velocity for several rocks and minerals are given in Table 1. These values are probably reasonable for porosity-free rocks at lower crustal and upper mantle

depths. Fewer data are available on the influence of temperature on rock velocities. It has been well known since the early work of Ide (1937) that the application of temperature to a rock at atmospheric pressure results in the creation of cracks that often permanently damage the rock (Fig. 7). 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 (for example, see Birch, 1943; Kern, 1978; Christensen, 1979). An example of data showing the influence of temperature on velocities is shown in Figure 8. Temperature derivatives of velocities for some common rocks and minerals are tabulated in Table 1.

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:

$$\frac{\mathrm{d}V}{\mathrm{d}z} = \left(\frac{\partial V}{\partial P}\right)_{\mathrm{T}} \frac{\mathrm{d}P}{\mathrm{d}z} + \left(\frac{\partial V}{\partial \mathrm{T}}\right)_{\mathrm{P}} \frac{\mathrm{d}\mathrm{T}}{\mathrm{d}z} \tag{8}$$

where z is depth, T is temperature, and P is pressure. For regions with normal geothermal gradients $(25^{\circ} \text{ to } 40^{\circ}\text{C/km})$, the change





Figure 7. Compressional velocity vs. temperature for granite at atmospheric pressure (Ide, 1937).

Figure 9. Shear-wave Q vs. pressure for several rocks (Katahara and others, 1982).

in compressional velocity with depth dV_p/dz is close to zero (Christensen, 1979). However, in high heat-flow regions, crustal velocity reversals are expected if compositional changes with depth are minimal.

All investigations have found that Q increases with increasing confining pressure. Beginning with Birch and Bancroft (1938), every experimental technique has observed 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 vs. P curve is generally attributed, therefore, to the



Figure 8. Compressional velocity vs. temperature at 200 MPa (Christensen, 1979).

closure of microcracks. This is illustrated in Figure 9 for several crustal and upper-mantle rocks, and pressure derivatives of Q for these rocks are given in Table 2. 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 (Johnston and Toksoz, 1980b; Tittmann and others, 1974). At the onset of partial melting, Q decreases (Spetzler and Anderson, 1968).

Pore pressure

The influence of pore pressure on velocity and attenuation has been widely studied for sedimentary rocks; however, only a limited amount of data are available for crystalline rocks. Raising the pore pressure has approximately the same effect on velocities as lowering the confining pressure by the same amount (Todd and Simmons, 1972; Christensen, 1984, 1986). This is illustrated in Figure 10 for a gabbro, where velocities are shown as functions of confining pressure at atmospheric pore pressure and constant differential pressure (confining pressure minus pore pressure). The dramatic lowering of velocities with increasing pore pressure appears to be a common feature for crystalline continental rocks and may be one of many possible explanations for crustal lowvelocity zones. Of significance, increases of pore pressure in

Rock	∂Q _s /∂P (MPa ⁻¹ at 300-500 MPa)		
Shikoku Eclogite	1.08		
Westerly Granite	1.50		
Duluth Gabbro	0.65		
San Carlos Gabbro	0.17		
Twin Sisters Dunite	0.57		

TABLE 2. ATTENUATION PRESSURE DERIVATIVES*

crustal regions will be accompanied by marked increases in Poisson's ratios (Christensen, 1984). Although pore pressure attenuation data is sparse for crystalline rocks, the investigations on sedimentary rocks have found that fluid flow in microcracks is the mechanism responsible for the observed drop in Q with increasing saturation (Winkler and Nur, 1979; Clark and others, 1980). This effect should be seen in crystalline rocks as well.

Mineralogy

When pores and cracks are closed, velocities are primarily a function of mineralogy. For many suites of continental rocks, systematic changes in velocities with mineralogy have been observed (Birch, 1961; Christensen, 1965, 1966; Christensen and Fountain, 1975). In general, rocks rich in alkali feldspar, quartz, and mica have compressional wave velocities below 6.5 km/sec, whereas the mafic rocks with Ca-rich feldspar, amphibole, and pyroxene have compressional wave velocities close to 7.0 km/sec. Unaltered ultramafic rocks and eclogites usually have velocities in excess of 8.0 km/sec. A more detailed discussion of estimates of crustal and upper-mantle mineralogy from seismic velocities is presented in another chapter (Fountain and Christensen, this volume).

Vp/Vs ratios appear to be diagnostic of several common rock types (Christensen, 1972; Christensen and Fountain, 1975; Christensen and Salisbury, 1975). For example, Vp/Vs ratios are generally low in quartz-rich rocks and high in anorthosites and serpentinites.

Attenuation is dependent on the density and shape of microcracks and their state of saturation. Mineralogy may be only of secondary importance, although higher velocity rocks usually have higher Q values (Johnston, 1981). Additional studies are needed to further explore the relationship between attenuation and mineralogy.

Seismic anisotrophy

The interpretation of seismic data is greatly simplified if it is assumed that rocks beneath the Earth's surface behave as iso-



Figure 10. Velocities as functions of confining pressure and differential pressure *Pd* (Christensen, 1986).

tropic solids. Laboratory studies, however, clearly show that most rocks are anisotropic. Early laboratory measurements of compressional wave velocities from cores cut in three mutually perpendicular directions demonstrated that common continental crustal metamorphic rocks often have anisotropies greater than 10 percent (Birch, 1960; Christensen, 1965). In addition, it was observed that shear waves split into components with different polarizations and different velocities (Christensen, 1966). This shear-wave splitting, also referred to as acoustic double refraction and shear-wave birefringence, is observed in the laboratory by rotating the polarization directions of the shear-wave transducers relative to the rock fabric (Fig. 11).

Anisotropy originates from a number of mechanisms, the two most important being preferred orientation of minerals and alignment of cracks. Common continental crustal rocks such as schists, gneisses, and amphibolites are anisotropic due to orientation of micas and amphiboles (Christensen, 1965). In the continental upper mantle, olivine and pyroxene alignments are likely to produce anisotropy. At confining pressures less than approximately 20 MPa, laboratory anisotropy measurements are influenced by crack orientation as well as mineral orientation. In schistose rocks, crack orientation usually enhances mineral anisotropy at low pressures (Fig. 12). In addition, anisotropy is likely to originate from aligned stress-induced cracks in dilatancy zones in seismic regions (Crampin, 1978; Crampin and others, 1980; Crampin and McGonigle, 1981). Experimental studies of the effects of this phenomenon on seismic anisotropy are limited (Tocher, 1957; Nur, 1971; Gupta, 1973).

It is generally assumed, at least implicitly, that attenuation is independent of direction, and a dearth of experimental data on Q



Figure 11. Oscilloscope traces for shear wave propagation through (a) slate and (b) dunite. The transducers are oriented at 0° to receive the higher velocity shear wave. Rotation of the transducers through 90° emphasizes the first arrival of the slower of the two shear waves (Christensen, 1971).

anisotropy neither confirms nor refutes this assumption. What little work that has been done (a taconite by Singh [1976], a granite by Lockner and others [1977], an oil shale by Johnston and Toksoz [1980a], and a deep-sea carbonate by Kim and others [1983]), however, points to a directional dependence of Q for some rocks. More measurements on igneous and metamorphic specimens are required to determine if Q anisotropy is an important factor in the Earth's crust.

Velocity-density relations

The dependence of velocity on density has been of interest in crustal gravity investigations where seismic data are also available. Several velocity-density relations for common rocks are listed in Table 3. These are usually expressed as a linear solution in the form $V = a + b\rho$, although for some rock suites a nonlinear solution may be more appropriate.

It is now understood that a single value dependence of velocity on density is valid only for limited compositions. Birch (1961) found that, to a first approximation, the compressional wave velocity can be expressed in terms of density and mean atomic weight. Mean atomic weights (m) were calculated from standard rock chemical analyses by

$$\mathbf{m} = (\Sigma \mathbf{x}_i \mathbf{m}_i)^{-1} \tag{9}$$

where x_i is the proportion by weight of the *i*th oxide in the rock, and m_i is the mean atomic weight of the oxide (the formula

TABLE 3. VELOCITY-DENSITY LEAST-SQUARES SOLUTIONS OF THE FORM V = a + $b\rho^*$

Data Set	Pressure, MPa	a km/sec	b <u>km/sec</u> g/cm ³	r ² , %	Reference [†]
$V_{\rm n}$ (all rocks), m ≈ 21	10 ³	-2.01	3.16	94	1
V_s (all rocks), m \cong 21	10 ³	-0.88	1.63	92	2
V _p , basait	10 ²	-4.44	3.64	95	3
V _s , balsat	10 ²	-2.79	2.08	94	3

*V_p in km/sec, ρ in g/cm³, r = correlation coefficient.

[†]1. Birch, 1961; 2, Christensen, 1968; 3. Christensen and Salisbury, 1975.

weight divided by the number of particles in the formula). Most common rocks have mean atomic weights between 20 and 22, with the highest values for relatively iron-rich rocks. A velocity-density solution for m = 21 given by Birch (1960) and a similar calculation for shear velocities (Christensen, 1968) are presented in Table 3.

Frequency

Field-measured seismic velocities and attenuations are generally lower than those measured from sonic logs; these, in turn, are lower than ultrasonic velocities. The effect of frequency on velocity is slight, however, and is usually on the order of the uncertainty in the laboratory measurements (Birch, 1961). The bulk of the attenuation data indicates that Q is frequency-independent for dry rocks; for saturated and partially saturated samples, however, Q may vary significantly with frequency (Born, 1941; Wyllie and others, 1962; Tittmann and others, 1981).

CONCLUSIONS

Beginning with the pioneering work of Birch (1960), much has been learned about elastic wave propagation in rocks. Velocities are now available for most common rock types at pressures that exist in the continental crust and upper mante. Additional important contributions include laboratory measurements of shear-wave velocities, Vp/Vs ratios, and the influence of temperature and pore pressure on velocities.

For continental lithospheric studies, it is desirable in the future to obtain additional velocity data at high temperatures. This is especially true for shear-wave measurements. As more detailed crustal velocity data become available, it may become apparent that seismic anisotropy is an important crustal feature. Thus, systematic laboratory studies of compressional wave anisotropy and shear-wave splitting could be critical in understanding crustal composition, just as they have been in investigations of the



Figure 12. Seismic anisotropy [(V_{max}-V_{min}]/V_{avg}) as a function of pressure for a quartz-mica schist.

upper oceanic mantle. In lower crustal regions, anisotropy resulting from mineral orientation is likely, whereas in the shallow crust, crack anisotropy is thought to be important.

Many researchers have obtained high-quality attenuation data, and a number of loss-mechanism theories have proven quite successful. Although a variety of methods have been employed to measure O, certain trends are evident. The magnitude may vary from one study to the next, but the relative effects of pressure, temperature, and saturation state are fairly consistent, and efforts to relate these conditions to the micro- and macrostructure of rocks should continue. Also, it would seem that the behavior of Q parallels some already well-established velocity characteristics such as dependence on confining and pore pressure and frequency. This raises some interesting questions: Is Q characterized by mineralogy at high pressures rather than cracks and their saturation condition? Is there a Q-density relation similar to the velocity-density relationship mentioned above? Is attenuation anisotropic in crustal and upper-mantle rocks? Are Qp/Qs ratios diagnostic, as suggested by Winkler and Nur (1979)? Is an effective pressure law applicable for Q? Finally, far too little laboratory attenuation work has been done on crystalline rocks. This must become a priority if significant use is to be made of seismic attenuation data of the continental crust and upper mantle.

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