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Composition of the continental crust and upper mantle; A review

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ABSTRACT

At present there are many ambiguities involved in arriving at reasonable models for crustal and upper-mantle compositions. We review several approaches previously employed by geologists and geophysicists to estimate bulk chemical and mineralogic compositions. Recent studies focusing on the petrology of xenolith suites and geologic mapping of high-grade metamorphic massifs, such as the Ivrea zone, support the thesis of heterogeneity in crustal composition. Even though crustal composition varies laterally, there is strong evidence pointing to the importance of metamorphic grade, which generally increases with depth, in controlling crustal petrology. Seismic refraction and reflection methods show the most promise for understanding the lateral variability in petrology. Laboratory studies of the seismic properties of rocks at crustal and upper-mantle pressure and temperature conditions show that seismic data can provide valuable constraints on crustal and upper-mantle composition. Seismic anisotropy is likely an important property of the continental lithosphere at all levels. Field and laboratory experiments carefully designed to investigate this directional dependence of seismic velocities will provide valuable constraints on the fabric and composition of the continental crust and upper mantle. Within the upper crust, physical properties are likely to be strongly influenced by the presence of fractures containing fluids at high pore pressures. A model for the continental crust and upper mantle, emphasizing probable extreme lateral variability, is constructed from information available on exposed deep crustal sections, xenoliths, and laboratory and field seismic studies.

INTRODUCTION

Models depicting the petrology, chemistry, and structure of the continental crust and upper mantle are currently being refined by integration of new geophysical data with the powerful constraints offered by geological investigations. Geological observations that provide important constraints on crustal composition include studies of xenoliths and high-grade metamorphic terrains now exposed on the Earth's surface. Geophysical methods most useful for understanding the nature of the crust and upper mantle are primarily seismic refraction and reflection profiling with important additional insights coming from electrical and magnetic investigations. When coupled with experimental data on seismic properties of continental rocks at pertinent temperatures and pressures, the geological and geophysical data can be used to correlate field measurements of seismic velocities with mineralogic composition at depth.

This chapter reviews the geologic and geophysical evidence pertaining to the chemistry and mineralogy of the continental crust and upper mantle with the goal of developing better constrained models of these horizons of the Earth. We first summarize previous estimates of chemical compositions of the continental crust. Evidence for the mineralogy of the continental lithosphere based on significant observations from xenolith suites

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is reviewed. The composition of the intermediate and deep levels of the continental crust is discussed in terms of observations of exposed crustal cross sections. Next, limitations that laboratory seismic data place on the estimation of crustal and upper-mantle composition are examined in detail. Finally, seismic constraints on crustal and mantle composition are summarized, and a model for the petrology and structure of the continental crust and upper mantle is proposed.

ESTIMATES OF CRUSTAL CHEMISTRY

Ever since recognition of the Mohorovičić discontinuity in the early part of the century, earth scientists have devised a variety of quantitative estimates of the chemical composition of the continental crust (Table 1 [Tables 1 through 9 are on pages 728 through 736, at the end of this chapter.]). Early estimates were based on the premise that the compositional average of crystalline rocks exposed at the surface of the continents represented the bulk composition of the entire continental crust (Clarke, 1924; Clarke and Washington, 1924; Goldschmidt, 1933, 1954). With the discovery of the Conrad discontinuity in Europe in 1925, however, scientists recognized that the composition of the deeper part of the continental crust may differ significantly from the upper crust. Later estimates, therefore, were predicated on the assumption that crustal composition could be approximated by a mixture of granitic and basaltic compositions (Poldervaart, 1955; Vinogradov, 1962; Taylor, 1964; Pakiser and Robinson, 1966; Ronov and Yaroshevsky, 1967, 1969; Galdin, 1974). Some of these estimates incorporated important geophysical and geochemical data to constrain the bulk composition. Pakiser and Robinson (1966) correlated seismic velocity with rock composition and used the result to estimate composition from seismic refraction data. Taylor (1964) fit the rare-earth element pattern of sediments by a mixture of equal parts of basalt and granite and consequently assumed this reflected the relative abundance of those two components in the crust. Laboratory velocity studies, however, demonstrated compositional models based on igneous rock abundances may be in error because seismic velocities in the upper and lower crust also correspond with a wide range of metamorphic compositions (Christensen, 1965).

More recent estimates were derived either from models of crustal growth or models of crustal structure based on geology of deep terrains. Taylor (1977, 1979) and Taylor and McLennan (1981) modified their models by assuming that 75 percent of the present-day crust was composed of Archean crustal material and the remaining 25 percent was derived by island-arc magmatism. Using whole-rock analyses of xenoliths from the Moses Rock dike, McGetchin and Silver (1972) estimated chemical composition of the crust beneath the Colorado Plateau. Holland and Lambert (1972) assumed that the Lewisian and Dalradian terrains in Scotland constituted a typical crustal section, and their average composition should be representative of the entire crust. Weaver and Tarney (1984a, b) used the upper crustal estimate presented by Taylor (1977) in conjunction with Lewisian terrain data from middle and lower crustal levels to derive their estimate. Smithson (1978) developed a chemical model based on mean crustal velocity that suggested a bulk crustal composition of quartz diorite.

The similarities between the results of crustal composition estimates are surprising, considering the wide variety of methods employed. In fact, the average SiO_2 percentage from all these estimates (60 ± 2.7 percent) is not substantially different from the earliest estimate of Clarke (1924). Certain researchers have obtained a more silicic crust (Smithson, 1978; Weaver and Tarney, 1984a,b), whereas others have estimated a more mafic crust (Pakiser and Robinson, 1966; McGetchin and Silver, 1972; Taylor, 1979; Taylor and McLennan, 1985). It is important to realize, however, that all these estimates are simply models based on a wide variety of simplified, generalized assumptions.

In addition to estimates of bulk crustal composition, some attempts have been made to estimate composition of the crustal seismic layers. Data presented in Table 2 summarize estimates of the composition of the lower continental crust. These values were derived from: (1) analyses of exposed deep-level metamorphic terrains (Lambert and Heier, 1968; Sheraton and others, 1973; Mehnert, 1975; Leyreloup and others, 1977; Smithson, 1978; Weaver and Tarney, 1980, 1984a; Maccarrone and others, 1983); (2) models of lower crustal genesis (Taylor, 1977, 1979; Taylor and McLennan, 1981, 1985); or (3) the assumption that the lower crust is mafic in composition (Poldervaart, 1955; Pakiser and Robinson, 1966). Because these estimates are model-dependent or are based on a variety of deep crustal analogs, it is not surprising that they differ significantly from one another.

Estimates of the composition of the upper crust are based primarily on extensive geochemical data of average chemical compositions of crystalline shield-area surface rocks (Sederholm, 1925; Grout, 1938; Poldervaart, 1955; Eade and others, 1966; Reilly and Shaw, 1967; Shaw and others, 1967; Condie, 1967; Fahrig and Eade, 1968; Ronov and Yaroshevsky, 1969; Eade and Fahrig, 1971; Bowes, 1972; Taylor and McLennan, 1985). These analyses are remarkably similar, yielding an average SiO_2 content of approximately 65 percent, and are well summarized by the upper crustal estimates presented in Table 3. Other workers estimated the petrology of the upper crust from geologic maps of shield areas (Tables 4, 5). This approach is important because it emphasizes the lithologic heterogeneity of the upper crust, an observation obscured in the averaging processes associated with development of geochemical data presented in Tables 1, 2, and 3, but evident in studies of crustal xenoliths and possible exposed crustal sections.

Evidence for Crust and Upper-Mantle Composition from Xenoliths

Xenolith suites provide an excellent catalog of rock types that potentially constitute the crust and upper mantle (e.g., Dawson, 1977; Kay and Kay, 1981; Griffin and O'Reilly, 1987). The geophysical properties of these rocks can be measured or theoret-

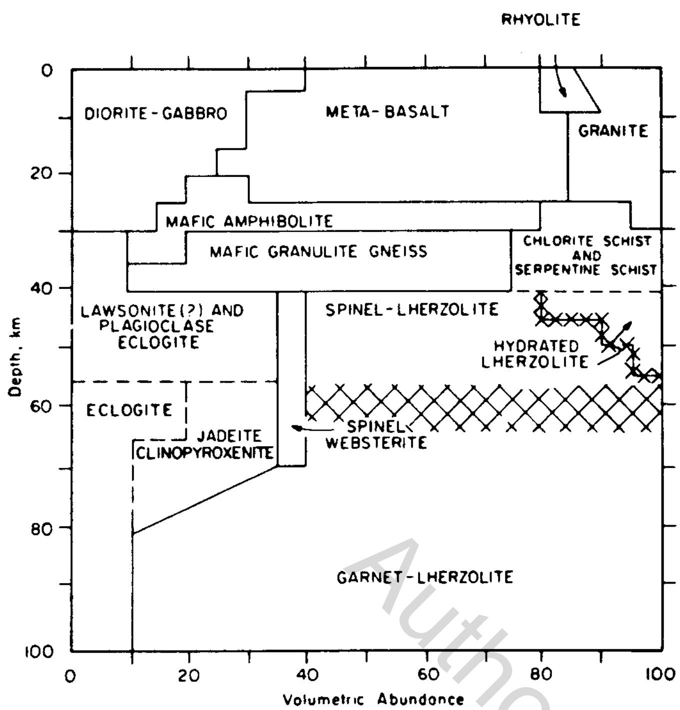


Figure 1. Model of the crust and upper mantle of the Colorado Plateau based on xenoliths (McGetchin and Silver, 1972).

ically calculated so that we can assess their potential importance as lithospheric constituents. The lithologies of xenoliths recovered from the numerous sites in North America are summarized in Table 6. Pressure-temperature (P-T) conditions of metamorphism are listed when reported. No attempt has been made to assess the volumetric abundance of each rock type or to evaluate the quality of the geothermometric and geobarometric data.

Evident in this table is the extent of the lithologic heterogeneity of the lithosphere under many localities. Some xenolith sites (e.g., State Line, central Montana, Navajo field, Moses Rock) exhibit diverse suites that indicate detailed vertical crustal and mantle heterogeneity must exist. In some cases, xenolith researchers have attempted to construct models of the crust and upper mantle by arranging xenoliths into a depth column in an order determined by various parameters such as P-T data, volumetric abundance, and xenolith size.

McGetchin and Silver (1972) presented one of the earliest crustal stratigraphy reconstructions based on xenoliths from the Colorado Plateau (Fig. 1). They envisioned an upper crust dominated by metabasalt, diorite, gabbro, and granite underlain at deep crustal levels by amphibolite, mafic granulite, chlorite schist and serpentine schist. Assignment of specific rock types to various crustal depths was determined by xenolith size. The upper mantle of the plateau was inferred to be lithologically complex, showing significant lateral and vertical heterogeneity. The presence of hydrated rocks at lower crustal levels was questioned by Padovani and others (1982), who, based on petrographic and scanning electron microscope (SEM) analysis, proposed that these rocks were hydrothermally altered during emplacement.

A complex upper-mantle model was presented for the Colorado-Wyoming Front Range (Fig. 2) by McCallum and Eggler (1976), who inferred that the mantle below the Moho was spinel peridotite with zones of garnet websterite. The crustal portion of this section was typified by a wide variety of rocks (Table 6) dominated by granulites and pyroxenites at the base overlain by upper crustal gneisses and granites. Brookins and Meyer (1974) presented a more detailed and complex model for the lithosphere beneath Kansas (Fig. 3). A variety of mafic rocks of generally high metamorphic grade constitute the lower crust and the crust-mantle transition. The upper mantle in this model is dominated by eclogites.

Because portions of the continental lithosphere have likely originated by accretion of island-arc terrains, xenoliths found in island arcs provide information on possible continental crustal composition as well as arc petrology. Compositions of Aleutian lava xenoliths summarized by Kay and Kay (1986) are given in Table 6. Based on xenolith studies and crystal fractionation calculations, Kay and others (1986) have estimated that the upper crust of the Aleutian arc consists of layered volcanic and volcaniclastic sediments intruded by plutons, whereas cumulates from arc rocks, early arc lavas, and original oceanic crust of this model

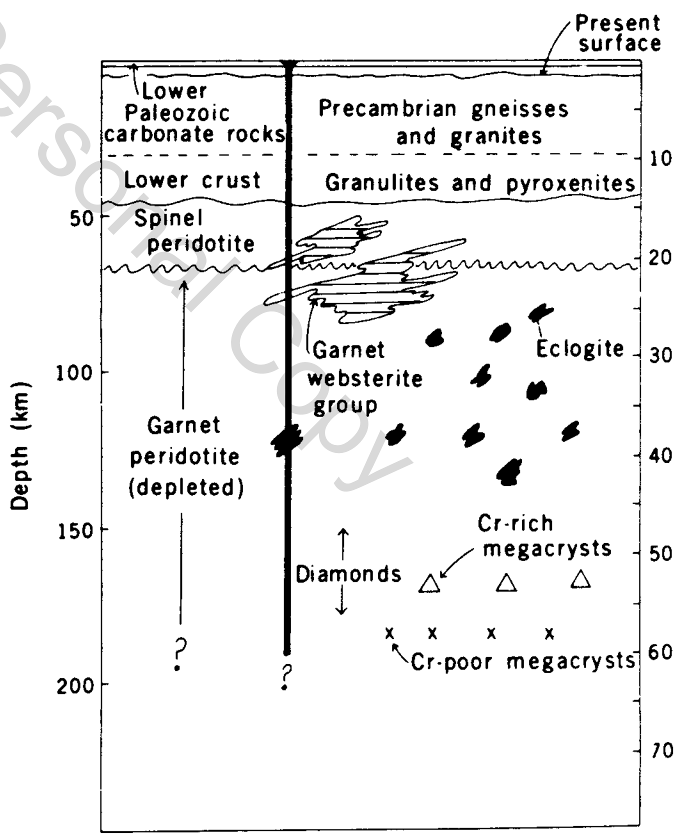


Figure 2. Schematic model of the crust and upper mantle beneath the Colorado-Wyoming Front Range area based on xenoliths (McCallum and Eggler, 1976). Heavy vertical line represents the kimberlite.

vary from basaltic to dacitic in composition and the lower crust is largely basaltic. Arc-related ultramafic cumulates form the uppermost mantle.

Xenoliths have distinct advantages for the study of crustal and upper-mantle composition because they represent samples of the lithosphere between the source of the transporting magma and the site of eruption, they commonly show contact relations between various rock types, and their pressure and temperature conditions of equilibrium can be derived from the compositions of coexisting minerals. They do not, however, show large-scale geometric relationships, they may represent a biased sample set, and they may not show equilibration along an ambient geotherm (e.g., Harte and others, 1981; Griffin and O'Reilly, 1987). In addition, recent isotopic studies show that certain deep crustal xenoliths recovered from Kilbourne Hole, New Mexico (Davis and Grew, 1978), and the Snake River Plain (Leeman and others, 1985) yield Proterozoic and Archean dates of peak metamorphism, respectively, indicating these xenoliths were at deep crustal levels during Precambrian peak metamorphism. Later tectonism may have elevated them to shallower crustal levels, implying these rocks were once in the deep crust or upper mantle but may not necessarily represent contemporary deep crust. The information provided by xenoliths on crustal composition is particularly useful when their lithologies are compared with portions of continental crustal sections now exposed on the Earth's surface.

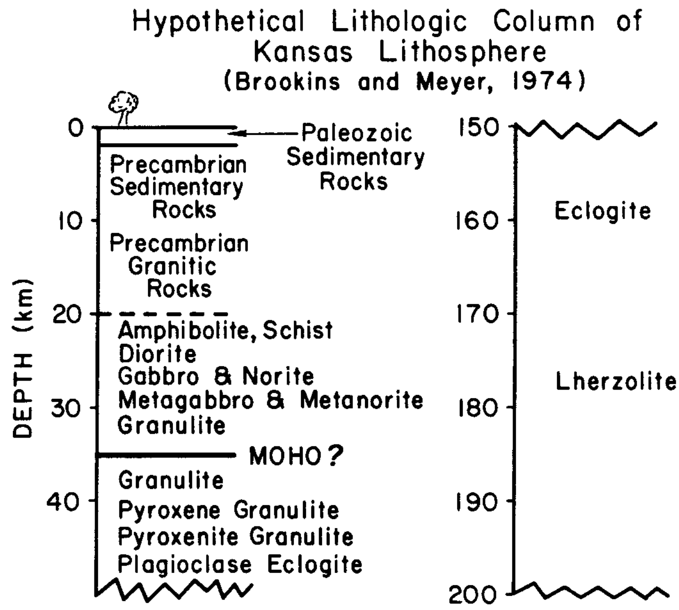


Figure 3. Hypothetical lithologic column for crust and upper mantle of Kansas based on xenoliths (Brookins and Meyer, 1974).

CRUSTAL STRUCTURE AND COMPOSITION BASED ON CRUSTAL CROSS SECTIONS

Several classic papers in the late 1960s (Berckhemer, 1968, 1969; Ansorge, 1968; Giese, 1968) suggested that the deep crust of the Po basin in Italy was thrust to the surface of the Earth and is presently exposed in the Ivrea zone of northern Italy. Evidence for this remarkable structure was found in a large positive Bouguer gravity anomaly and analysis of seismic refraction data showing a high-velocity slab projecting to near-surface depths. Laboratory seismic velocity data for Ivrea-zone rocks (Fountain, 1976) added support to the hypothesis. Later, Fountain and Salisbury (1981) postulated that analogous situations exist elsewhere in the world, and thus, there may be several localities where partial cross sections of the continental crust are exposed for direct observation. In addition to the Ivrea zone, examples of exposed crustal cross sections included the Sachigo-Pikwitonei subprovinces (Manitoba), the Fraser and Musgrave Ranges (Australia), and the Kasila series (West Africa). Several other possible crustal cross sections have recently been identified; these include the Kapuskasing structural zone in Ontario (Percival and Card, 1983), Fiordland in New Zealand (Oliver and Coggon, 1979; Oliver, 1980; Priestley and Davey, 1983), the Tehachapi Mountains in California (Ross, 1985), and Calabria in southern Italy (Schenk and Scheyer, 1978; Moresi and others, 1978-1979). In Table 7 we tabulate the major rock types of the crust based on several of these cross sections.

Two of these examples represent sections through the Archean Superior Province in Canada. Here, the upper crustal levels are dominated by a variety of tonalitic, granodioritic, and granitic rocks, which surround metavolcanic and metasedimentary packages (greenstone belts). In the Pikwitonei example (Table 7), the granite-greenstone terrain passes, with inferred increasing depth, into upper amphibolite to granulite facies tonalitic gneiss with inclusions of remnant greenstone belt lithologies. In contrast, deeper crustal levels in the Kapuskasing structural zone (Table 7) are characterized by various granulite facies rocks (mafic gneiss, tonalite gneiss, metasedimentary gneiss, anorthosite). Percival and Card (1985) have speculated that these rocks are remnants of the pre-greenstone belt crust.

The Ivrea section exhibits different structural and lithologic characteristics than those shown in the Canadian examples. Upper crustal levels are dominated by greenschist to amphibolite facies metasedimentary rocks and quartzo-feldspathic gneisses. Deep crustal levels are composed of upper amphibolite to granulite facies pelitic rocks, marble, amphibolite, gabbroic rocks, and garnet granulite. Mafic granulites dominate along strike of this zone. Ultramafic bodies occur at the base of the exposed section. The transition zone from mafic granulites to the ultramafic bodies has been interpreted as an exposed example of the Moho (Berckhemer, 1969; Hale and Thompson, 1982), but the boundary may be tectonic (e.g., Shervais, 1978-1979).

Geobarometric data for several of these sections indicate that the maximum depths exposed equilibrated at pressures no greater than 700 to 900 MPa. Thus between 3 and 10 km of lower crust is missing if one assumes a mean crustal thickness of 35 km. The crust-mantle transition is generally not observed, so we lack critical information about the nature of the Moho and the uppermost continental mantle. Also, the tops of these sections commonly contain greenschist facies assemblages, suggesting the uppermost crust is missing.

These sections illustrate the great lateral and vertical lithologic heterogeneity that may characterize the crust in one area and over continent-scale distances. This heterogeneity results from the complex, protracted and specific sequence of geologic events experienced by each crustal section. Thus, unlike similar oceanic crustal modeling based on ophiolites, we cannot draw many generalities about overall crustal composition from these sections, but rather we should analyze the crust of each region individually.

SEISMIC CONSTRAINTS ON CRUSTAL AND UPPER-MANTLE PETROLOGY

The models examined to date provide information on crustal and mantle petrology over limited regions. Refraction and reflection methods provide the most promise for understanding lateral variability in the continental lithosphere. Of critical importance in the interpretation of the seismic data are laboratory studies of the compressional (V_p) and shear (V_s) wave velocities of rocks at crustal and upper-mantle P-T conditions. In this section we review some of the more important results from these investigations that are relevant to the interpretation of seismic data, and the constraints that field seismic data place on the estimation of crustal and upper-mantle composition are examined in detail. In this discussion we consider the upper crust to make up the upper third to half of a crustal column. Traditionally, the lower crust is viewed as the portion of the crust below the Conrad discontinuity, but because this discontinuity is not always observed and many velocity profiles show several discontinuities, we regard the lower crust as the bottom third to half of the crust, as defined by seismic methods. In general, V_p is greater than 6.5 km/sec at these depths, and commonly greater than 6.8 km/sec.

Seismic properties of crustal and upper-mantle rocks

Summarized in Table 8 are compressional-wave velocity data for typical continental crystalline rocks that appear to be abundant based on xenolith studies and the petrology of exposed crustal cross sections. The data are tabulated as a function of confining pressure to, in most cases, 1 GPa (10 kbar). Certain samples were classified with reported modal mineralogy according to the IUGS system (Streckeisen, 1976).

For most rocks, the changes of velocity with confining pressure are well known. At low confining pressures, the gradients are large (Table 8), apparently due to the closure of low-aspect ratio microcracks (Birch, 1961; Thill and others, 1969). At confining pressures from 200 to 1,000 MPa, the pressure derivative of compressional wave velocity ($\partial V_p/\partial P$) approaches characteristic single crystal values and at higher pressures reflects the intrinsic elastic behavior of the constituent minerals (Christensen, 1974). Changes in velocity with pressure for isotropic harzburgites with nearly identical mineralogy from the Twin Sisters Range, Washington, and Kilbourne Hole, New Mexico (xenolith), are shown in Figure 4. The differences in velocities and initial behavior with increasing pressure are explained by the high porosity of the xenolith, which apparently originated during rapid ascent to the Earth's surface. Even at the high pressures simulating those of the lower crust and upper mantle, velocities of the xenolith are low because pressure alone is insufficient to close pore spaces. For this reason, we selected data for Table 8 from rocks collected from crystalline terrains that have not experienced this violent emplacement history.

Compared to measurements at elevated pressure, data on the influence of temperature on velocities are scarce. Recent measurements define the temperature derivatives of compressional wave velocity ($\partial V_p/\partial T$) for a few common crustal and upper-mantle lithologies (Meissner and Fakhimi, 1977; Ramananantoandro and Manghnani, 1978; Christensen, 1979; Kern and Richter, 1981; Kern and Schenk, 1985). In general, these studies show that $\partial V_p/\partial T$ typically ranges from 4 to 6×10^{-4} km sec^{-1} $°C^{-1}$, but larger and smaller values are reported. Large gradients reported by Christensen (1979) at temperatures greater than 300°C apparently result from crack enlargement associated with anisotropic thermal expansion of minerals at confining pressures of 200 MPa.

To illustrate this effect of temperature, we show the variation of V_p with depth in a granite and a mafic garnet granulite (Fig. 5) calculated on the basis of temperature and pressure derivatives presented in Christensen (1979) and with geothermal gradients given in Pollack and Chapman (1977). These gradients correspond to surface heat-flow values of 40 mW/m^2 (average shield) and 90 mW/m^2 (Basin and Range) and are dashed where extrapolated above 300°C. Also shown, for comparison, are the

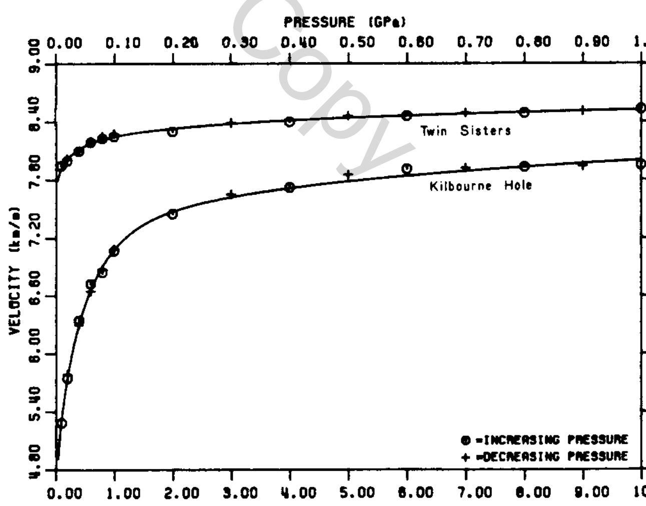


Figure 4. Compressional wave velocity vs. confining pressure for harzburgites from the Twin Sisters Range, Washington, and Kilbourne Hole, New Mexico (N. I. Christensen, unpublished data).

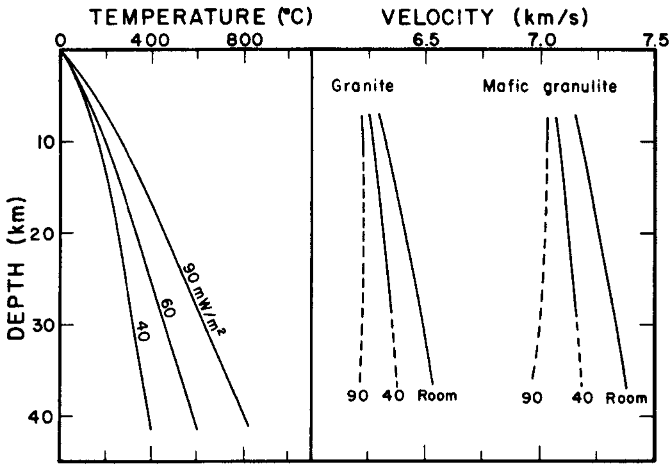


Figure 5. Calculated variation of velocity with depth for a granite and garnet-bearing mafic granulite (right) for room temperature conditions and geothermal gradients corresponding to 40 and 90 mW/m^2 heat-flow provinces (left). Curves are dashed where values are extrapolated above experimental conditions.

room-temperature variations of V_p strictly with pressure. These two rock types represent near end-member lithologies for the continental crust and exhibit similar behavior with temperature. Velocities tend to increase with depth for low geothermal gradients. At depths below approximately 20 km, both rocks show velocity reversals. Thus, decreasing velocity with depth should be common in crustal regions where lithology is uniform. In regions characterized by extremely high temperature gradients, partial melting, or the alpha-beta phase transition in quartz will generate low-velocity zones (Gutenberg, 1959; Kern, 1979). Other possibilities include high pore pressure (Christensen, 1986) and intrusions of low-velocity rocks (Mueller, 1977).

It is well known that many rocks exhibit seismic anisotropy. For crustal rocks, this behavior is most notable in metamorphic rocks (Christensen, 1965) including mylonites (Fountain and others, 1984), which exhibit strong fabric elements such as lineation, schistosity, and layering. Anisotropy is reported in Table 8 calculated from $(V_{max}-V_{min})/V_{mean}$ expressed as a percentage. At low pressure, anisotropy can often be related to anisotropy in microcrack orientation (Christensen, 1965; Simmons and others, 1975), whereas at high pressures, preferred orientation of highly anisotropic minerals is primarily responsible for seismic anisotropy. Although most significant in metamorphic rocks, including peridotite tectonites, seismic anisotropy is also detectable in some igneous rocks (Table 8).

The most significant parameter influencing V_p in continental rocks is mineralogy, which is a function of chemical composition and petrologic evolution. Christensen (1965) demonstrated that V_p varies in a systematic manner in coarse-grained igneous rocks (Fig. 6). For instance, there is a consistent increase in V_p between granite and diorite, a change duplicated in the high-grade equivalents, the felsic gneisses (see Table 8). That these changes are related to mineralogy is illustrated by calculations of V_p for combinations of quartz, potassium feldspar, plagioclase, and hornblende following the procedure outlined by Christensen (1966). Instead of presenting simple triangular diagrams, we show a four-component system using four adjacent triangles. These results, calculated for room temperature and midcrustal depths, assuming typical $\partial V_p/\partial P$ values for minerals, are displayed in Figure 7. For the plagioclase composition selected, there is only slight variation of V_p in the quartz-potassium feldspar-plagioclase field. An increase in the amount of the mafic component will produce a significant velocity increase, as will an increase in the anorthite content of plagioclase (Birch, 1961).

Figure 8 shows a similar diagram for a mafic system composed of pyroxene, hornblende, plagioclase (calcic), and garnet. This combination not only describes various gabbroic rocks, but also a variety of metamorphosed mafic rocks, including garnet granulites and amphibolites. The range of V_p for this system is very large, as it is strongly controlled by garnet. This diagram explains the significant increase in V_p associated with the transformation from gabbroic rocks to the higher grade garnet granulites and eclogites. For ultramafic rocks, a triangular diagram presented by Christensen (1966) illustrates the relation between mineralogy and velocity for aggregates of olivine, orthopyroxene, and serpentine (Fig. 9). We emphasize that these diagrams are for isotropic mineral assemblages, but laboratory measurements demonstrate that anisotropy can have greater effects on V_p than significant changes in mineral percentages. For example, average anisotropy appears to be about 7 percent in peridotites. Referring to Figure 9, this corresponds to the total range of velocities between olivine and orthopyroxene.

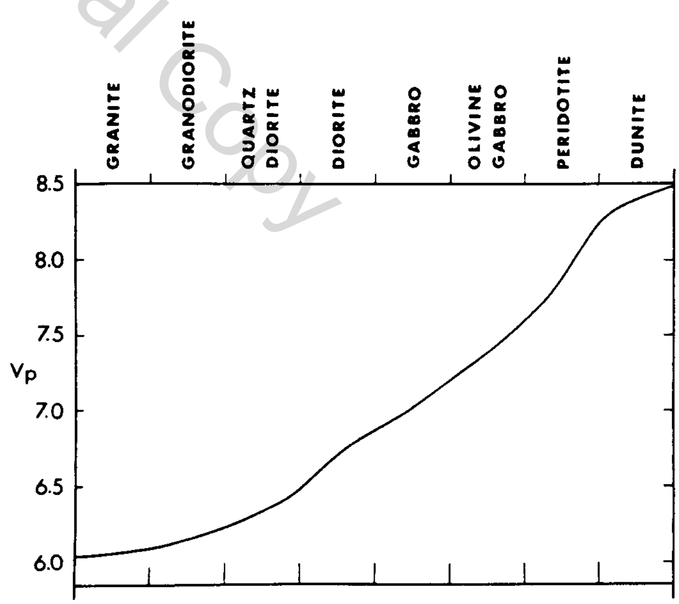


Figure 6. Diagram showing variation of compressional wave velocities at 150 MPa for crustal and upper-mantle rock types (Christensen, 1965).

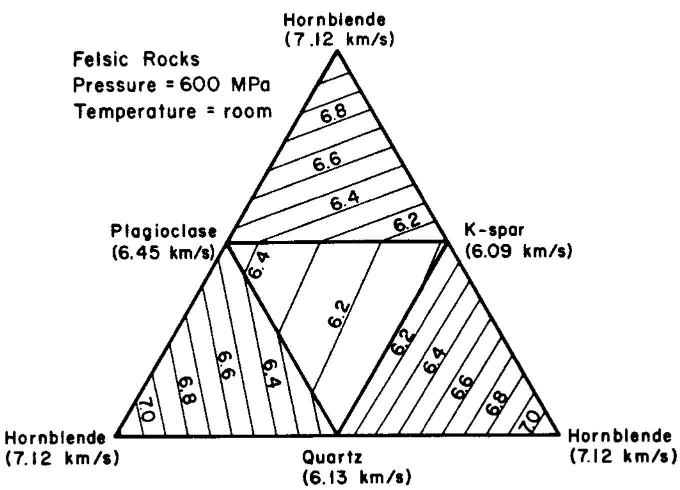


Figure 7. Calculated velocities for the four-component system hornblende-quartz-potassium feldspar-plagioclase (albite). Mineral velocities used in the calculations were obtained from Birch (1961), Alexandrov and Ryzhova (1961, 1962), and McSkimin and others (1965).

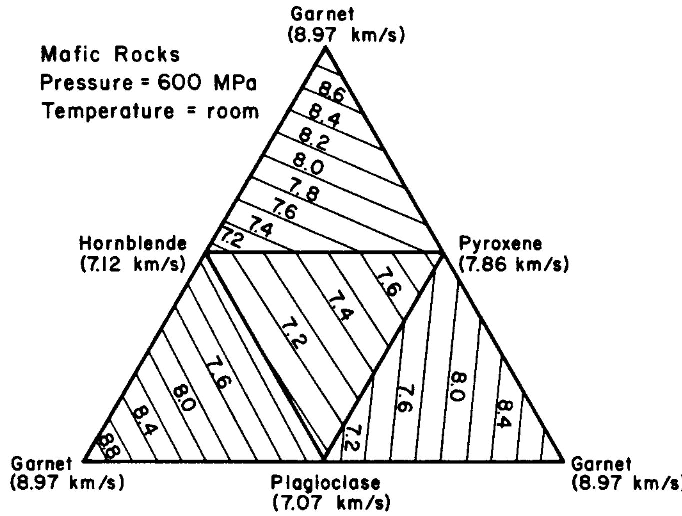


Figure 8. Calculated velocities for the four-component system garnet-pyroxene-hornblende-plagioclase (anorthite). Mineral velocities used in the calculations were obtained from Birch (1961), Alexandrov and Ryzhova (1961), Frisillo and Barsch (1972), and Babuska and others (1978).

Seismic constraints on upper-mantle composition

A major source of information on the petrologic nature of the upper continental mantle is P_n velocity. The availability of data for North America varies significantly from region to region. Western North America has been extensively studied, whereas coverage of the eastern and central regions is limited. Several compilations resulted in contour maps of upper-mantle velocity and crustal thickness for North America or portions of North America (e.g., Pakiser and Steinhart, 1964; Warren and Healy, 1973; Smith, 1978; Blair, 1980; Allenby and Schnetzler, 1983; Braile and others, this volume). The P_n contour map of Blair (1980), given in Figure 10, was constructed by computer contouring seismic data at 349 locations from the compilation of Christensen (1982). A histogram of these upper-mantle velocities is presented in Figure 11.

A comparison of the velocities in Figure 11 with laboratory-measured velocities places major constraints on upper-mantle composition. The major rock types with sufficiently high velocities are dunite, peridotite, and some eclogites (Table 8), as shown in the histogram of Figure 12. Some of the eclogite data in Figure 12 were obtained from xenoliths and, as discussed earlier, V_p may be low because of porosity. In addition, the velocities plotted in Figure 12 do not take into account the effects of accessory minerals (including alteration products) and temperatures, both of which lower P_n velocities significantly. A comparison of Figures 11 and 12 shows that P_n data, such as that in Figure 10, does not discriminate a dominantly eclogitic from a peridotitic upper-mantle model. Thus, additional constraints, such as seismic anisotropy, should be considered.

During the past two decades, numerous papers cited evidence for seismic anisotropy in the upper mantle. Azimuthal anisotropy of P_n velocities is a common property of the oceanic upper mantle (e.g., Raitt and others, 1969; Morris and others, 1969; Shimamura and others, 1983) that, based on studies of ultramafic sections of ophiolities, originates from preferred orientation of highly anisotropic olivine crystals (e.g., Christensen and Salisbury, 1979; Christensen, 1984a). Continental upper-mantle anisotropy has been observed in Europe (Bamford, 1973, 1977;

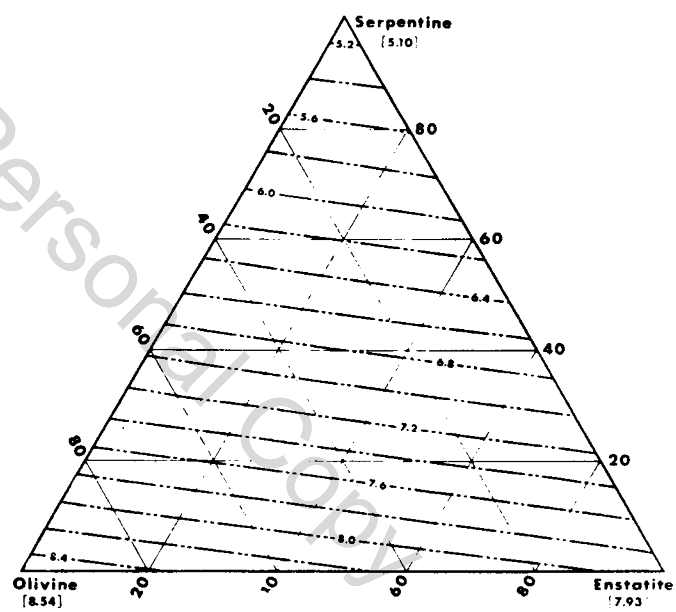


Figure 9. Calculated velocities for the three-component system olivine-enstatite-serpentine at 1 GPa from Christensen (1966).

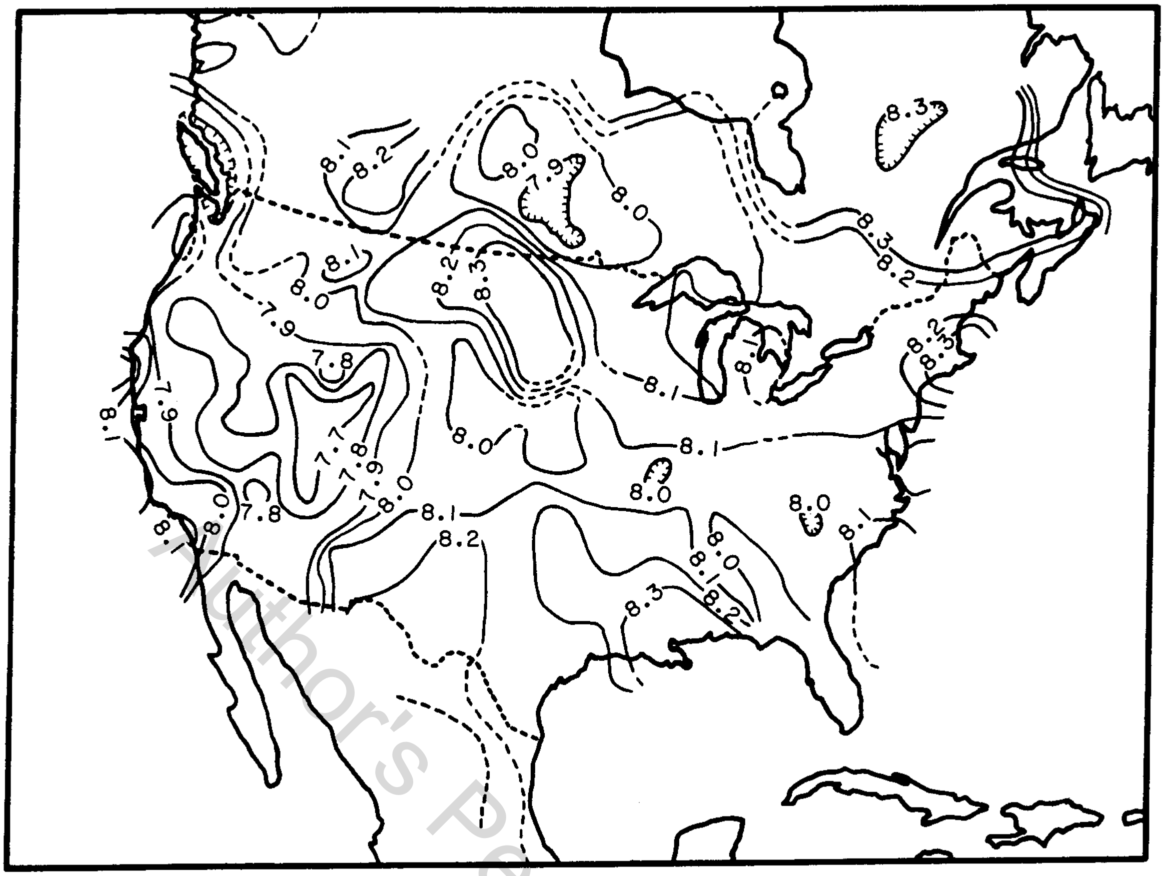


Figure 10. Contour map of P_n velocities for North America from Blair (1980). See Braile and others (this volume) for comparison.

Fuchs, 1975, 1983), the western United States (Bamford and others, 1979; Vetter and Minster, 1981), the USSR (Chesnokov and Nevskig, 1977) and Australia (Leven and others, 1981). Analyses of P_n data have failed to detect anisotropy in the Mojave region of southern California (Vetter and Minster, 1981), the eastern United States (Bamford and others, 1979), and northern Britain (Bamford and others, 1979). It should be emphasized that if upper-mantle anisotropy is transversely isotropic (Christensen and Crosson, 1968) and has a vertical axis of symmetry, P_n anisotropy would not be observed. Significantly, possible upper-mantle peridotite from the Ivrea zone exhibits this type of anisotropy (Fountain, 1976).

Additional observations of the existence of anisotropy to several hundred kilometers depth beneath the continents are based on surface-wave studies (Forsyth, 1975; Mitchell and Yu, 1980; Anderson and Dziewonski, 1982). A global map of 200-sec Rayleigh wave azimuthal variation by Tanimoto and Anderson (1984) was correlated with return flow directions in the asthenosphere.

Anisotropy has been studied in detail in several possible upper-mantle rocks (e.g., Christensen, 1966; Christensen and Ramananantoandro, 1971; Babuska, 1972; Ramananantoandro and Manghnani, 1978). Studies of rock specimens suggest both orthorhombic and axial (transverse isotropy) as possible upper-mantle symmetry. Investigations of anisotropy in ophiolites, believed to represent on-land exposures of oceanic crust and upper mantle, provide strong evidence for an oceanic upper mantle composed of peridotite with olivine *a*-axes producing fast velocities parallel to paleospreading directions (Christensen, 1984a). Our level of understanding of anisotropy beneath the continents is poor because of our lack of knowledge about mantle dynamics and composition. For example, eclogites, which may be an upper continental mantle constituent (Fig. 12), are nearly isotropic or weakly anisotropic. This general low anisotropy of eclogites is illustrated in Figure 13 where anisotropies of possible upper-mantle rocks are plotted against their densities. Examination of this figure shows the strong anisotropy common to most peridotites and dunites and the relatively low anisotropy of many eclogites. Several eclogites, however, do possess significant anisotropy that is likely a consequence of strong preferred orientation of pyroxene. Detailed upper continental mantle seismic studies have the potential to map isotropic versus anisotropic regions and thereby resolve major questions on upper continental mantle composition and dynamics.

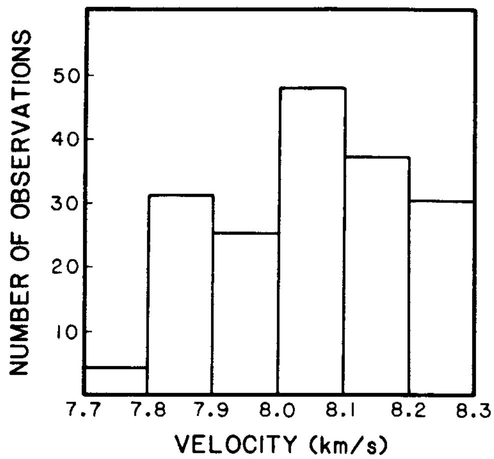


Figure 11. Histogram of P_n velocities for North America.

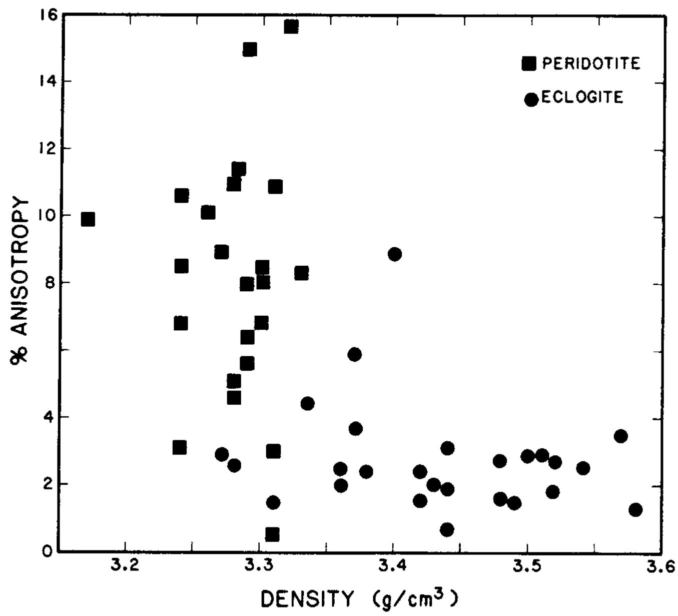


Figure 13. Density versus percent anisotropy for peridotite (squares) and eclogite (circles). Data derived from compilation in Christensen (1982) and Christensen (unpublished data).

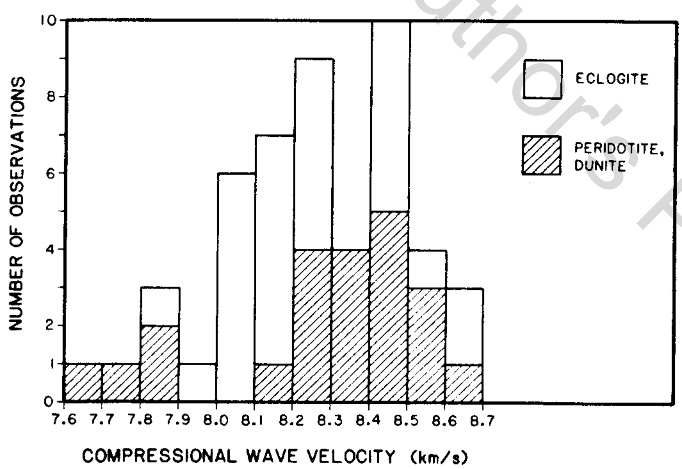


Figure 12. Histogram of compressional wave velocities for eclogite, peridotite, and dunite at 1,000 MPa showing overlap between eclogite and ultramafic rock velocities. Data derived from compilation in Christensen (1982).

Seismic constraints on lower crustal composition

Since Mohorovičić used earthquake arrival-time data to identify the crust-mantle boundary, a variety of seismological methods have been used to investigate the deeper portions of the continental crust. Seismic refraction and reflection profiling are widely used to examine the crust below 10 km depth, and presently provide a more voluminous data set and data of higher resolution than the less widely used methods of magnetic and electrical surveys. In this section we review the seismic evidence bearing on the petrology of middle to lower crustal levels.

Interest in the lower portions of the continental crust has increased lately because of recognition that the deep crust generates significant reflections. One of the earliest examples of deep crustal reflections came from the first Consortium for Continental Reflection Profiling (COCORP) survey in Hardeman County, Texas (Oliver and others, 1976), and is exemplified by recent data from the Basin and Range region (Fig. 14). Since the first COCORP results, reflections from the deep crust have been reported from surveys in Great Britain (e.g., Matthews and Cheadle, 1985), France (e.g., Bois and others, 1985), Germany (Meissner and Lueschen, 1983), Switzerland (Finckh and others, 1985), Australia (Mathur, 1983), and many North American sites (e.g., Oliver and others, 1983). Many seismic reflection workers note that reflective lower crust tends to occur under young extensional terrains such as the Basin and Range (Allmendinger and others, 1983), and the North Sea and the continental shelf of Great Britain (Matthews and Cheadle, 1985), suggesting that deep crustal reflectivity may result from processes associated with crustal extension. Deep crustal reflection events, however, are observed in older terrains with more enigmatic tectonic histories, such as the Adirondacks (Klemperer and others, 1983) and Hardeman County (Oliver and others, 1976). We point out that the cumulative line length over extensional terrains collected by British Institutes Reflection Profiling Syndicate (BIRPS) and COCORP clearly exceeds the length over stable shield areas, that may result in a somewhat biased view of deep crustal reflectivity.

Where reflections are reported, they are generally horizontal to subhorizontal events that exhibit complex, multi-cyclic waveforms with relatively high amplitudes. These characteristics suggest that the deep crust in some regions must be composed of layers of differing acoustic impedances with thicknesses appropriate to cause constructive interference (Fuchs, 1969; Meissner,

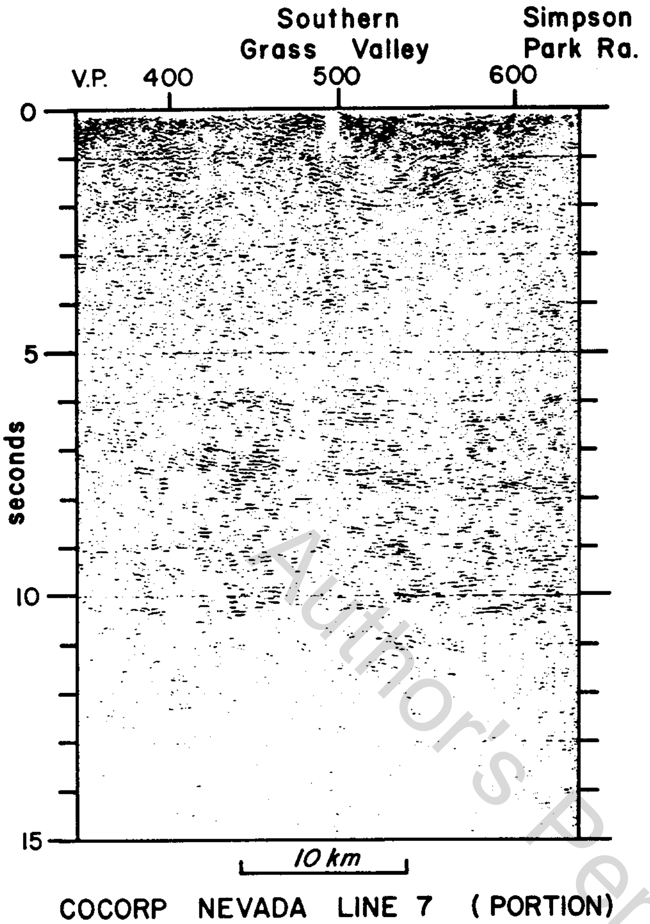


Figure 14. Unmigrated stacked seismic section of the eastern part of COCORP Nevada line 7 (from Potter and others, 1987) from the Basin and Range province showing reflective lower crust between 6- and 10-sec two-way traveltime. Right side of profile is west and left side is east.

1973; Clowes and Kanasewich, 1970; Hurich and others, 1985). This implies that layering on a scale of less than 150 m thick must be prevalent in some regions. Several mechanisms may be responsible for producing this apparent acoustic layering, including variations in petrology, pore pressure, percentages of partial melt, and the degree of anisotropy (e.g., see discussions by Fountain and others, 1984; Matthews and Cheadle, 1985). The relative importance of these mechanisms is poorly understood. Also, to date, there are few assessments of the magnitude of the necessary reflection coefficients (Sandmeier and Wenzel, 1986); thus we have poor constraint on the velocity and density contrasts within the heterogeneous lower crust. Laboratory and modeling studies on Ivrea and Kapuskasing rocks (Hale and Thompson, 1982; Fountain, 1986; Fountain and others, 1987) indicate that reflection coefficients in compositionally layered deep crustal sequences can be very large (absolute value, $\leqslant 0.15$).

In many regions, the middle portion of the crust is reported to be transparent because reflections are either absent or not abundant. This generality was pointed out in BIRPS data (e.g., Smythe and others, 1982) and has been reported elsewhere. Reflections have been observed at this level by groups using shooting or recording parameters different from those used by COCORP (Hurich and others, 1985; Coruh and others, 1987), suggesting that, in some cases, absence of midcrustal reflections may be related to recording and shooting methods. It is important to note that absence of reflections does not imply that the middle crust is lithologically or structurally homogeneous. Inspection of Table 8 shows that there is little significant variation in velocity and density in quartzo-feldspathic lithologies, although there are clear petrologic distinctions. A complex middle crust composed of plutons and various granitic to tonalitic gneisses would not generate large reflections when compared to deep crustal events generated by layers of granulite-facies mafic and tonalitic gneisses (see Table 8).

Whereas reflection methods provide constraints on certain scales of deep crustal geometry, refraction methods and other approaches to velocity inversion can provide velocity data that can, perhaps, better constrain crustal composition. Although numerous refraction surveys have been conducted in the United States, those run in the past 5 to 10 yr are of particular note because of the use of modern recording methods, relatively small station spacing, and new interpretation techniques. Most of these newer surveys were conducted over interesting tectonic features such as rift zones, basins, and recently active magmatic systems. Few have been run over stable shield or platform areas. Thus, the view of the deep crust obtained from these more recent data may be somewhat biased. With that caution in mind, we present some velocity models from these surveys with the rock velocity curves from Figure 5 superimposed for reference.

The Basin and Range and geographically associated Snake River Plain–Yellowstone Plateau (SRP–YP) has the densest refraction coverage of any region in the United States. Figure 15 presents data from the Wasatch Front area (Braile and others, 1974) and the SRP–YP region (Braile and others, 1982; Sparlin and others, 1982; Smith and others, 1982). The deeper crust under the SRP–YP is characterized by a high-velocity (6.8 km/sec) layer that, for a high geothermal gradient, approaches the V_p curve for the mafic garnet granulite gneiss. A variety of mafic lithologies, anorthosites, or even high-grade metapelites could easily match this V_p at appropriate pressure and temperature conditions. Midcrustal levels under these terrains are about 6.5 km/sec, perhaps reflecting a more felsic composition.

Data from the western margin of the United States are shown in Figure 16, which displays velocity profiles for the Coast Range of California (Walter and Mooney, 1982) and the Oregon Cascades (OC) (Leaver and others, 1984). The Cascade model exhibits a thick midcrustal zone characterized by V_p of about 6.5 km/sec, perhaps indicative of felsic rocks. This zone overlies a high V_p (>7 km/sec) lower crust that closely matches the V_p for garnet granulite for the geothermal gradient corresponding to a 90 mW/m^2 heat-flow province. The Diablo and Gabilan Ranges of California exhibit differing V_p profiles and are characterized by

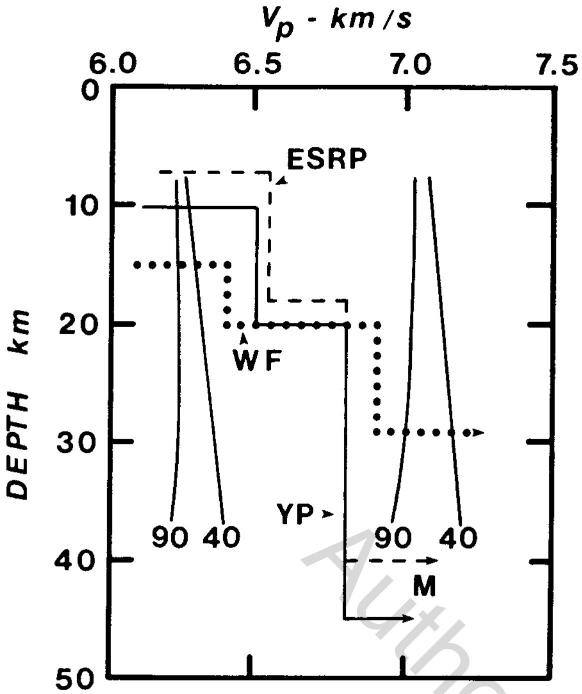


Figure 15. Seismic velocity structure for the eastern Snake River Plain (ERSP), Yellowstone Plateau (YP), and Wasatch Front (WF). M corresponds to Moho. The curves marked 40 and 90 correspond to the variation of velocity with depth for granite (left pair) and mafic garnet granulite (right pair) for heat-flow regimes of 40 and 90 mW/m^2.

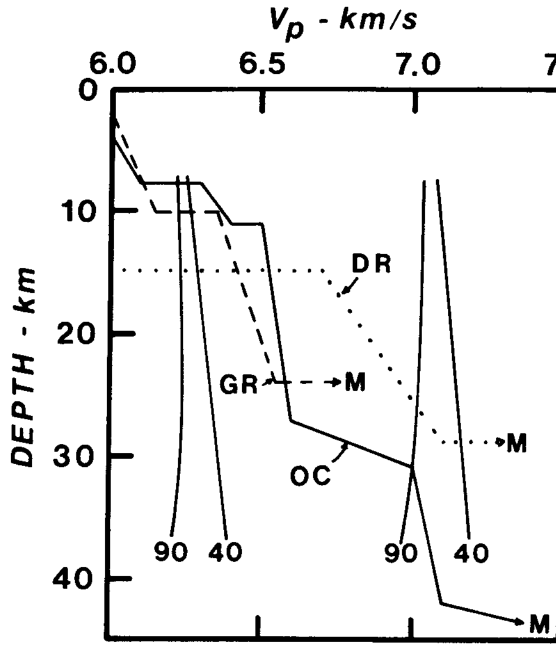


Figure 16. Seismic velocity structure for the Diablo Range (DR) and Gabilan Range (GR) of the California Coast Ranges, and Oregon Cascades (OC) with the same velocity curves for granite and mafic granulite as in Figure 15.

velocity gradients too high to be explained by pressure and temperature changes in a single rock type. The gradients, if real, must reflect significant compositional changes with depth. The Gabilan Range (GR) section is apparently dominated by felsic rocks at depth, whereas felsic midcrustal levels in the Diablo Range (DR) must give way to mafic-dominated lithologies in the lowermost crust. Lin and Wang (1980) and Walter and Mooney (1982) attributed the high velocities under the Diablo Range to a predominantly mafic lower crust, perhaps similar in nature to gabbroic rocks locally found in the dominantly metasedimentary Franciscan complex. Alternative lithologies for the deepest crustal levels might include anorthosite, blueschist, partially serpentinized peridotite, or metapelites.

Recent refraction studies in the midcontinent region are sparse, but two interesting data sets were presented for the Mississippi embayment (Mooney and others, 1983) and the Canadian portion of the Williston basin (Hajnal and others, 1984). The Mississippi embayment was once an active continental rift system (Ervin and McGinnis, 1975). In both sections (Fig. 17), the velocity profile exhibits values between 6.4 and 6.6 km/sec at midcrustal depths, with significant increases to values in excess of 7 km/sec in the lower crust. These high velocities exceed those for garnet granulite, and suggest—assuming isotropic conditions—abundant higher velocity components (e.g., garnet, olivine, pyroxene, hornblende) at this level.

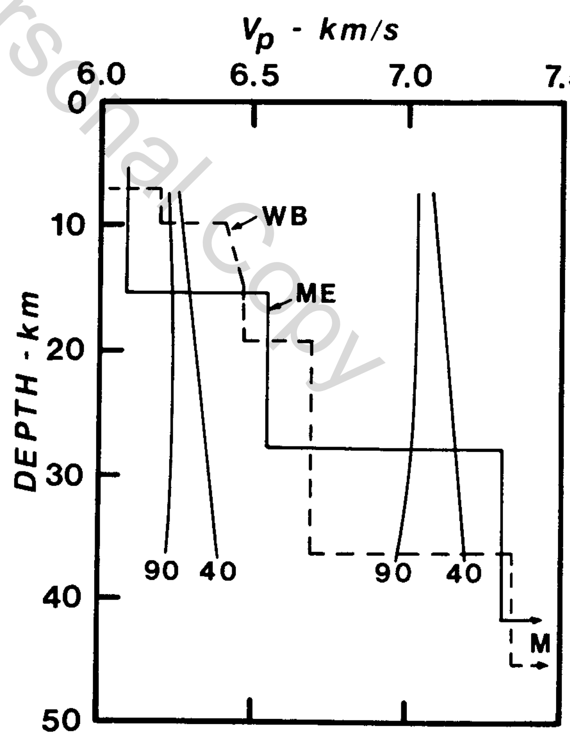


Figure 17. Velocity structure for Mississippi embayment (ME) and Williston basin (WB) with the same velocity curves for granite and mafic granulite as in Figure 15.

Many of these profiles show an intermediate-velocity middle crust and a high-velocity lower crust. These high velocities are generally between 6.8 and 7 km/sec, with a few sections showing higher values. Lower crustal zones dominated by various mafic lithologies, perhaps in upper amphibolite or granulite facies, can explain this range (Table 8). Velocities greater than 7.3 km/sec, however, appear to require the addition of some high-velocity components such as eclogitic or ultramafic rocks if the lower crust is assumed to be isotropic.

Although general compositional information can be gained from interpretation of P-wave velocity data from refraction surveys, the approach has many limitations. First, velocity interpretations based on first arrivals exclusively give the velocity of the refracting horizon and not the entire layer beneath it. Solutions employing later arrivals, amplitudes, and raytracing can give a much more realistic velocity structure. Second, path lengths and wavelengths used in refraction experiments average small-scale heterogeneities so that the velocity determined should not necessarily correspond to a single rock type. Third, direct comparison of V_p from refraction surveys with average V_p for rocks does not take seismic anisotropy into account. Although deep crustal anisotropy has not been reported, laboratory data (Table 8) suggest it may be an important control of wave propagation in the lower crust. Fourth, data in Table 8 indicate that V_p is not a unique function of rock composition, making direct comparisons of field and laboratory data ambiguous. In our previous discussion of the velocity profiles, we elected to correlate V_p with general lithologic categories rather than specific rock types because of this ambiguity. Finally, as stated above, recent refraction studies in the United States have been conducted over areas of tectonic interest, but few have been run over stable shield or platform areas in the midcontinent, resulting in a lack of reference velocity profiles for comparison.

Seismic investigations of the oceanic crust showed that knowledge of shear-wave velocities can provide important constraints on lithologic composition of the crust (Christensen, 1972; Spudich and others, 1978). Shear-wave velocity data for the deeper levels of the continental crust are scarce, generally unreliable, and are derived from a variety of methods including refraction surveys and surface wave dispersion. Available lower crustal shear-wave data for the United States are presented in Table 9. In general, there is considerable variation of V_s in the deep crust, suggesting considerable petrologic differences.

To assess the possibility of constraining deep crustal compositions with shear-wave data, we plot V_p versus V_s for the families of crustal rock types in Table 8 (where shear-wave velocity data are available) at pressures of 600 MPa (Fig. 18). On the same figure we show the V_p and V_s data from Table 9. In some cases there are correspondences between the lower crustal data and the fields of various rock types; in other cases, the field values can be explained by some combination of two or three of the categories.

Shear-wave velocity studies may improve resolution of fine-scale structure of the lower crust. Recently, Owens and others (1984) utilized teleseismic P-waveforms to determine the shear-wave structure beneath the Cumberland Plateau. The resultant models (see Taylor, this volume) show significant lateral heterogeneity in a small geographic region and provide evidence for a thick, laminated crust-mantle transition zone. The P-wave velocity data for the Cumberland Plateau were obtained from a reanalysis of an old U.S. Geological Survey line by Prodehl and others (1984) and are also shown in Taylor (this volume) for comparison with the shear-wave data. In the zones where V_p does not change, we can estimate Poisson's ratio for the southeast profile. For both these zones, Poisson's ratio is close to 0.25, which is low for the fields of crustal rocks presented in Figure 18 but similar to other seismic data. Either the laboratory data are inappropriate for comparison or other factors need to be considered.

It is important to realize that the studies of Owens and others (1984) and Prodehl and others (1984) compare wave energy from different sources traveling in different directions, suggesting that, if the crustal rocks are anisotropic, calculation of Poisson's ratio may be inappropriate. Shear-wave anisotropy in

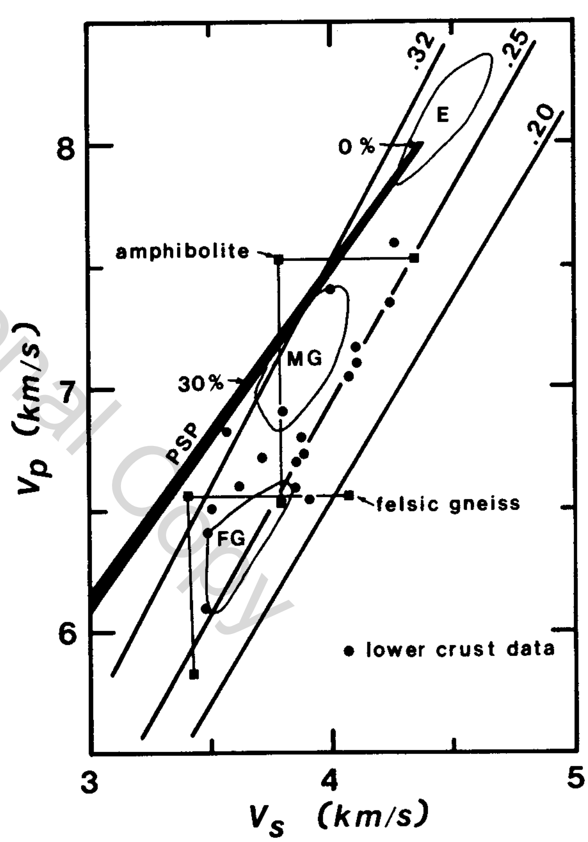


Figure 18. Compressional wave velocity vs. shear-wave velocity showing fields for eclogites (E), mafic gneisses (MG), felsic gneisses (FG), partially serpentinized peridotites with 0 and 30 percent serpentine points marked (PSP = solid area), reported crustal seismic data (circle), and lines of constant Poisson's ratio (0.20, 0.25, and 0.32). Also shown are data for an amphibolite and a felsic gneiss (squares connected by lines) in which shear waves are polarized yielding two shear-wave velocity values for one compressional wave velocity.

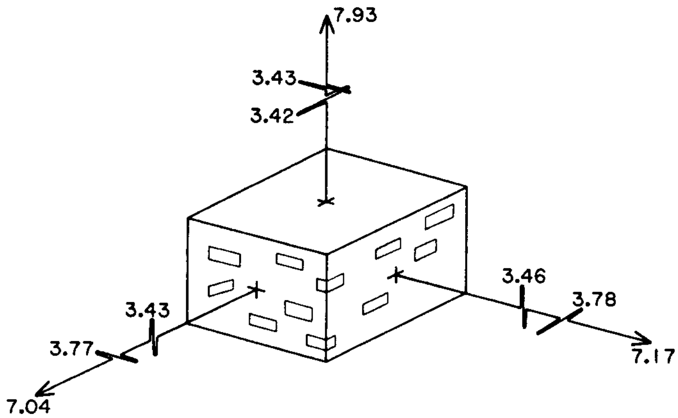


Figure 19. Compressional and shear-wave velocities in different propagation directions for anorthosite (N. I. Christensen, unpublished data). Rectangles schematically represent orientation of plagioclase crystals.

crustal rocks can place severe limitations on the use of V_p-V_s plots such as that presented in Figure 18. In anisotropic rocks, two polarized shear waves with different phase velocities usually propagate in a single direction. Thus, two Poisson's ratios could be calculated for that direction. This effect is illustrated in Figure 19, which shows compressional and shear-wave velocities for several propagation directions for an anorthosite (N. I. Christensen, unpublished data). This behavior introduces considerable complication to the interpretation of V_p-V_s diagrams because the mean V_p and V_s values may not be the appropriate values to use for comparison. This interpretation is also demonstrated in Figure 18, where we show, on a V_p-V_s plot, a single P-wave velocity with its two corresponding polarized shear velocities for several rocks (Christensen, 1971b). Also shown is the slow P-wave velocity normal to foliation and its single corresponding S-wave velocity. These data practically encompass the entire variation of V_p and V_s for the deep crustal data. Furthermore, if shear-wave splitting occurs in large volumes of the deep crust, as it does in some fault zones (e.g., Booth and others, 1985), we may not really know which S-wave was observed, thereby introducing considerable ambiguity in the interpretation of the data. Christensen (1984b) also demonstrated that increasing pore pressure can change V_p/V_s, leading to an increase in Poisson's ratio. This effect raises the possibility that high Poisson's ratios in the lower crust could be related to high pore pressure.

Another issue bearing on the interpretation of velocities in the lowermost crust and upper mantle is the role of the gabbro-eclogite phase transformation, as originally discussed in a series of papers by Ringwood and Green (1964, 1966), Green and Ringwood (1967, 1972), D. Green (1967), T. Green (1967), Cohen and others (1967), and Ito and Kennedy (1968, 1970, 1971). Recently, Furlong and Fountain (1986) estimated the compressional-wave velocities along various geothermal gradients through the gabbro-garnet granulite-eclogite fields, based on new calculations of the phase boundaries (Wood, 1987) and on old estimates of the volumetric abundances of mineral phases through the transformations (Green and Ringwood, 1967). The resulting velocity profiles for an olivine gabbro and quartz tholeiite composition are shown in Figure 20. These curves show no abrupt velocity discontinuities that would correspond to a Moho, but instead show smooth velocity gradients. Of particular importance is the prediction of high velocities (>7 km/sec) at levels roughly equivalent to typical Moho depths. These calculations lend support to the suggestion that high lower crustal velocities may reflect the dominance of garnet-bearing assemblages, as originally suggested by Yoder and Tilley (1962), and in some cases, eclogites in the crust-mantle transition zone. Velocities transitional between crustal and mantle values, such as those predicted in the model, are commonly reported in crust-mantle transition zones in extensional regimes such as the Wasatch Front (Braile and others, 1974).

Constraints on upper crustal composition

Seismological constraints on the composition of the upper portions of the continental crust are primarily derived from interpretation of the P_g phase in refraction studies, upper crustal velocity models from earthquake location algorithms, and seismic reflection data. Traditionally, the upper crust has been considered granitic in composition because of the similarity of P_g velocities to

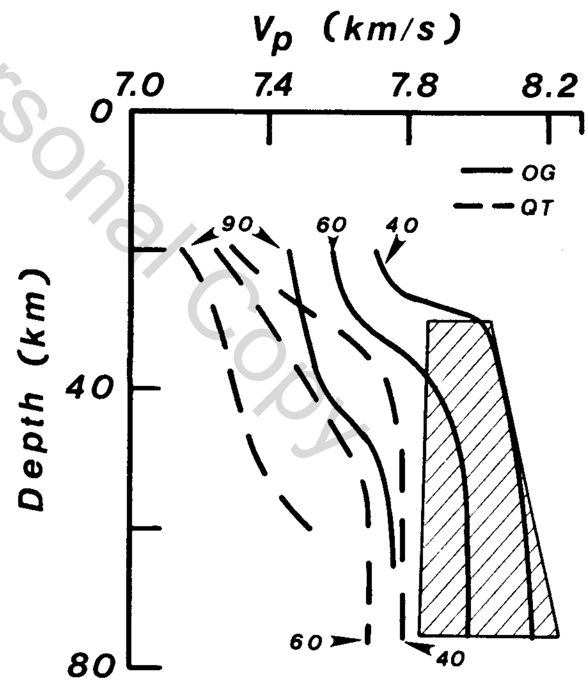


Figure 20. Variation of compressional wave velocity with depth for olivine gabbro (OG) and quartz tholeiite (QT) composition through the gabbro-garnet granulite-eclogite transition based on calculations from Furlong and Fountain (1986). Different curves correspond to calculations for geothermal gradients for 40, 60, 90 (wM/m^2) heat-flow provinces. The shaded area corresponds to velocities for typical ultramafic rocks.

early laboratory compressional wave velocity measurements in granite (Birch, 1958). The surface compositional estimates for shield areas (Table 5), which average all exposed rock types including metavolcanic rocks, suggest, however, an average granodioritic to tonalitic composition. Furthermore, Bott (1982) argued that the average density of the upper crust must be about 2.75 g/cm^3 to account for the significant negative Bouguer anomalies observed over granite and granodiorite plutons. This average is certainly closer to tonalite values than granite values, which is consistent with the conclusion of Woollard (1966) based on density measurements of crystalline rocks. Finally, inspection of the crustal cross sections (Table 7) shows that granites are not dominant, and most likely, that the upper crust is laterally and vertically heterogeneous.

It appears that upper crustal seismic velocities for the frequency range of refraction, reflection, and logging experiments are often lower than expected for the exposed rock types because the upper crust is penetrated by large-scale fractures. Data from the Soviet deep hole in the Kola Peninsula (e.g., Kremenetsky and Ovchinnikov, 1986) suggest that, at least locally, fractures and fluids may be pervasive throughout much of the upper crust, indicating that velocities through this region will tend to be lower than laboratory velocities. Seismic velocities measured in boreholes in crystalline rocks tend to be lower than laboratory-measured values for water-saturated rocks recovered from the hole. Examples of this effect have been reported for a borehole in the Wind River Range (Simmons and Nur, 1968; Smithson and Ebens, 1971), the Matoy well in Oklahoma (Simmons and Nur, 1968), the Stone Canyon well in California (Stierman and Kovach, 1979), and a borehole in the Flin Flon district in Saskatchewan (Hajnal and others, 1983). Cross-hole seismology (Wong and others, 1983) and borehole geophysical tomography (Daily and Ramirez, 1984) identified significant large-scale fractures in borehole environments, an observation that supports the hypothesis that large-scale cracks in the rock mass affect measurements of crustal seismic velocities obtained by field seismic methods. Laboratory velocity measurements represent the maximum velocity of the rock, that is, the value the field measurements would record if the rock mass were free of large-scale fractures (e.g., Hyndman and Drury, 1967; Salisbury and others, 1979; Stierman and Kovach, 1979). Furthermore, rock velocities will be significantly lowered if pore pressure is in excess of hydrostatic pressure (see Christensen, this volume). Thus, estimation of upper crustal composition from velocities requires knowledge of porosity and pore pressure, as has been illustrated by data from the Kola drillhole (see review by Christensen, this volume).

Detailed geometric aspects of the upper crust have been revealed by extensive exploration of the crust with seismic reflection techniques over the past decade. In the United States, these studies revealed important structural detail not evident in seismic refraction data. COCORP and other groups demonstrated that upper crustal faults, fault zones, and faulted terrain geometry can be imaged with reflection techniques. This is well illustrated by data from the Wind River Range (Smithson and others, 1978), the southern Appalachian Mountains (Cook and others, 1979), the Sevier Desert (Allmendinger and others, 1983), southern Oklahoma (Brewer and others, 1983), Kettle dome (Hurich and others, 1985), and the Mojave Desert (Cheadle and others, 1986). Buried basins and associated faults have been identified in Michigan (Brown and others, 1982), Hardeman County in Texas (Brewer and others, 1981), and Kansas (Serpa and others, 1984). The internal seismic structure of fault zones is reviewed by Mooney and Ginzburg (1986). A magma chamber in the upper crust was imaged in the Rio Grande rift (Brown and others, 1980; Brocher, 1981).

There are additional approaches that may yield reliable information about the composition of the upper continental crust. Clearly, surface geology can, in many cases, be reliably extrapolated to depths perhaps as great as 10 km. In some cases such extrapolation can be augmented by shallow borehole information, and future deep drilling will provide a wealth of information on the petrologic nature of the upper continental crust. Because seismic methods can recognize velocity contrasts in the upper crust, high-resolution refraction and reflection techniques can provide key information on the geometry of upper crustal bodies in some cases. In a few situations, exposed cross sections of the crust provide a window through many kilometers of the upper crust. These methods should be used on a case-by-case basis to develop site-specific upper crustal models.

SUMMARY AND CONCLUSIONS

In this chapter, we reviewed several approaches earth scientists have employed to determine the compositions of the crust and upper mantle. Most estimates of bulk chemical composition of the crust center around 60 percent SiO_2, although more silicic and mafic estimates have been proposed. Published estimates of the chemical composition of the lower crust vary from mafic to intermediate, whereas estimates of upper crustal composition, based on averages of surface rocks, tend to be similar (e.g., mean SiO_2 of 65 percent). In contrast, xenolith suites and nearly complete cross sections of the crust, exposed in orogenic belts, demonstrate the lithologic heterogeneity of the crust and upper mantle. Xenolith suites form the basis of several models that exhibit mafic granulites in the lower crust. Upper-mantle models show a complex mixture of various ultramafic and eclogitic rocks. Of significance, most cross sections show the crust is vertically zoned by metamorphic grade, with various composition granulites composing the lower crust.

Constraints offered by seismological methods provide significant information on crustal and upper-mantle composition. Laboratory studies demonstrate that cracks, confining pressure, pore pressure, and temperature, in addition to mineralogy, influence crustal seismic properties. Seismic anisotropy is an important parameter for crustal and upper-mantle rocks. In some cases its effects are greater than compositional changes. Quartzofeldspathic rock types show relatively small variations in compressional wave velocity with composition. In contrast, mafic rocks show a large variation of compressional wave velocity that is primarily related to variable proportions of garnet and plagioclase, or garnet and clinopyroxene. Seismic anisotropy, however, in many rocks can dominate the effect of compositional variations on compressional wave velocity. Constraints offered by seismological methods, combined with laboratory investigations, provide the following significant information on crustal and upper-mantle composition:

1. Upper-mantle velocities vary significantly in North America and correspond to laboratory velocities of dunite, peridotite, and eclogite, or mixtures of these rock types.
2. Seismic anisotropy, when recognized in the continental mantle, provides evidence for a peridotite-dominated upper mantle.
3. In many regions, seismic reflections are observed in the lower crust. Although variations in pore pressure and anisotropy could produce these reflections, we favor an origin due to compositional variation, as is observed in exposures of deep crustal rocks.
4. Midcrustal levels may be seismically transparent because of the dominance of quartzo-feldspathic lithologies.
5. Several recent North American seismic refraction lines, primarily over Phanerozoic tectonic features, suggest that the lower crust is dominated by various mafic rocks (in amphibolite to granulite facies).
6. The predominance of high-velocity components (garnet, olivine, pyroxene) is necessary to explain velocities in the crust-mantle transition zone.
7. Poisson's ratios calculated from compressional and shear-wave velocity data, where available, can be interpreted in terms of specific lithologic composition and variable pore pressure. For both laboratory and field seismic studies, the combined use of shear- and compressional-wave velocity data offers the greatest opportunity in the future to decipher the crustal and upper-mantle composition (e.g., see Christensen and Fountain, 1975).
8. The gabbro-garnet granulite-eclogite transition is likely important in deep, mafic portions of the crust. Garnet granulites can yield velocities intermediate between typical lower crustal values and upper-mantle values.
9. The uppermost crust sampled by P_g phases is likely not granite. Its composition can be approximated by tonalite, but it is lithologically heterogeneous.
10. Upper crustal physical properties, in many regions, are probably strongly controlled by large-scale porosity and fluids. These properties make direct translation of velocity into compositional information difficult.

To this point, our discussion indicates that there are many ambiguities and unknown parameters involved in the interpretation of crustal and upper-mantle composition. We have also discussed many important constraints that available data place on the problem. Despite the mentioned shortcomings of exposed cross sections, these terrains still provide the most accurate view of large portions of the Earth's crust. When the additional constraints of xenolith composition and seismic results are considered, we can formulate some realistic models of the crust and upper mantle in an approach similar to that developed by Smithson and Brown (1977). In Figure 21, we show a crustal and upper-mantle model constructed by interpolating the geologic relations between several exposed cross sections and, by inference, of geologic relations below the sections. The model is intended to be consistent with exposed cross sections, xenoliths, and the seismic nature of the crust and upper mantle. The inset shows which cross sections were used to construct various portions of the model and the variation of metamorphic grade in the model. The greenschist, amphibolite, and granulite boundaries are constrained by their position in the exposed cross sections. We realize that these boundaries may or may not coincide with equilibrium conditions predicted by present geothermal gradients. The granulite-eclogite facies boundary was placed on the basis of calculations from Wood (1987) for low geothermal gradients on the right side of Figure 21 and high gradients on the left side.

There are many characteristics of the model that merit comment. The right side of Figure 21 is dominated by sections derived from Archean and Proterozoic shield areas and can be regarded as examples of ancient crust. The left side of the diagram includes the Ivrea and Fiordland sections, terrains with complex Phanerozoic histories. This portion of the figure presents models for crustal structure and composition in active continental tectonic environments. The model shows significant lateral and vertical changes in lithologic composition. In some regions the crust is dominated by felsic rocks; in others the crust is compositionally zoned, with mafic rocks dominating lower crustal levels. Structural complexity is pervasive throughout, although scale limitations prevent exhibition of the fine-scale layering we expect at many crustal and upper-mantle levels. Metamorphic boundaries vary in depth in the model, and in some cases, high-grade metamorphic assemblages are shown tectonically transported to the surface. The upper mantle in the model shows significant lateral and vertical lithologic heterogeneity, as prescribed by the xenolith models. In some areas the distinction between the lower crust and upper mantle is unclear because mafic rocks are found embedded in peridotite and vice versa.

Despite our attempt to integrate a great deal of data into our model, much of the model is still relatively unconstrained. This will remain the case until seismic and potential field methods are able to provide the enhanced resolution that will allow mapping of small-scale bodies in the crust and upper mantle. Importantly, we need to explore methods to improve our capability to translate field data into compositional information. This will require better understanding of seismic anisotropy, pore pressure, shear-wave velocities, and the effect of rock composition on seismic properties.

In this chapter we have emphasized methods of evaluating the composition of the crust and upper mantle through the combined use of field observations and seismological techniques. An understanding of crustal and upper-mantle composition provides a foundation upon which we can develop insights into the evolu-

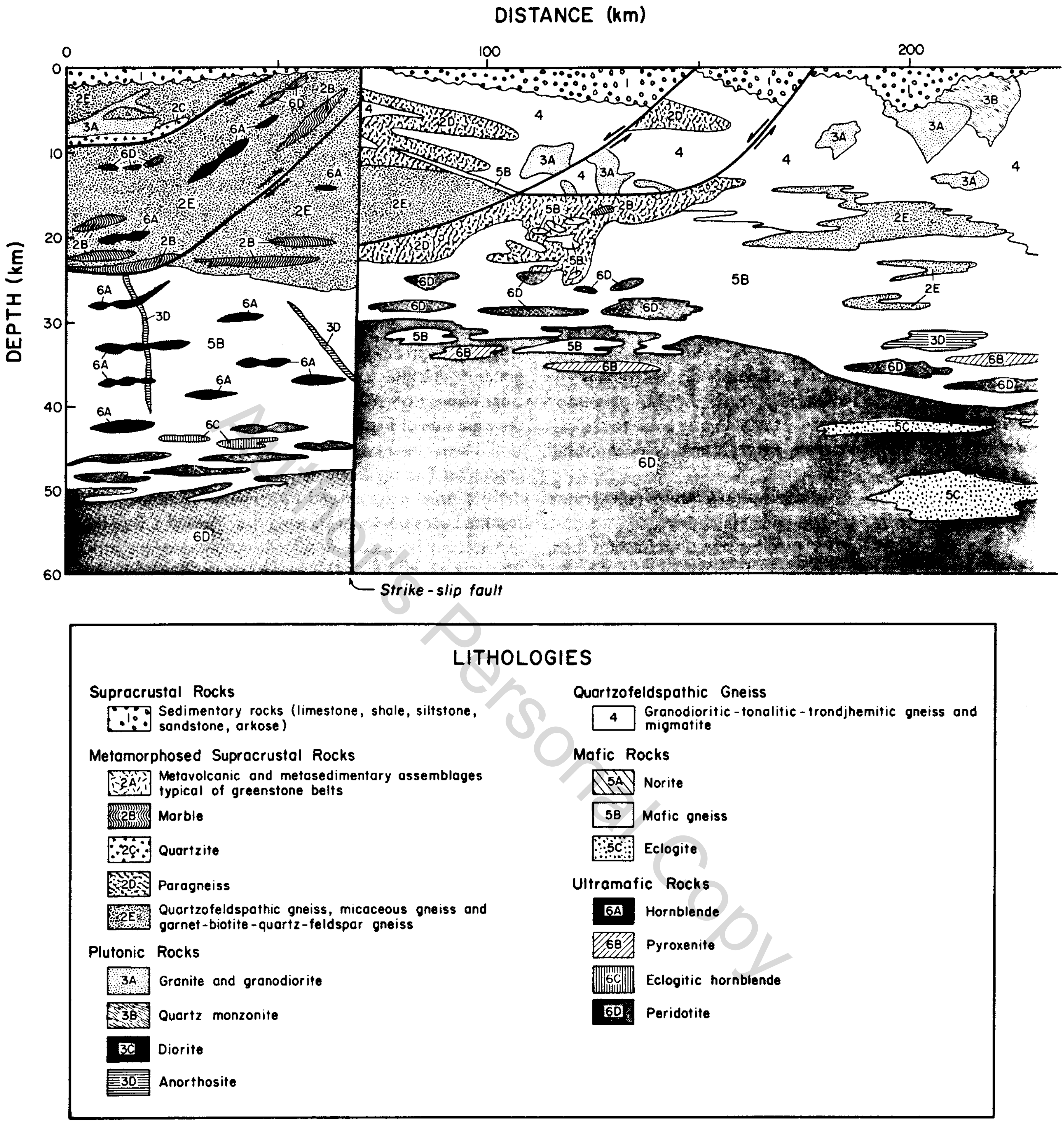
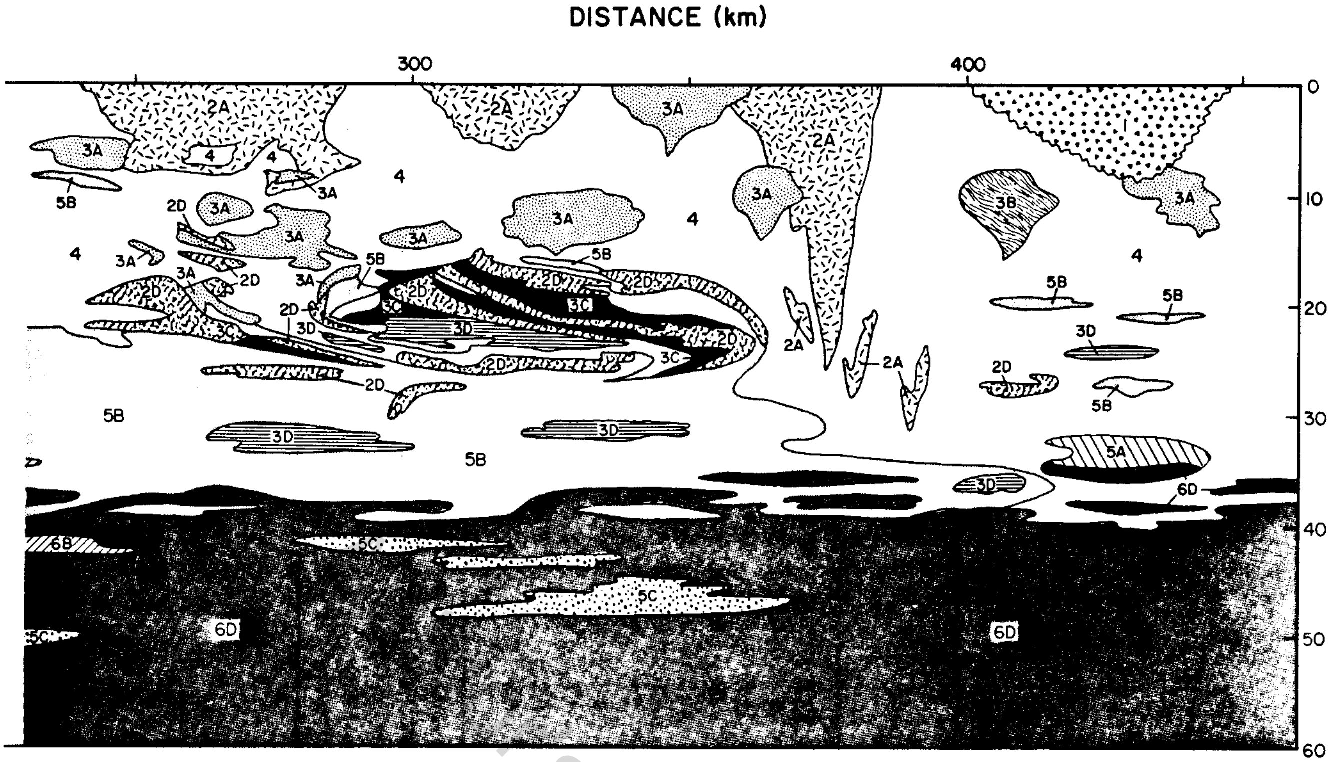
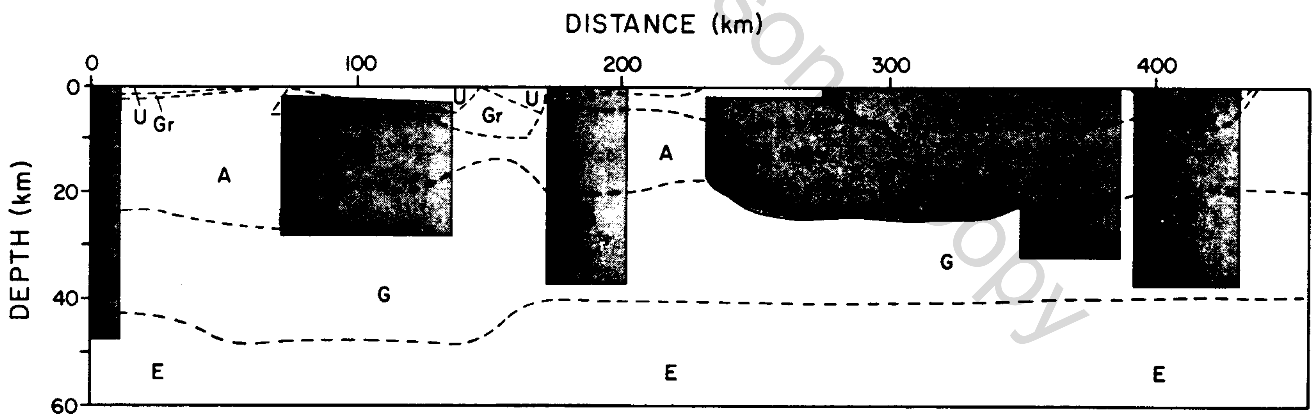


Figure 21. Hypothetical cross section of the continental crust and upper mantle; metamorphic facies and data sources are shown in lower right. Metamorphic facies are unmetamorphosed (U), greenschist (Gr), amphibolite (A), granulite (G), and eclogite (E). Sources for data are (1) Fiordland, New Zealand (Oliver and Coggon, 1979; Oliver, 1980); (2) Ivrea zone, northern Italy (Hunziker and Zingg, 1980; Fountain, 1986); (3) Fraser Range, Australia (Fountain and Salisbury, 1981); (4) Kapuskasing structural zone, Ontario (Percival and Card, 1985); (5) Pikwitonei-Sachigo Provinces, Manitoba (Weber and Scoates, 1978; Manitoba Mineral Resources Division, 1979; Arima and Barrett, 1984); (6) Musgrave Range, Australia (Fountain and Salisbury, 1981). The balance of the model was inferred from data and hypotheses outlined herein.





tion of the continental lithosphere. The lithosphere is a complex mosaic of diverse blocks that have evolved through lateral accretion (e.g., Coney and others, 1980), tectonic underplating (e.g., Yorath and others, 1985), magmatic underplating (e.g., Holland and Lambert, 1975; Furlong and Fountain, 1986), Andean margin magmatism (e.g., Taylor, 1977; Hamilton, 1981), and crustal extension (e.g., Wickham and Oxburgh, 1985). Each of these modifies lithospheric structure and composition. In the future, the recognition of deep continental regions that evolved through these processes will require integration of new detailed seismological laboratory data, deep drilling data, and additional geological observations.

ACKNOWLEDGMENTS

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TABLE 1. ESTIMATES OF BULK CHEMICAL COMPOSITION OF CONTINENTAL CRUST

Major Oxide	Author Reference 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20
SiO_2	59.0	59.1	61.9	59.2	59.8	58.4	63.4	60.3	57.9	60.2	61.9	52.4	62.5	60.6	58.0	63.0	58.0	63.2	63.7	57.3
TiO_2	1.0	1.1	1.1	0.8	1.2	1.1	0.7	1.0	1.2	0.7	0.8	0.7	0.7	0.6	0.8	0.7	0.8	0.6	0.5	0.9
Al_2O_3	15.2	15.3	16.7	15.8	15.5	15.6	15.3	15.6	15.2	15.2	15.6	15.8	15.6	16.3	18.0	15.8	18.0	16.1	15.8	15.9
Fe_2O_3	3.1	3.1	....	3.4	2.1	2.8	2.5	....	2.3	2.5	2.6	2.6	6.1	2.1	....	2.0	....	....	....	....
FeO	3.7	3.8	6.9	3.6	5.1	4.8	3.7	7.2	5.5	3.8	3.9	5.6	....	4.7	7.5	3.4	7.5	4.9	4.7	9.1
MnO	0.1	0.1	....	0.1	0.1	0.2	0.1	....	0.2	0.1	0.1	0.1	0.1	0.1	....	0.1	0.1	0.1	0.1	....
MgO	3.5	3.5	3.4	3.3	4.1	4.3	3.1	3.9	5.3	3.1	3.1	7.7	3.2	4.2	3.5	2.8	3.5	2.8	2.7	5.3
CaO	5.1	5.1	4.2	3.1	6.4	7.2	4.6	5.8	7.1	5.5	5.7	7.7	6.0	5.3	7.5	4.6	7.5	4.7	4.5	7.4
Na_2O	3.7	3.8	3.4	2.1	3.1	3.1	3.4	3.2	3.0	3.0	3.1	1.1	3.4	3.0	3.5	4.0	3.5	4.2	4.3	3.1
K_2O	3.1	3.1	3.0	3.9	2.4	2.2	3.0	2.5	2.1	2.8	2.9	0.2	2.3	1.9	1.5	2.7	1.5	2.1	2.0	1.1
P_2O_5	0.3	0.3	....	0.2	0.2	0.3	0.2	0.2	0.3	0.2	0.3	0.1	0.2	....	....	....	....	0.2	0.2	....
H_2O	1.7	1.6	:.:	3.7	....	....	....	....	....	2.8	....	....	....	....	....	....	....	....	....	....

1, Clarke (1924)
2, Clarke and Washington (1924)
3, Goldschmidt (1933)
4, Goldschmidt (1954)
5, Poldervaart (1955)–shield
6, Poldervaart (1955)–fold belt
7, Vinogradov (1962)
8, Taylor (1964)
9, Pakiser and Robinson (1966)
10, Ronov and Yaroshevsky (1967)
11, Ronov and Yaroshevsky (1969)
12, McGetchin and Silver (1972)
13, Holland and Lambert (1972)
14, Galdin (1974)
15, Taylor (1977, 1979)
16, Smithson (1978)
17, Taylor and McLennan (1981)
18, Weaver and Tarney (1984a)
19, Weaver and Tarney (1984b)
20, Taylor and McLennan (1985)

TABLE 2. ESTIMATES OF BULK CHEMICAL COMPOSITION OF LOWER CONTINENTAL CRUST

Major Oxide	Author Reference 1	2	3	4	5	6	7	8	9	10
SiO_2	48.8	60.6	61.2	57.4	54.0	56.3	59.0	55.4	56.2	59.2
TiO_2	1.8	0.9	0.5	1.4	0.9	1.1	0.8	1.6	1.4	0.9
Al_2O_3	15.6	15.4	15.2	16.00	19.0	17.1	17.1	17.8	18.2	17.2
Fe_2O_3	2.8	....	2.3	2.1	....	3.8	2.0	....	....	....
FeO	8.2	7.3	5.7	7.3	9.0	4.5	3.4	12.0	10.0	6.1
MnO	0.1	0.2	0.2	0.2	....	0.1	0.1	0.2	0.2	0.1
MgO	1.1	3.9	5.6	5.6	4.1	5.0	2.8	5.6	4.8	3.4
CaO	2.6	5.7	7.5	7.1	9.5	5.5	4.6	2.4	3.9	5.9
Na_2O	3.9	2.8	3.0	2.0	3.4	2.1	4.0	2.0	2.0	4.0
K_2O	3.8	2.2	2.0	0.9	0.6	1.4	2.7	2.9	2.5	2.4
P_2O_5	0.2	....	....	....	....	....	....	....	....	0.3

1, Poldervaart (1955); Pakiser and Robinson (1966)
2, Lambert and Heier (1968)
3, Sheraton and others (1973)
4, Mehnert (1975)
5, Taylor (1977, 1979); Taylor and McLennan (1981)–gneisses
6, Leyreloup and others (1977)
7, Smithson (1978)
8, Maccarrone and others (1983–without dioritic and tonalitic gneisses
9, Maccarrone and others (1983)–with dioritic and tonalitic
10, Weaver and Tarney (1984a)

TABLE 3. ESTIMATES OF BULK COMPOSITION OF UPPER CONTINENTAL CRUST

Reference	Major Oxide % SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	H_2O
Galdin (1974)	62.6	0.6	16.5	1.9	3.8	0.1	2.9	4.3	3.3	2.2	....	....
Taylor (1977, 1979)	66.0	0.6	16.0	....	4.5	0.1	2.3	3.5	3.8	3.3	0.2	....
Taylor and McLennan (1985)	66.0	0.5	15.2	....	4.5	....	2.2	4.2	3.9	3.4	....	....

TABLE 4. MINERALOGIC COMPOSITION OF SHIELD AREAS

Mineral	Shaw and others (1967)	Wedehpol (1969)	Nesbitt and Young (1984)*	Nesbitt and Young (1984)†
Plagioclase	39.3	41.0	30.9	34.9
K-Feldspar	8.6	21.0	12.9	11.3
Quartz	24.4	21.0	23.2	20.3
Glass	0.0	0.0	0.0	12.5
Amphibole	0.0	6.0	2.1	1.8
Biotite	11.2	4.0	8.7	7.6
Muscovite	7.6	0.0	5.0	4.4
Chlorite	3.3	0.0	2.2	1.9
Pyroxene	0.0	4.0	1.4	1.2
Olivine	0.0	0.6	0.2	0.2
Oxides	1.4	2.0	1.6	1.4
Other	4.7	0.5	3.0	2.6

*Average mineralogic composition of upper crust.
†Average composition of exposed crust.

TABLE 5. PERCENTAGES OF ROCKS IN SHIELD AREAS*

Lithology	Superior–Slave–Wyoming	Churchill	Central	Grenville
Sedimentary rock	5	18	2	20
Felsic volcanics	0.1	0.5	20	4
Basic volcanics	12	6	3	3
Granitic rock†	76	70	70	66
Diorite and quartz diorite	2	1	0.01	0.01
Peridotite	0.1	Trace	Trace	Trace
Other	4	4	5	5

*From Engel (1963).
†Includes quartz monzonite, granodiorite, quartz porphyry, and gneisses pervasively veined by granite.

TABLE 6. LITHOLOGY OF XENOLITHS FROM NORTH AMERICAN LOCALITIES*

Location	Rock Types	References
1. Rhode Island	Spinel lherzolite (9–25/1,000–1,200), spinel harzburgite, werhlite	Leavy and Hermes (1979)
2. Ithaca, New York	Mafic syenite, garnet clinopyroxenite	Schulze and others (1978); Kay and others (1983)
3. Elliot County, Kentucky	Garnet peridotite	Schulze (1984)
4. Riley County, Kansas	Granite, schist, diorite, gabbro, norite, amphibolite, granulite facies metagabbro, garnet-sillimanite-sapphrine–bearing granulite, pyroxenite, eclogite, plagioclase eclogite	Brookins and Meyer (1974); Meyer and Brookins (1976)
5. State Line, Colorado-Wyoming, and Iron Mountain, Wyoming	Charnockite(?), anorthosite, gabbronorite, hypersthene monzogabbro, two-pyroxene granulite, two-pyroxene garnet granulite, cpx-garnet granulite, hypersthene granulite, garnet-kyanite granulite, augite granulite, kyanite eclogite, eclogite (700-1,300), dunite, garnet-spinel harzburgite (8–20/590–775), spinel lherzolite (900–950), garnet-spinel lherzolite (15–25/650–750), garnet-spinel-olivine websterite d(15–25/650–750), garnet-spinel clinopyroxenite (15–25/650–750)	Eggler and McCallum (1974); McCallum and Maberek (1976); Ater and others (1984); Bradley and McCallum (1984); Kirkley and others (1984)
6. Leucite Hills, Wyoming	Anorthosite, gabbroic granulite (<10/875) granite, quartz diorite, granite gneiss, diorite gneiss, quartz diorite gneiss, schist, dunite	Kay and others (1978); Sperr (1985)
7. Big Belt Mountains, Montana	Cpx-opx-plag-spinel granulite, pyroxenite, harzburgite, wehrlite,lherzolite	Eggler and McCallum (1974)
8. Central Montana	Schist, gneiss, amphibolite, granulite, mafic granulite, mafic amphibolite, spinel peridotite, spinel pyroxenite, dunite, garnet lherzolite (50–50/1,220–1,350), garnet lherzolite (23-42/830-990), garnet pyroxenite (50–60/ 1,220–1,350), garnet harzburgite (50–60/ 1,230–1,350), garnet dunite (50–60/ 1,220–1,350)	Hearn and Boyd (1975); Hearn and McGee (1984)
9. Snake River Plain, Idaho	Rhyolite, pumice, welded tuff, sedimentary rock, biotite-garnet gneiss (730), charnockite, opdalite (hypersthene granodiorite), enderbite, hypersthene diorite, anorthosite, norite (4–8/700–800 for metamorphic rocks)	Leeman (1979); Matty (1984); Leeman and others (1985)
10. Rio Grande Rift (New Mexico)		
a. Black Range	Metagabbro, spinel metagabbro, olivine–spinel metagabbro (11–14/1,100–1,200), spinel lherzolite, spinel clinopyroxenite, clinopyroxenite	Fodor (1978)
b. Kilbourne Hole	Charnockite, anorthosite, sillimanite-bearing garnet granulite, 2-pyroxene granulite (6–10/800–1,000), granet granulite (≤5.4/750–1,000), garnet orthopyroxenite, spinel lherzolite (14-22/900-1100)	Padovani and Carter (1977); Reid and Woods (1978)
c. Engle Basin	Charnockite (7–14/899), pyroxene-plagioclase granulite (5–11/883–934), pyroxenite (8–17/965-1,011), lherzolite (8–20/1,062–1,211)	Warren and others (1979)
d. Elephant Butte	Two-pyroxene granulite (6–13/900—980), clinopyroxenite, spinel lherzolite (14–20/935–1,030)	Baldridge (1979)
e. Abiquiu	Plagioclase-bearing pyroxenite, orthopyroxenite, websterite (877–907)	Baldridge (1979)
f. Cieneguilla	Granulite, harzburgite (943–992)	Baldridge (1979)

TABLE 6. LITHOLOGY OF XENOLITHS FROM NORTH AMERICAN LOCALITIES* (continued)

Location	Rock Types	References
11. Central Colorado Plateau		
a. Navajo Field	Garnetiferous granitic rocks, sillimanite-garnet-biotite gneiss, amphibolite schist, felsic granulite, mafic garnet granulite (>6/555-635), eclogite, clinopyroxenite, websterite, dunite, harzburgite, garnet lherzolite (43/93–1,230)	Ehrenberg (1979); Ehrenberg and Griffin (1979)
b. Moses Rock, Mule's Ear, and Garnet Ridge	Rhyolite porphyry, granite, basalt, chlorite schist, serpentine schist, quartz monzonite, granite gneiss, metabasalt, diorite mafic amphibolite (475–750), garnet amphibolite, garnet-sillimanite schist (750–850) felsic granulite, intermediate granulite, granulite metagabbro gneiss (500–800), eclogite, plagioclase eclogite, garnet pyroxenite (15.5/710), jadeite-clinopyroxenite, spinel websterite, spinel lherzolite	Watson and Morton (1969) McGetchin and Silver (1972); Padovani and others (1982)
c. Buell Park and Green Knobs	Granite, quartz monzonite, granodiorite, felsic volcanic rocks, sandstone, low-grade meta-felsite, biotite schist, amphibolite (530), meta-diabase, two pyroxene granulite (5–8/700–800), amphibole-bearing two-pyroxene granulite, garnet granulite (6–9/640–750), lherzolite-websterite (700–1,100), opx-rich websterite, ultramylonitic and mylonitic peridotite	Smith and Levy (1976); Smith (1979); O'Brien (1983)
12. Colorado Plateau Margin (Arizona)		
a. Chino Valley	Quartzite, schist, two-pyroxene feldspathic granulite, garnet-bearing amphibolite, eclogite (550-700), pyroxenite, garnet pyroxenite, (550–700), garnet websterite (10–20/700–1,000)	Arculus and Smith (1979); Schulze and Helmstaedt (1979; Aoki (1981)
b. Carefree	Plagioclase-bearing amphibolite (assume P = 10/650), garnet-rich amphibolite, garnet-cpx-plagioclase granulite (<8), eclogite (assume P = 10/900), garnet clinopyroxenite	Esperanca and Holloway (1984)
c. San Francisco Volcanic Field	Charnockite, norite, anorthosite, olivine gabbro, two-pyroxene gabbro, two-pyroxene granulite, hypersthene gabbro, pyroxene granulite, plagioclase pyroxenite, wehrlite, olivine clinopyroxenite, clinopyroxenite, websterite, olivine websterite, spinel pyroxenite, dunite	Cummings (1972; Stoesser (1973, 1974)
d. Geronimo volcanic field	Two-pyroxene granulite, harzburgite, websterite, wehrlite, clinopyroxenite, amphibole peridotite, spinel lherzolite	Kempton and others (1984)
13. California		
a. Dish Hill	Granite, two-pyroxene-plagioclase granulite, hornblende clinopyroxenite, clinopyroxenite, spinel lherzolite, spinel wehrlite, garnet clinopyroxenite (15-20/1,100)	Shervais and others (1973)
b. Central Sierra Nevada	Pyroxenite (13), harzburgite, lherzolite, mafic granulites, gabbro	Dodge and others (1986)
14. British Columbia		
a. Kettle River, Lassie Lake, Lightening Peak	Metamorphic lherzolite, dunite, wehrlite (all 9-23/950-1,300)	Ross (1983)
b. Big Timothy Mountain, Jacques Lake	Lherzolite, harzburgite, websterite, dunite, wehrlite (all 10-27/900-1,200), lherzolite tectonite (1,085)	Ross (1983); Littlejohn and Greenwood (1974)

TABLE 6. LITHOLOGY OF XENOLITHS FROM NORTH AMERICAN LOCALITIES* (continued)

Location	Rock Types	References
14. British Columbia (continued)		
c. Castle Rock	Lherzolite and websterite (10-24/1,000-1,300), lherzolite (>1,600)	Ross (1983); Littlejohn and Greenwood (1974)
d. Summit Lake	Wehrlite, spinel lherzolite, clinopyroxenite, olivine websterite, websterite, dunite (18-20/1,080-1,100 or 9-20/950-1,250)	Ross (1983); Brearley and others (1984)
e. Haggens Point, Boss Mountain and Nicola Lake	Lherzolite tectonite (1,085), cumulate lherzolite (840)	Littlejohn and Greenwood (1974)
15. Yukon River and Selkirk Cone, Yukon	Lherzolite (850-1,160), lherzolite, websterite, dunite, wehrlite (all 12-28/950-1,200)	Sinclair and others (1978); Ross (1983)
16. Aleutian arc	Gabbro, diorite, ultramafic cumulate, dunite	Kay and Kay (1981); Kay and others (1986)

TABLE 7. LITHOLOGIES IN EXPOSED CROSS SECTIONS OF THE CONTINENTAL CRUST

Exposed Cross Section	Greenschist Facies and Unmetamorphosed Rocks	Amphibolite Facies	Granulite Facies
Ivera-Verbano and Strona-Ceneri, Italy	Limestone, volcanic rocks, phyllite, orthogneiss, schists	Granite, schist, paragneiss, orthogneiss, amphibolite, marble	Meta-pelite, marble, granetiferous mafic gneiss, mafic gneiss, peridotite, pyroxenite, hornblendite, amphibolite
Sachigo-Pikwitonei, Manitoba, Canada	Metabasalt, metarhyolite, metagraywacke, tonalite, granite, metaconglomerate	Metabasalt, metarhyolite, metagraywacke, tonalite, granite, tonalitic gneiss, granodioritic gneiss	Tonalite gneiss, metagranodiorite, silicate-oxide iron formation, mafic gneiss, ultramafic rocks
Michipicoten-Wawa-Kapuskasing, Ontario, Canada	Metabasalt, metarhyolite, meta-andesite, metagraywacke, granite, tonalite metaconglomerate, iron formation, chert	Granite, granodiorite, tonalite, tonalite-granodiorite gneiss, diorite-monzonite, metavolcanic, rocks, metagraywacke, amphibolite	Anorthosite, tonalite-granodiorite gneiss, mafic gneiss, paragneiss
Fraser Range, Australia	....	Granite, adamellite gneiss, migmatite, quartzo-feldspathic gneiss	Mafic gneiss, metasedimentary gneiss, felsic gneiss, anorthosite, olivine gabbro, norite
Musgrave Range, Australia	Arkose, sandstone	Granite, adamellite, granite gneiss, quartzo-feldspathic gneiss	Maficultramafic plutons, anorthosite, granodioritic gneiss, mafic gneiss, quartzite, calcareous rocks, adamellite
Kasila Series, Western Africa	....	Quartzo-feldspathic gneiss, calc-silicate gneiss, metabasic gneiss	Mafic gneiss, layered gabbro and anorthosite complexes, quartz-magnetite gneiss, quartz-diopside gneiss, sillimanite-bearing gneiss, granite gneiss
Calabria, Italy	Limestone, phyllite, gneiss	Diorite gneiss, tonalite, granite	Metapelite, mafic gneiss, metaquartz monzo-gabbronorite
Fiordland, New Zealand	Granite, granodiorite, diorite, gabbro, sedimentary rocks	Granite, granodiorite, quartzo-feldspathic and micaceous gneiss, schist, metabasite, marble, calc-silicate	Metagabbroic diorite, feldspathic gneiss, ultramafic rocks, anorthosite veins

TABLE 8. COMPRESSIONAL WAVE VELOCITIES AND DENSITIES OF SELECTED CRUSTAL AND MANTLE ROCKS

Rock	Density (g/cm³)	P(MPa) (km/sec) 10	50	100	200	400	600	1000	Anistropy (%)	Reference*
Granite										
Westerly, Rhode Island	2.619	4.1	5.63	5.84	5.97	6.10	6.16	6.23	3.7	1
Barre, Vermont	2.655	5.1	5.86	6.06	6.15	6.25	6.32	6.39	1.3	1
Barriefield, Ontario†	2.672	5.7	6.21	6.29	6.35	6.42	6.46	6.51	1.5	1
Tonalite										
Val Verde, California	2.763	5.1	....	6.33	6.43	6.49	6.54	6.60	1.1	1
San Luis Rey, California	2.798	5.1	....	6.43	6.52	6.60	6.64	6.71	0.7	1
Diorite										
Dedham, Massachusetts	2.906	5.5	....	6.46	6.53	6.60	6.65	6.70	0.1	1
Gabbro and norite										
Mellen, Wisconsin	2.931	6.8	7.04	7.07	7.09	7.13	7.16	7.21	3.5	1
French Creek, Pennsylvania	3.054	5.8	6.74	6.93	7.02	7.11	7.17	7.23	1.0	1
Transvaal, Africa	2.978	6.6	7.02	7.07	7.11	7.16	7.20	7.28	1.8	1
Anorthosite										
Adirondacks, #1	2.707	5.96	6.40	6.58	6.74	6.83	6.89	6.95	2.3	2
Stillwater Complex, Montana	2.770	6.5	....	6.97	7.01	7.05	7.07	7.10	3.9	1
Bushveld Complex	2.807	5.7	6.92	6.98	7.05	7.13	7.16	7.21	3.1	1
Amphibolite facies granitic gneiss										
Gneiss 1, Connecticut	2.654	4.5	....	5.85	6.06	6.18	6.24	6.33	2.2	3
Gneiss 2, Connecticut	2.643	4.8	....	5.97	6.12	6.22	6.27	6.35	1.9	3
Amphibolite facies granodioritic gneiss										
New Hampshire	2.758	4.4	....	5.95	6.07	6.16	6.21	6.30	4.1	1
Amphibolite facies tonalitic gneiss										
Gneiss 3, Connecticut	2.755	5.1	....	6.15	6.32	6.43	6.49	6.57	0.3	3
Gneiss 4, Connecticut	2.824	4.8	....	6.03	6.25	6.40	6.47	6.58	5.2	3
Pelitic schist										
Iverea Zone, Italy (IV-21)	2.700	....	....	5.65	5.92	6.13	6.26	6.37	9.6	4
Thomaston, Connecticut (Garnet Schist)	2.760	5.2	5.95	6.20	6.32	6.43	6.50	6.59	9.9	3
Litchfield, Connecticut	1.750	....	6.06	6.27	6.39	6.52	6.58	6.68	21.0	3
Quartzite										
Montana	2.647	5.6	....	6.11	6.15	6.22	6.26	6.35	3.1	1
Clarendon Springs, Vermont	2.630	5.5	5.89	6.05	6.12	6.19	6.24	6.30	0.3	3
Greenstone										
Yreka, California	2.910	6.7	6.80	6.84	6.90	6.96	6.99	....	3.3	5
Luray, Virginia	2.930	6.5	6.54	6.58	6.65	6.71	6.75	....	1.0	5
Marin County, California	2.880	5.9	....	6.35	6.46	6.59	6.70	....	2.1	5

TABLE 8. COMPRESSIONAL WAVE VELOCITIES AND DENSITIES OF SELECTED CRUSTAL AND MANTLE ROCKS (continued)

Rock	Density (g/cm³)	P(MPa) (km/sec)							Anistropy (%)	Reference*
		10	50	100	200	400	600	1000		
Amphibolite										
Canyon Mountain, Oregon	2.925	....	6.66	6.70	6.76	6.80	6.83	6.86	8.6	6
#2 Bantam, Connecticut	3.030	5.5	6.30	6.63	6.87	7.04	7.10	7.18	14.1	3
Ivrea Zone, Italy (IV-16)	3.044	....	....	6.06	6.60	7.04	7.21	7.32	6.1	4
Hypersthene granodiorite gneiss (charno-enderbite)										
Tichbourne, Ontario	2.712	....	....	6.06	6.17	6.31	6.41	6.48	0.7	7
Pallavaram, India	2.740	6.15	....	6.24	6.30	6.36	6.40	6.46	1.2	1
Hypersthene tonalite gneiss (enderbite)										
#3, Adirondacks	2.703	5.97	6.25	6.34	6.41	6.48	6.53	6.60	3.1	2
New Jersey Highlands	2.681	....	....	6.36	6.44	6.52	6.57	6.63	2.2	7
#2, Adirondacks	2.702	5.78	6.23	6.34	6.42	6.51	6.57	6.64	2.7	2
Hypersthene quartz monzonite (mangerite)										
#5, Adirondacks	2.739	6.04	6.20	6.28	6.35	6.44	6.48	6.54	0.5	2
#8, Adirondacks	2.826	6.20	6.37	6.43	6.53	6.57	6.60	6.67	1.7	2
Saranac Lake, New York	2.830	....	....	6.40	6.54	6.66	6.71	6.77	1.0	7
Granulite facies metapelitic gneiss										
Ivera Zone, Italy (IV-23)	2.954	....	....	6.60	6.74	6.87	6.96	7.05	7.6	4
Ivrea Zone, Italy (IV-7)	3.104	....	....	6.99	7.12	7.27	7.35	7.43	1.0	4
Hornblende-pyroxene granulite (mafic gneiss)										
Santa Lucia, California	2.899	....	....	6.75	6.83	6.91	6.95	7.02	3.3	7
#14,Adirondacks	3.170	5.84	6.80	6.95	7.01	7.11	7.16	7.25	4.5	2
Ivrea Zone, Italy (IV-15)	3.080	....	....	7.12	7.26	7.38	7.45	7.51	5.0	4
Pyroxene-garnet granulite (mafic gneiss)										
Ivrea Zone, Italy (IV-9)	2.942	....	....	6.31	6.58	6.85	6.99	7.09	1.6	4
Ivrea Zone, Italy (IV-17)	2.910	....	....	6.90	7.08	7.21	7.28	7.35	1.2	4
Ivrea Zone, Italy (IV-20)	3.047	....	....	6.69	7.12	7.38	7.48	7.57	0.4	4
Serpentinite										
Mount Boardman, California	2.513	....	4.93	4.96	5.03	5.15	5.25	5.42	1.2	6
Paskenta, California	2.517	....	4.94	5.00	5.07	5.19	5.31	5.49	3.8	6
Canyon Mountain, Oregon	2.535	....	5.28	5.31	5.35	5.42	5.50	5.62	2.1	6
Pyroxenite										
Stillwater Complex, Montana	3.279	7.42	....	7.62	7.65	7.72	7.75	7.83	1.3	1
Sonoma County, California	3.247	6.8	....	7.73	7.79	7.88	7.93	8.01	4.7	1
Bushveld Complex	3.288	....	7.40	7.49	7.60	7.75	7.85	8.02	4.1	1
Dunite										
Addie, North Carolina	3.304	7.7	....	7.99	8.05	8.14	8.20	8.28	7.2	1
#4 Twin Sisters, Washington	3.30	8.3	8.30	8.33	8.39	8.43	8.46	8.51	6.8	8
#11 Twin Sisters, California	3.30	8.1	8.22	8.28	8.35	8.43	8.48	8.52	8.1	8

TABLE 8. COMPRESSIONAL WAVE VELOCITIES AND DENSITIES OF SELECTED CRUSTAL AND MANTLE ROCKS (continued)

Rock	Density (g/cm^3)	P(MPa) (km/sec) 10	50	100	200	400	600	1000	Anistropy (%)	Reference*
ECLOGITE										
#13 Sunmore, Norway	3.539	7.6	7.91	7.98	8.06	8.13	8.17	8.23	2.4	2
#14 Sunmore, Norway	3.577	7.8	8.05	8.14	8.22	8.30	8.35	8.42	3.4	2
#15 Grytinvaag, Norway	3.585	7.7	7.79	8.05	8.16	8.28	8.35	8.43	1.2	2

*References: 1, Birch (1960); 2, Manghnani and others (1974); 3, Christensen (1965); 4, Fountain (1976); 5, Christensen (1970); 6, Christensen (1978); 7, Christensen and Fountain (1975); 8, Christensen (1971a).
†Altered, but typical of many granites.

TABLE 9. POISSON'S RATIO AND COMPRESSIONAL AND SHEAR WAVE VELOCITY DATA FOR THE MIDDLE AND LOWER CRUST OF NORTH AMERICA

Location	Depth to Top of Horizon (km)	V_p (km/sec)	V_s (km/sec)	σ*	Method†	Reference§
New Madrid	25.2	7.17	4.10	0.26	E	1
SW Oklahoma	18.0	6.72	3.88	0.25	R	2
	26.0	7.05	4.07	0.25	R	
Eastern New Mexico	23.0	6.72	3.88	0.25	R	2
	32.0	7.10	4.10	0.25	R	
	41.0	7.35	4.24	0.25	R	
Colorado Plateau	27.0	6.80	3.87	0.26	SWD	3
Eastern Basin and Range	22.5	6.70	3.71	0.28	SWD	3
Colorado Plateau–Basin	14.7	6.5	3.50	0.30	R	4
and Range Transition	24.7	7.4	4.0	0.29	R	
Wasatch Front	19.0	6.90	3.80	0.28	R	5
	28.0	7.60	4.25	0.27	R	
Basin and Range	20.0	6.60	3.85	0.24	SWD	6
Northern Basin and Range	20.0	6.60	3.61	0.29	SWD	6
High Lava Plain, Oregon	15.0	6.70	3.85	0.25	SWD	6
Eastern Snake River Plain	20.0	6.82	3.56	0.31	SWD	7

*Poisson's ratio
†SWD = surface wave dispersion; R = refraction; E = earthquake source.
§References: 1, Mitchell and Hashim (1977); 2, Mitchell and Landisman (1971); 3, Bucher and Smith (1971); 4, Keller and others (1975); 5, Braile and others (1974); 6, Priestley and Brune (1982); 7, Greensfelder and Kovach (1982).

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