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Measured and calculated elastic wave speeds in partially equilibrated mafic granulite xenoliths: Implications for the properties of an underplated lower continental crust

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Abstract. Ultrasonic compressional wave velocities measured at 1.0 GPa and room temperature are compared with calculated velocities (based on single-crystal data and modal mineralogy) for a suite of mafic granulite xenoliths from the Chudleigh volcanic province, north Queensland, Australia. The xenoliths have nearly constant major element compositions but widely variable modal mineralogy, reflecting recrystallization under variable pressure-temperature conditions at depth in the continental crust (20-45 km). They thus provide an excellent opportunity to investigate velocity variation with depth in a mafic lower crust. Measured *P* wave velocities, corrected for the decompression-induced breakdown of garnet, range from 6.9 to 7.6 km/sec and correlate with derivation depth. These velocities are 5-12% lower than the calculated velocities (7.5-8.0 km/sec), apparently as a result of grain boundary alteration as well as irreversible changes that occurred in the xenoliths during rapid decompression. Calculated *P* wave velocities are similar to those estimated by Furlong and Fountain (1986) and Sobolev and Babeyko (1989) for mafic granulites formed through basaltic underplating of the continental crust. Depending upon in situ temperature, *P* wave velocities in the deepest samples may be interpreted as crustal (e.g., 7.3-7.6 km/sec, if heat flow is high) or mantle (7.7-7.8 km/sec, in areas of low heat flow). The range of velocities in the xenolith suite is larger than predicted for a fully equilibrated underplated basaltic layer, highlighting the importance of kinetic effects in determining the ultimate velocity profile of magmatically underplated crust. Comparison of our results with seismic profiles illustrates that the lower crust rarely reaches such high velocities, suggesting quartz-bearing rocks (country rocks?) are present within magmatically underplated layers of the deep crust.

Introduction

There is mounting evidence from crustal seismic studies that the lowermost continental crust is dominated by mafic rock types [*Drummond and Collins*, 1986; *Holbrook et al.*, 1992]. Such mafic lower crust may arise in several ways: (1) intrusion and crystallization of basaltic or picritic magma at or near the crust-mantle boundary (basaltic underplating), (2) foundering of midcrustal mafic intrusions into the lower crust [*Glazner*, 1994], and (3) partial melting of evolved rock types, leaving behind mafic residue. Studies of lower crustal xenoliths indicate that much of the lower crust formed through igneous intrusion rather than by partial melt removal (see *Rudnick* [1992], and references therein). And, although basalts do not become neutrally buoyant until they reach midcrustal depths (e.g., *Glazner*, 1994), rheological considerations suggest that mafic magmas may indeed pond at the crust-mantle boundary [*Parsons et al.*, 1992].

Furlong and Fountain [1986] and *Sobolev and Babeyko* [1989] modeled the physical characteristics of a lower crust formed by intrusion of basaltic magma compositions. They considered various mineralogies predicted from experimental petrologic data, differing rates of mineral reequilibration, and various geotherms in order to estimate the range of *P* wave velocity and to model the V_p-depth profiles produced by basaltic underplating.

In this paper we report density and ultrasonic *P* wave velocity measurements for a suite of compositionally uniform lower crustal xenoliths which formed as plagioclase-rich cumulates from basaltic magmas as they intruded the lower continental crust. We compare these results with calculated *P* wave velocity for the same rocks and the model velocity-depth profiles of *Furlong and Fountain* [1986] and *Sobolev and Babeyko* [1989] in order to determine the seismic structure of a lower crust of uniform mafic chemical composition. We then compare our result to seismic refraction profiles in order to evaluate the volumetric significance of basaltic underplating.

Samples and Geologic Setting

The suite of lower crustal xenoliths investigated here comes from <1 Ma cinder cones in the Chudleigh volcanic province of north Queensland, Australia (Figure 1). These erupt near the southern extension of the Burdekin fault zone, a steeply westward dipping thrust fault that separates Proterozoic

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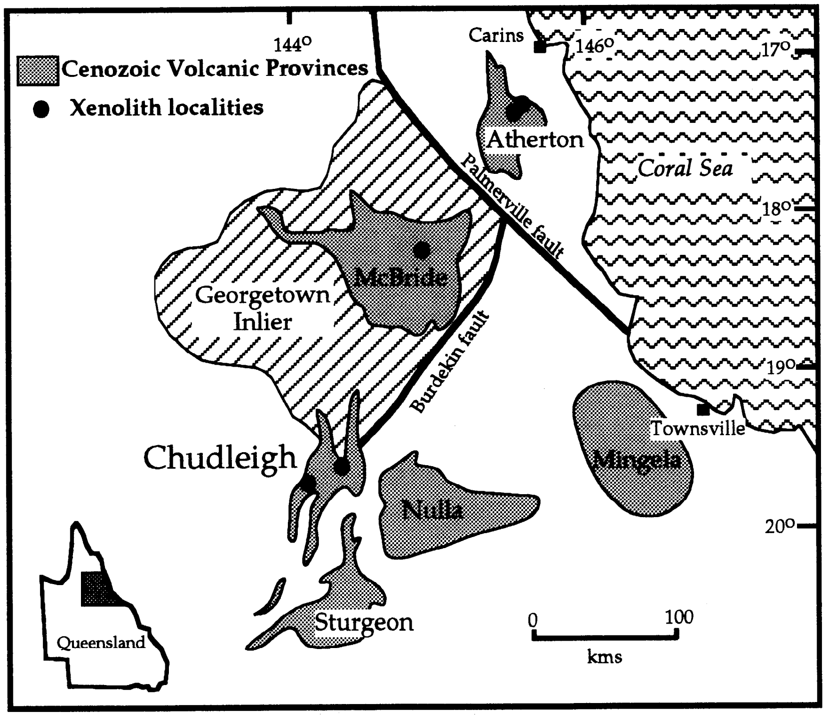

Figure 1. Generalized geologic map showing the location of the Pleio-Pleistocene Chudliegh volcanic province, from which the xenoliths described here were derived.

granites and metamorphic rocks of the Georgetown Inlier from Paleozoic sedimentary and volcanic rocks of the Tasman fold belt (Figure 1). Crustal thickness and structure in this area are not well known. Depth to the Moho is estimated to be ~40-45 km, based on gravity and early seismic refraction data [see *Dooley* , 1980, and references therein]. The nearest recent seismic refraction survey was conducted in the Eromanga basin [*Finlayson, et al.*, 1984], over 900 km to the south, and it is hazardous to extrapolate these findings to the northern volcanic provinces. Nevertheless, the Tasman fold belt in eastern Australia is generally characterized by relatively thick crust (40-50 km) having high *P* wave velocities in the deeper portions [*Finlayson, et al.*, 1980, *Finlayson, et al.*, 1984, *Finlayson and Leven*, 1987].

Petrography, mineral chemistry, whole rock major element and trace element geochemistry, as well as isotopic compositions of the xenoliths are given by *Rudnick et al.* [1986] and *Rudnick and Taylor* [1991]. These papers also discuss in detail the xenoliths' equilibration conditions and petrogenesis, which shall only be summarized here (Table 1).

The xenoliths are interpreted to be cogenetic mafic cumulates that crystallized in the deep crust less than 100 m.y. ago [*Rudnick, et al.*, 1986]. Plagioclase-rich xenoliths are the most common variety and show little variation in major element composition over a very wide range in mineralogies. Table 2 gives the average composition of the six plagioclase-rich xenoliths studied here, along with mafic compositions for which subsolidus phase assemblages have been determined.

Table 1. Description of Mafic Granulite Xenoliths Investigated Here, Listed in Order of Increasing Derivation Depth

Sample	Rock Type	Grain Size, mm	Depth of Origin, km	T°C*	Mineralogy,† modal %	Mean Atomic Weight
83-107	olivine-two pyroxene granulite	~0.3	< 20	800-850	pl(60)-px(18)-ol(13)-sp(9)	21.4
83-114	garnet-two pyroxene granulite	~1.5	26-33	750-800	pl(39)-px(34)-gt (21)-opaques(4)-sp(1)	21.6
83-127	garnet-two-pyroxene granulite	3	26-33	850-900	pl(46)-gar(27)-px(25)-sp(2)	21.4
83-117	garnet-two-pyroxene granulite	~0.6	26-33	900-940	gt(46)-px(28)-pl(25)-sp(2)	21.3
83-125	garnet granulite	0.5-4	36-49	850-910	gt(40)-pl(34)-px(25)	21.3
BC	garnet granulite	~0.3	36-49	950-1040	px(42)-gt(32)-pl(24)-sp(1)	21.4

*Temperatures were determined from either *Wells* [1977] two-pyroxene thermometry or *Ellis and Green* [1979] garnet-clinopyroxene thermometry.

†Modal percentages were established by point counting (≥ 1000 points; minerals are pl, plagioclase; ol, olivine; px, pyroxene; sp, spinel; gt, garnet).

Table 2. Average Composition of Plagioclase-Rich Chudleigh Granulite Xenoliths Compared With Experimentally Investigated Compositions

	Composition					
	1		2	3	4	5
	$\bar{x}$	σ				
SiO_2	50.0	0.46	50.3	49.93	50.3	52.16
TiO_2	0.42	0.56	0.70	1.34	1.7	1.86
Al_2O_3	19.5	1.20	17.0	16.75	17.0	14.60
FeO*	6.88	1.10	8.05	11.40	8.27	9.50
MnO	0.13	0.02	0.17	0.18	0.16	0.14
MgO	10.4	1.00	8.3	7.59	7.8	7.36
CaO	10.0	1.50	11.2	9.33	11.4	9.44
Na_2O	2.58	0.20	2.8	2.92	2.8	2.68
K_2O	0.18	0.07	0.33	0.37	0.18	0.73
P_2O_5	0.07	0.04	0.08	0.19	--	0.18
Total	100.23		98.93	100.00	99.60	98.65
Mg #	72.9		64.8	54.2	62.7	58.0
Mean atomic weight	21.4		21.9	21.9	21.8	22.1

Compositions are 1, average of the six plagioclase-rich Chudleigh xenoliths studied here; 2, delegate two pyroxene lower crustal granulite xenolith (R698 of *Irving* [1974]); 3, olivine tholeiite [*Ito and Kennedy*, 1971]; 4, high-Al basalt [*Green and Ringwood*, 1967]; and 5, olivine tholeiite (B) [*Green and Ringwood*, 1971]. Mg # = 100 (Mg/Mg + ΣFe). Mean atomic weights after *Birch* [1961]. Total Fe as FeO.

Compared with melt compositions, the xenoliths have higher Mg # (Mg # = 100 (Mg/Mg + ΣFe)) and Al_2O_3, and lower TiO_2, reflecting their cumulate character and their abundant plagioclase.

Given the similarity in bulk composition between the different xenoliths studied here, variations in mineralogy can be directly related to the PT conditions that each experienced. Figure 2 shows the experimentally determined equilibrium PT fields for a bulk composition comparable to those of the Chudleigh xenoliths. Were there no kinetic barrier, the mineralogy of the unit represented by each xenolith (including the compositions of coexisting phases) would be expected to have adjusted progressively to maintain thermodynamic equilibrium as the rock cooled below the solidus, toward the ambient geotherm. Under these idealized circumstances, the resulting mineralogy would follow that predicted for a given geotherm (Figure 2) and would appropriately be described as fully "equilibrated" (cf. *Furlong and Fountain* , 1986).

However, the Chudleigh xenoliths contain in their textures and phase compositions, a clear record of incomplete equilibration, which is the consequence of increasingly sluggish diffusion as the temperature falls well below the solidus. The shallowest-derived xenoliths contain relict igneous textures and the lowest grade mineralogies (olivine-bearing), whereas the deepest-derived xenoliths contain fully metamorphic textures and highest-grade mineralogies, approaching eclogite [*Rudnick, et al.*, 1986, *Rudnick and Taylor*, 1991]. Reaction coronas of (1) olivine rimmed by orthopyroxene and spinel-pyroxene symplectites and (2) spinel rimmed by garnet suggest isobaric cooling and reequilibration of the rocks in the deep crust between 17 and 44 km depth, as shown by the arrows in Figure 2. The heads of the arrows indicate the temperatures (determined by application of cation exchange thermometry to analyses performed on adjacent mineral rims) below which no further equilibration was able to occur, i.e. those at which the mineralogy was frozen in or "quenched". These temperatures diverge markedly from the steady state conductive geotherm at midcrustal depths; at greater depths the temperatures of last equilibration approach the steady state geotherm more closely. We attribute this feature to the incomplete reequilibration of the xenoliths, which is most pronounced at the shallowest levels, where temperatures are cooler and reactions more sluggish. In this respect it is interesting that many of the cation exchange temperatures record values just slightly below the experimentally determined phase boundaries (Figure 2). This suggests that the isobaric cooling reactions facilitated cation exchange, as postulated by *Ellis and Green* [1985].

Due to the paucity of geophysical data for the crust in this region, it is difficult to ascertain how representative these xenoliths are of the lower crust. Geochemical data [*Rudnick, et al.*, 1986] provide compelling evidence that these rocks formed by mixing between mantle-derived basaltic magma and more evolved crustal compositions; the latter of which are not found in the Chudleigh xenolith suite but must be present in the deep crust in order to explain the observed trace element and isotopic compositions of the xenoliths. Moreover, models for the thickness of an underplated layer that can be produced from adiabatically upwelling mantle generally do not

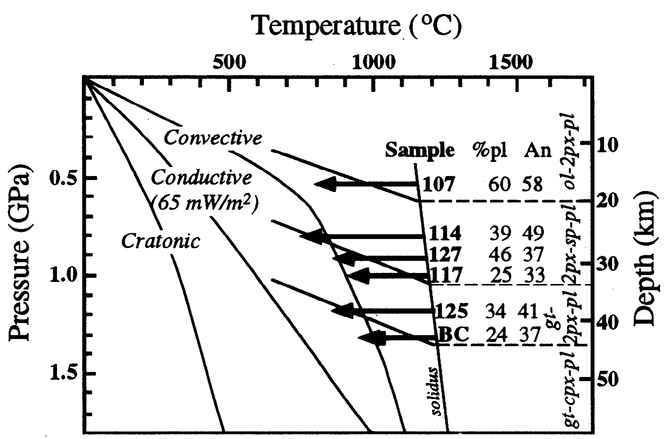

Figure 2. Temperature versus depth plot showing relative derivation depths for Chudleigh granulite xenoliths. Heavy arrows show subsolidus cooling paths which xenoliths followed, as deduced from reaction coronas. The size of the arrow head is equal to the range of calculated temperatures from rim-rim analyses [from *Rudnick and Taylor*, 1991]. Subsolidus stability fields for Delegate xenolith [from *Irving*, 1974] are marked by solid straight lines; mineral assemblage labels are at right (also see Table 1). Numbers to the right of the arrows are the sample number, modal percent of plagioclase present, and the anorthite content of the plagioclase, respectively. Three geotherms are illustrated, the convective geotherm of *O'Reilly and Griffin* [1984], determined for xenoliths from southeastern Australia, a conductive geotherm from *Chapman and Furlong* [1992], based on the surface heat flow of 65 mW/m² [*Cull*, 1991], and a conductive geotherm derived from studies of cratonic peridotite xenoliths [*Boyd*, 1973].

range above 10 km [*Furlong and Fountain*, 1986]. Thus it is unlikely that the lower 25 km of crust in this region is composed entirely of underplated mafic rocks. We will return to the question of how much crust is added by underplating in a later section.

The mineralogies, grain sizes, inferred derivation depths and mean atomic weight of the xenoliths studied here are provided in Table 1. Freshest samples were chosen for measurement in all cases except sample 83-117, which has a large amount of grain boundary alteration and hematite staining, thus allowing for qualitative comparisons of alteration effects on measured *P* wave velocity.

One feature common to all garnet-bearing xenoliths of this suite is the partial to complete replacement of garnet with kelyphite, a secondary alteration product (Figure 3). In the samples measured here, no pristine garnet is preserved. The kelyphite retains the composition and morphology of garnet but consists of submicron acicular to equant crystals of spinel, anorthite, and pyroxenes (identified through transmission electron microscopy). Previous investigations suggest that kelyphite forms due to garnet breakdown caused by decompression of the xenolith during the rapid eruption of the host [*Garvie and Robinson*, 1984, *Kay and Kay*, 1983, *Padovani and Carter*, 1977] and may even involve melting and quench crystallization [*Richet*, 1988]. The presence of kelyphite-rimmed cracks within partially replaced garnets (Figure 3) suggests that such an origin applies to the kelyphite in the Chudleigh xenoliths, as cracks are unlikely to develop in rocks under the high-pressure, fluid-absent conditions in the lower crust. Since plagioclase and pyroxene have markedly lower *P* wave velocities than the original garnet, kelyphite formation results in measured velocities that are significantly lower than those of the original garnet-bearing assemblage in the lower crust. For these reasons, the effects of kelyphite formation (and any other decompression-induced alteration) must be taken into account when comparing measured velocities with data from seismic profiles.

Laboratory Techniques

Three orthogonal cores, ≥ 25 mm in length and 15 mm in diameter, were prepared from each of the selected xenoliths. Most samples showed no preferred mineral orientation [*Rudnick, et al.*, 1986], but where observed (e.g. 83-107), the cores were cut parallel and orthogonal to it. The ends of the cores were ground flat and parallel within ±0.01 mm. Densities were measured by immersion in ethanol and are reported in Table 3. Due to the large numbers of microcracks, the specimens were saturated for several hours before measurement. However, it is not likely that the alcohol penetrated all cracks, so the reported densities represent minimum values. The specimens were jacketed in copper sleeves sealed with O-rings against hardened steel end pieces in order to exclude the fluid pressure medium. After each successful run the steel end pieces were removed and the sample examined to ensure that fluid had not penetrated into the jacketted specimen. The velocity measurements were performed using a Harwood pressure vessel which operates semiroutinely to 1.0 GPa confining pressures. Details of the ultrasonic measurements are given by *Jackson and Arculus* [1984]. Rock velocities were calculated from the zero-pressure core length and measured travel time of the pulse. Estimated uncertainties of the measurements are 0.05 km/sec for *P* wave velocity and 0.01 g/cm^3 for ρ.

Results

Densities and 1.0 GPa room temperature ultrasonic compressional wave velocities for the xenoliths are given in Table 3, where the samples are listed in order of increasing metamorphic grade, hence derivation depth. The change in *P* wave velocity with increasing pressure is shown in Figure 4

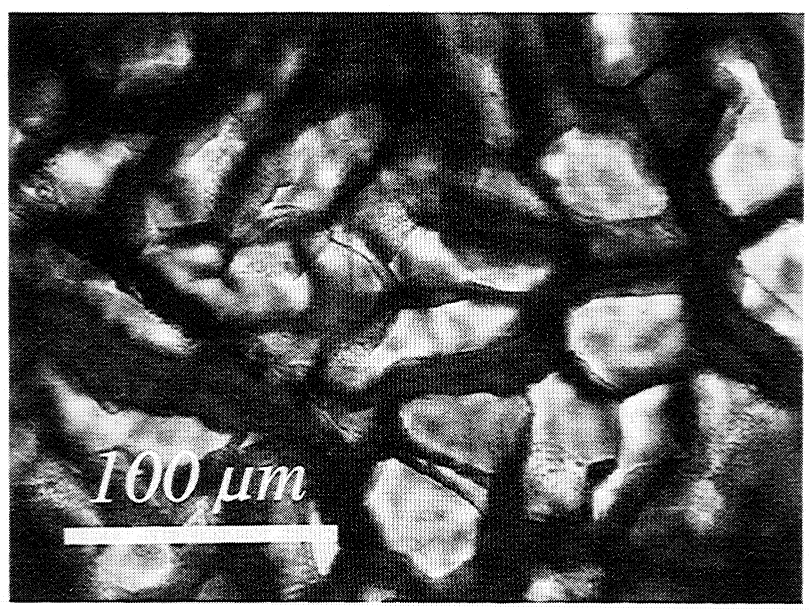

Figure 3. Partial replacement of garnet (light areas) by kelyphite (dark) in sample 83-133. Note the preferential development of kelyphite along cracks. Garnets are completely replaced by kelyphite in the samples investigated here.

Table 3. Results of Ultrasonic Measurements Performed at 1.0 GPa and Room Temperature

Sample (Core)	ρ, kg/m³	ρ,* kg/m³	V_p, km/s	V_p,* km/s	Percent Anisotropy	Calculated V_p†
83-107						
(1)	2830	---	6.78	---		
(2)	2830	---	6.74	---		
(3)	2910	---	6.92	---		
Mean	2850	---	6.86		2.6	7.47
83-114						
(1)	2940	3030				
(2)	2920	3010	6.89	7.06		
(3)	2920	3010	6.85	7.02		
Mean	2930	3020	6.87	7.04	0.5	7.68
83-127						
(1)	2940	3060				
(2)	2950	3070	7.26	7.52		
(3)	2950	3070	7.02	7.28		
Mean	2950	3070	7.12	7.40	3.4	7.68
83-117						
(1)	3040	3210	6.82	7.16		
(2)	2920	3090	6.83	7.17		
(3)	2980	3150	6.86	7.20		
Mean	2980	3150	6.84	7.18	0.6	8.09
83-125						
(1)	2970	3120	7.00	7.30		
(2)	2950	3100	7.04	7.34		
(3)	2950	3100	7.06	7.36		
Mean	2960	3110	7.03	7.33	0.8	7.92
BC (1)	2.99	3.12	7.26	7.52		
(2)	3.02	3.15	(7.2)	(7.4)		
(3)	3.05	3.18	(7.4)	(7.7)		
Mean	3.02	3.15	(7.3)	(7.6)	2.6	7.99

Samples listed in order of increasing derivation depth. V_p in parentheses were extrapolated to 1.0 GPa from 0.8 GPa.

*Corrected for kelyphite; see text for description of correction procedure.

†Corrected for 1 GPa pressure using 0.12 km s^{-1} GPa^{-1}. See text for explanation of calculations.

for sample 83-114, this curve is similar to that observed in the other samples. Note that the velocity appears to increase linearly with pressure only above about 0.6 GPa. This linear regime yields pressure derivatives ranging from 0.2 to 0.5 kms^{-1} GPa^{-1}, which are significantly greater than those determined by *Christensen* [1974] for a variety of mantle rock types at 3.0 GPa pressures (these ranged from 0.10 to 0.14 kms^{-1} GPa^{-1}, consistent with single-crystal elasticity data for the constituent minerals). This suggests that even 1 GPa confining pressure is insufficient to close all cracks in the samples, possibly due to irreversible changes in the crystal boundaries caused by rapid decompression of the xenolith in the pipe. A pressure derivative of 0.12 kms^{-1} GPa^{-1} was used to correct the calculated 1 GPa velocities for in situ pressures when calculating velocity-depth profiles.

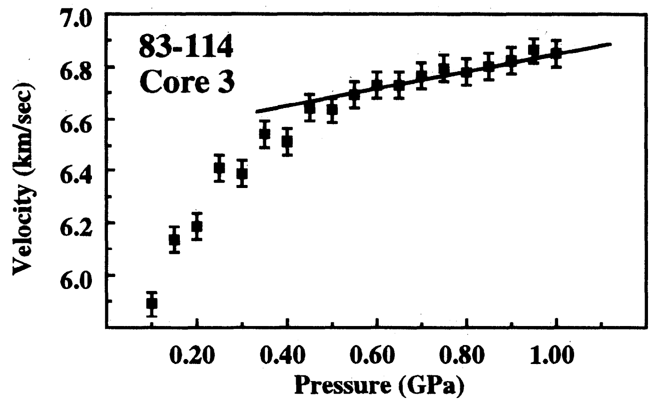

Figure 4. Pressure versus velocity for single core of 83-114. Error bars mark estimated uncertainty in velocity measurements. Note linear increase of velocity with pressure above about 0.6 GPa.

Effects of Kelyphite Formation on *P* Wave Velocity

For garnet-bearing samples (114, 127, 125, 117, and BC) the measured *P* wave velocities are lower than the expected in situ velocities due to the decompression-induced breakdown of the garnets. The predicted in situ *P* wave velocities may be calculated by comparison of kelyphite velocity to that of fresh garnet. Wide-beam electron probe analyses of the kelyphites show them to be relatively homogenous with a narrow compositional range from Py_{55}-Al_{31}-Gr_{14} to Py_{64}-Al_{22}-Gr_{13}. The preferred least squares solution II of *Babuska et al.* [1978] yields velocities of 8.90-8.97 km/sec for this compositional range; a value of 8.94 km/sec was therefore employed in the calculations. Transmission electron microscopy (TEM) of a kelyphitized garnet from BC shows it to consist of orthopyroxene, anorthite, spinel, and clinopyroxene. The relative proportions of these phases in the kelyphite were determined in two ways: (1) through a "point count" based on the identification of over 50 randomly selected grains in the TEM and (2) through mass balance calculations using least squares mixing of the garnet composition as determined from electron probe analyses and the kelyphite phase compositions as estimated from energy dispersive spectra from the TEM. The results of these estimates are given in Table 4. There is remarkable agreement between the two estimates, considering the low number of grains included in the "point count." Single-crystal velocity and density data for the kelyphite phases were taken from *Christensen* [1982] and are listed in

Table 4. Parameters Used in Corrections of V_p for Kelyphite

Mineral	Percent	V_p, km/s	ρ, kg/m³
Anorthite	21 (22)	7.2*	2760*
Spinel ($Mg_{69}Fe_{31}Al_2O_4$)	21 (26)	9.4	3790
Orthopyroxene ($En_{82}Fs_{17}$)	42 (40)	7.81	3350
Clinopyroxene ($Di_{92}Hd_8$)	16 (10)	7.82	3310
Σ (=kelyphite)	100	7.95	3300

Percentages in parentheses were determined through "point count."

* Extrapolated from data of *Christensen* [1982].

Table 4. Thus the kelyphite has an estimated density of 3.30 g/cm^3 and a *P* wave velocity of 7.95 km/sec (Table 4).

Using the modal abundance of kelyphite established from point counting of thin sections (see Table 1), in situ densities and velocities can be calculated. These are given in Table 3, columns 3 and 5.

Comparison of Measured and Calculated *P* Wave Velocity

The fidelity with which the measured velocities reflect the primary mineralogy of the xenoliths has been assessed by comparison of the measured velocities with those calculated for the modal mineralogy from appropriate single-crystal elasticity data. A summary of the strategy behind these calculations and the relevant elasticity data for mineral end members is given by *Jackson et al.* [1990].

The calculated velocities, given in Table 4, are systematically higher (by 5 to 12%) than the kelyphite-corrected measured ones. In theory, these discrepancies could be due to (1) scattering of elastic waves at grain boundaries, (2) the presence of significant Al and/or Na in the natural pyroxenes, which are not accounted for in the pyroxene end-members used for the calculations, (3) differences between the modal mineralogy determined on a single thin section from those in the actual rock core on which the measurements were made, (4) grain boundary alteration in the xenoliths, and/or (5) failure to close all pore spaces due to irreversible changes occurring on grain boundaries or in the process of kelyphtization. These possibilities are discussed sequentially.

Scattering at grain boundaries will become important if the wavelength of the acoustic disturbance becomes comparable to the grainsize. At frequencies near 1 MHz, wavelengths are a few millimeters and some scattering would be expected, especially in relatively coarse grained rocks for which the wavelength is not significantly greater than the grainsize. Indeed, in measurements with 5-Mhz transducers, as performed here, the dominant frequency of the transmitted signals is often somewhat less than 5 MHz because of scattering of the higher-frequency energy. However, provided that sufficient energy associated with lower frequencies travels to the receiver via the shortest available path, the correct velocity will still be measured using this technique, which is based on first-arrival traveltimes.

Although the pyroxenes in these xenoliths do have appreciable amounts of Al_2O_3 and Na_2O [*Rudnick and Taylor*, 1991], the effect of inclusion of the tschermaks and jadeite components would be to increase, rather than to decrease, the calculated velocities of the pyroxene solid solutions [*Kandelin and Weidner*, 1988]. This suggests that the calculated velocities are minimum estimates for the primary mineralogy of the xenoliths. It is likewise unlikely that the velocity discrepancies are due to inaccurate modes, since this would not be expected to produce a systematic shift between measured and calculated velocities.

Grain boundary alteration and residual porosity remain the most likely explanations for the differences. Grain boundaries in the Chudleigh xenoliths do show alteration, characterized by secondary oxides, glass, and fluid inclusions. In this respect they are similar to xenoliths from Calcutteroo, in which no systematic differences occur between measured and calculated velocities [*Jackson, et al.*, 1990]. Nevertheless, sample 83-117, which shows the largest amount of grain boundary alteration in the Chudleigh samples also shows the greatest discrepancy between the kelyphite-corrected *P* wave velocity and that calculated from single crystal data (Table 3), suggesting that alteration has played a role in reducing the measured velocities. In addition, the large pressure derivatives derived from our measurements, as discussed above, suggest that not all cracks and pores are closed at 1.0 GPa. For these reasons, the calculated *P* wave velocity will be used in subsequent discussions.

Discussion

The calculated, room temperature velocities for the xenoliths at 1 GPa range from 7.5 to nearly 8.1 km/sec and show a direct correlation with mineralogy, hence derivation depth (Table 3). In order to compare these velocities with those of refraction profiles, the effects of in situ temperature, which will lower the velocity approximately 6x10^{-4} km s^{-1} K^{-1} [*Christensen*, 1979], must be evaluated. The in situ temperature can be estimated in two ways: (1) by cation

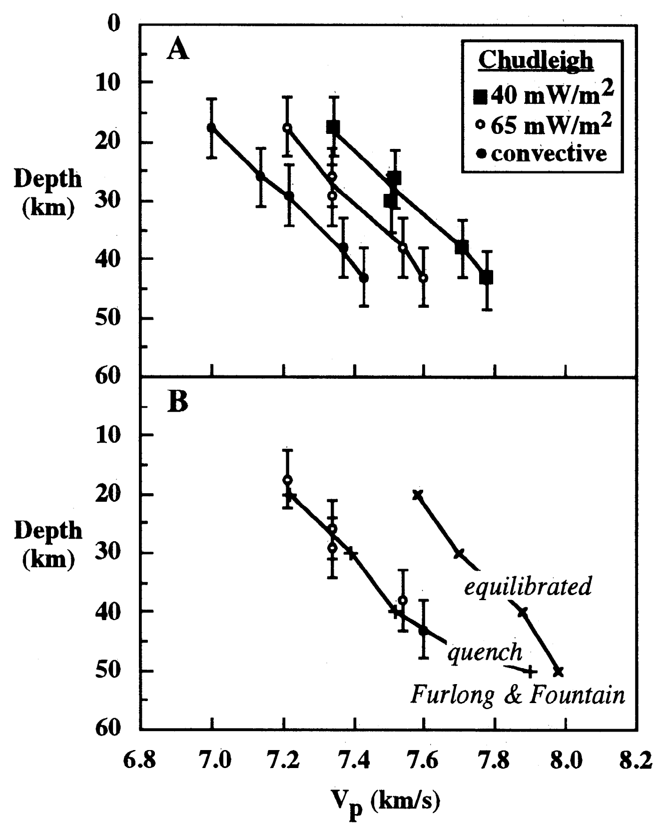

Figure 5. (a) Velocity-depth profile determined for Chudleigh xenoliths that have been temperature-corrected to different geotherms. (b) 65 mW/m^2 geotherm corrected *P* wave velocities for Chudleigh xenoliths compared with those calculated by *Furlong and Fountain* [1986] for underplated olivine tholeiite composition. Quench and equilibrium labels mark velocity profiles calculated for these degrees of equilibration. The *Sobolev and Babeyko* [1989] results for "cooled" olivine tholeiite closely match the equilibrated profile of Furlong and Fountain and are not shown here. Error bars on Chudleigh data represent ± 0.5 GPa.

exchange thermometry for each xenolith [see *Rudnick and Taylor* 1991] or (2) by projecting the inferred equilibration depth on to a geotherm. For this purpose several geotherms are available: a hot, xenolith-derived geotherm for southeastern Australia [*O'Reilly and Griffin*, 1984], a calculated steady state geotherm based on the observed surface heat flow of 65 mW/m^2 [*Cull*, 1991] (which is typical of many Phanerozoic regions), and a cold, cratonic geotherm [*Boyd*, 1973], which may be representative of the thermal state of Precambrian cratons. All three are shown in Figure 2.

Depending upon which geotherm is used to correct the calculated *P* wave velocities, the results range from values generally considered crustal (i.e., 7.0-7.4 km/sec for hot geotherms) to those that would be interpreted as mantle (i.e., > 7.7 km/sec for cold geotherms) on velocity profiles. Thus in regions that have experienced basaltic underplating, some ambiguity in the inferred depth of the crust-mantle boundary is to be expected.

The temperature-corrected velocities are plotted as a function of depth in Figure 5. Also shown are V_p-depth profiles calculated by *Furlong and Fountain* [1986] for the olivine tholeiite composition in a crust having a heat flow of 60 mW/m^2, similar to the north Queensland example. The Chudleigh samples have distinctly lower *P* wave velocity than the equilibrated profile of Furlong and Fountain, but closely match their "quench" profile in both absolute velocity and velocity gradient. This highlights the importance of reaction kinetics in defining the ultimate velocity profile for underplated crust [e.g., *Sobolev and Babeyko* , 1989].

The velocity profiles determined here are compared with profiles from other areas of eastern Australia as well as "type sections" derived from compilations of seismic data for various types of crust in Figure 6. Two features are apparent

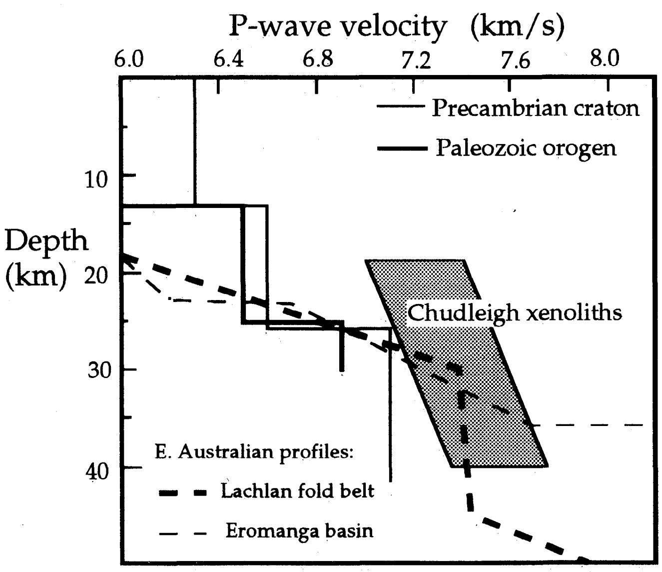

Figure 6. *P* wave velocity variations with depth for various regions compared with the results for the Chudleigh xenoliths. Profiles labeled "Precambrian craton" and "Paleozoic orogen" are type sections from Rudnick and Fountain (1995). Eastern Australian profiles are from *Finlayson et al.* [1984] (for Eromanga basin, southern Queensland) and *Finlayson et al.* [1980] (for Lachlan fold belt, southeastern New South Wales). Shaded region represents calculated velocity-depth variation for Chudleigh xenoliths under a range of crustal geotherms (see Figure 2).

from this diagram: (1) most lower crustal velocities although high, fall below those of the mafic granulites studied here and (2) high velocities, matching those of the mafic granulites, are generally restricted to depths of 30 km or more (as in the velocity profiles for eastern Australia). This suggests that high velocity (i.e., 6.9-7.2 km/sec) lower crustal layers that are so typical of refraction profiles [e.g., *Holbrook et al.*, 1992; Rudnick and Fountain, 1995] probably represent a mixture of mantle-derived basaltic lithologies and pre-existing, quartz-bearing crust. The presence of evolved rocks in the lower crust of north Queensland is supported by intermediate and metasedimentary xenoliths from the nearby McBride province [*Rudnick and Taylor*, 1987] (Figure 1) and the isotopic and trace element characteristics of the Chudleigh granulites, as discussed above.

Conclusions

Measured ultrasonic velocities for an isochemical suite of lower crustal xenoliths from the Chudleigh volcanic province, north Queensland, are systematically lower by 5-12% than velocities calculated from single-crystal data. We believe this is due to irreversible changes that occurred to the xenoliths during rapid decompression in the pipe. The *P* wave velocities calculated from single-crystal elasticity data show a large range and correlate well with the degree of metamorphic equilibration that the xenoliths experienced, which correlate in turn with their derivation depths. The xenolith derived from the shallowest level, ~18 km depth, contains olivine-cored coronas and has the lowest velocity (7.5 km/sec), whereas those derived from the deepest levels, ~44 km depth, are well equilibrated, approach an eclogite mineralogy, and have the highest velocities (8.0-8.1 km/sec). When corrected for in situ temperature, the *P* wave velocities range from crustal to mantle-like values. The temperature-corrected velocity-depth profile closely matches that of "quenched" basaltic underplate as modeled by *Furlong and Fountain* [1986]. However the Chudleigh xenoliths range from quench-type mineralogy to nearly fully equilibrated mineralogies at the deepest crustal levels, highlighting the importance of reaction kinetics in determining the final velocity profile of underplated material. The xenolith's *P* wave velocities are higher than the high-velocity layers that are common in the lower crust. If these layers formed by basaltic underplating it is likely that they contain some evolved lithologies in addition to the underplated material, a conclusion consistent with chemical and isotopic compositions for many lower crustal xenoliths [*Rudnick*, 1992].

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