Paper 5:

Multi-genetic origin of crustal reflectivity: a review of seismic reflection profiling of the continental lower crust and Moho

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1. Introduction

Seismic reflection profiles provide the highest resolution information on the insitu structure and physical properties of the lower continental crust and Moho. The seismic properties most readily obtained from reflection profiles are reflectivity patterns, which correlate with distinct geologic settings. When reflectivity patterns are interpreted with complementary seismic velocity and non-seismic crustal parameters, inferences regarding the composition and evolution of the lower crust can be made.

The study of the lower continental crust with seismic reflection data is a rapidly evolving field; improved data sets continually provide new insights and alter initial concepts based on more limited data. In this chapter we summarize the most important observations and conclusions derived from reflection seismology and relate these to models of the physical properties and evolution of the lower crust. We focus particularly on the structure of the lower crust in diverse tectonic settings, the nature of the Moho, and the multi-genetic origin of crustal reflectivity. The reader should consult the companion papers by Holbrook et al. (this volume) and Jones (this volume) for the related topics of seismic velocities and conductivity in the lower crust.

2. The seismic reflection method

Seismological observations at the beginning of this century led to the detection of the Earth's core in 1909, and the crust-mantle boundary, the Moho (Mohorovicic, 1910). Seismic refraction measurements of potential hydrocarbon traps within sedimentary structures also began at about the same time, but it was not until the end of the 1930's that useful seismic reflection data became available (Sheriff and Geldart, 1982). This relatively late start in the widespread application of reflection seismology was due to the great technical effort needed to enhance near-vertical reflected signals that often are 10,000 times weaker than the seismic ground-roll from near-surface layers. In the 1940's and 1950's new techniques, like common depth point (CDP) stacking, versatile seismic sources, and magnetic tape recording were developed and contributed to the emerging superiority of reflection profiling over refraction seismics for prospecting in sedimentary basins.

2.1. Early deep crustal seismic reflection studies

It was in the 1950's that the first sporadic observations of near-vertical reflections from the lower continental crust were reported (Junger, 1957; Galfi and Stegena, 1957). However, serious doubts about the reality of these deep reflections were expressed by some workers (e.g., Steinhart and Meyer, 1961). These doubts led to an "experimentum crucis" in 1964 near Augsburg, Germany, where deep crustal reflectors were observed at both near-vertical and wide-angle incidence (Meissner, 1967), thus confirming, at least in this area of the Bavarian Molasse Basin, the existence of deep reflectors. At about the same time, similar investigations in Australia by the Bureau of Mineral Resources (BMR) (Moss, 1962; additional work summarized in Moss and Dooley, 1988, and Dooley and Moss, 1988) and in Canada (Kanasewich and Cummings, 1965; Clowes et al., 1968) also gave conclusive evidence for intracrustal and Moho reflections at near-vertical incidence. The lower crust was shown to be highly reflective at near-vertical incidence, and the

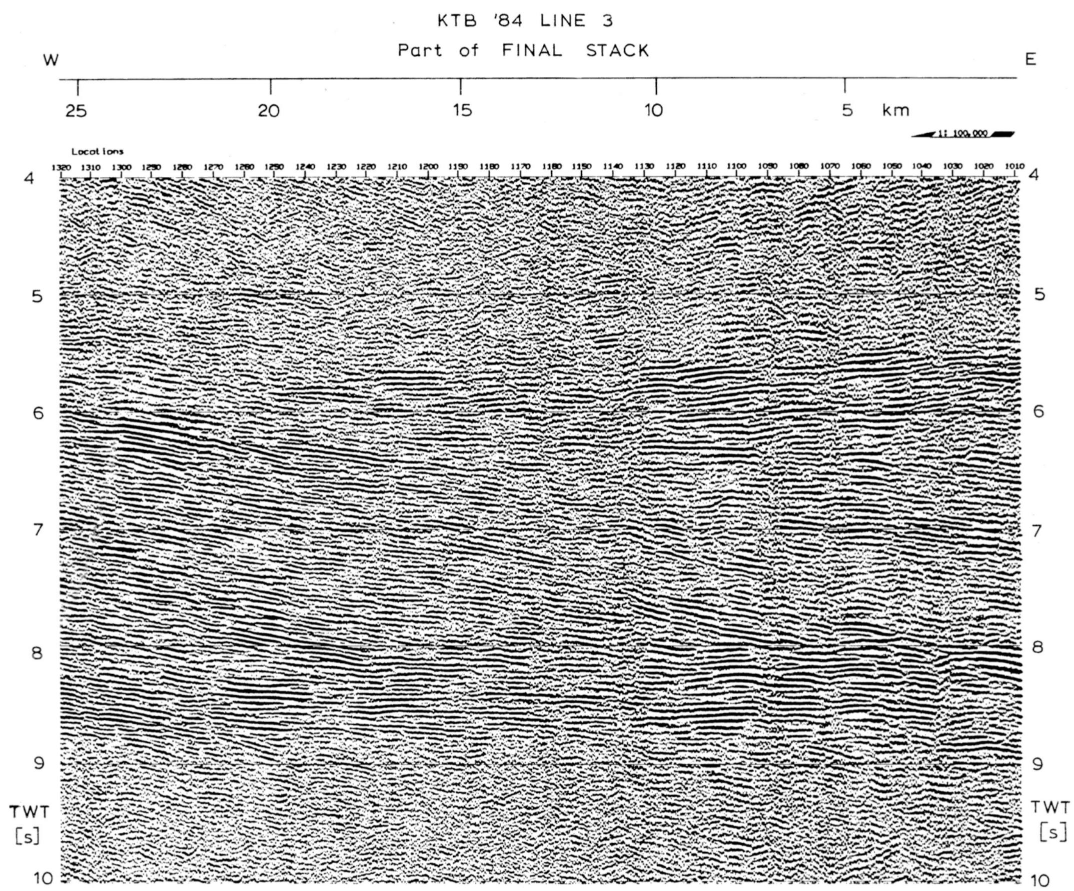

Fig. 2-1. Lower crustal seismic lamellae (or "laminae") as observed on a seismic reflection profile recorded in the Black Forest, SW Germany (Lueschen et al., 1987). Crustal reflectivity increases abruptly at about 5.5 s two-way time (TWT and terminates at 9 s at the Moho. The high amplitudes of the reflections requires the existence of significant velocity contrasts (± 5% or larger) in the lower crust. Lower crustal seismic lamellae have been recorded throughout the world with varying amplitudes, lateral continuity, and dips depending on the tectonic setting.

names "lamellae" or "laminae" for the seismic layering in the crust were coined (Glocke and Meissner, 1976). Seismic lamellae are depicted in Figure 2-1.

2.2. The last 15 years of deep crustal reflection studies

Sporadic seismic reflection campaigns continued in the early 1970's across important geological structures in Germany (Dohr and Meissner, 1975) and elsewhere. However, the year 1975 marks the beginning of a nearly continuously operating seismic crew by COCORP (Consortium for Continental Reflection Profiling) in the U.S. (Oliver et al. 1976; Brown et al., 1986). Since then, a concerted effort to study the continental crust has been made, and similar groups have been formed elsewhere in the USA (U.S. Geological Survey, Univ. of Wyoming, and CALCRUST), Great Britain (BIRPS), Canada (LITHOPROBE) and the Geological Survey of Canada), France (ECORS), Australia (ACORP and the BMR), Germany (DEKORP), Switzerland, and elsewhere (summarized in volumes edited by Barazangi and Brown, 1986a,b; Matthews and Smith, 1987, Leven et al., 1990).

Many of these research groups did not attempt to measure lower crustal seismic velocities. The use of spread lengths of about 10 km on land and about 3 km at sea did not permit the determination of seismic velocities to depths greater than about 10 km. The detailed picture available on near-vertical sections was so fascinating, and data were so quickly collected especially at sea that crustal velocities were of secondary interest. More recently, considerable effort has gone into the determination of seismic velocities (P-wave, S-wave) and anisotropy in an effort to place tighter constraints on crustal composition and physical state (e.g., Bartelson et al., 1982; Clowes et al., 1987; Mooney and Brocher, 1987; Wenzel et al., 1987; Fisher et al., 1989; Valasek et al., 1989; BABEL Working Group, 1990; Goodwin and McCarthy, 1990; Lueschen et al., 1990).

2.3. Resolution and uncertainties

The acquisition and processing of seismic reflection data has been well advanced by the hydrocarbon exploration industry, as described in such standard texts as Telford et al. (1976), Sheriffand Geldart (1982), Claerbout (1985), Waters (1987), and Yilmaz (1987). However, the application of reflection seismology to the deep continental crust presents special problems not often encountered in the exploration of sedimentary basins. Among the most important problems, as summarized by Mooney (1989), are:

(1) low reflection amplitudes due to geometrical spreading, scattering, and seismic attenuation;

(2) large static corrections (time shifts) for land profiles due to variations in elevation and near-surface geology;

(3) an inadequate knowledge of deep crustal velocities (including their lateral variations), which degrades the effectiveness of the common depth point (COP) stacking method; and

(4) poor data quality where there is high cultural noise, imperfect coupling of vibrators and geophones, or low, uneven-fold data arising from skips or crooked-line recording.

These factors can make the difference between recording coherent deep seismic reflections or incoherent noise (cf., Taner et al., 1970; Johnson and Smithson, 1986; Phinney and Roy-Chowdhury, 1989; Brocher and Hart, 1991).

Smithson and Johnson (1989) point out three other problems that reflection seismologists encounter when working in the crystalline crust: (1) reflection coefficients are smaller for layered crystalline rocks (about 0.1) in comparison with those for layered sedimentary rocks (about 0.3); (2) crystalline rocks are often more deformed and distorted than sedimentary strata; and (3) there is generally a lack of drill hole control to test interpretations. We discuss some cases where upper crustal drill hole control is available in Section 4.1.

The relative vertical resolution of seismic reflections is generally comparable to one-quarter of the wavelength of the seismic signal according to:

$$\lambda = V_p / f$$

where λ = wavelength (km), V_p = compressional wave velocity (km/s), and f = frequency (Hz, cycles/s).

Using a crustal V_p -value of 6.0 km/s and a typical frequency of f = 25 Hz, the seismic wavelength is 240 m and the optimal resolution in the seismic section will be equal to $\lambda/4$, or 60 m. High-frequency signals (f = 50 Hz) show higher resolution (30 m), and in sediments, with their lower V_p -values, resolution may reach 10-20 m. Layers thinner than one-quarter wavelength will also be resolved, up to approximately one-tenth of a wavelength (Fuchs, 1969; Widess, 1973), but signal amplitudes will be lower and favorable signal-to-noise conditions are necessary. The best resolution of the physical properties of crustal reflectors is obtained with broadband (approximately 8-80 Hz and broader) seismic data.

Horizontal resolution is less than vertical (thickness) resolution, and decreases with depth. The concept of the first Fresnel zone (Sheriff and Geldhart, 1982) describes the minimum length of a well-resolved reflecting horizon. Typical Fresnel zone radii (for 25 Hz data) in the crystalline crust are 1.5 km at 10 km depth, and 3 km at 30 km depth. Gaps that are smaller than a Fresnel zone will not be evident and, therefore, a discontinuous reflection horizon will appear to be more continuous than it actually is. Computer modeling of hypothetical structures provides quantitative evaluation of seismic resolution and uncertainties (Hunch and Smithson, 1987; Reston, 1987; Wenzel et al, 1987; Gibson and Levander, 1988, 1990; BIundell, 1990).

The determination of a reflector's absolute depth, for example, the Moho, depends on a knowledge of the velocity structure of the crust. This information normally comes from wide-angle reflection/refraction studies that generally use lower frequencies (5-10 Hz) and thus have correspondingly lower resolution compared with seismic reflection data. The error in the calculated depth (z) of a reflecting interface is proportional to the accuracy of the measured average velocity (V_p). V_p - z (velocity-depth) curves may easily assume errors of 5-6% (Holbrook et al., this volume).

3. Observations or continental lower crust and Moho

Some 40,000 km of deep crustal reflection data have been collected worldwide during the past 15 years in continental areas. About half of the data were obtained by marine surveys in continental shelf areas. Observations in similar geologic environments, for example, young mountain belts, are consistent from region to region. Significantly, reflectivity patterns do not generally appear to be controlled by data acquisition parameters, and data collected in similar geologic environments using vibroseis, explosives, or airguns usually show similar reflectivity patterns. The primary exceptions to this are that explosive and airgun sources appear to produce superior deep reflections for crustal thickness greater than about 25 km, and explosive sources sometimes produce fewer artifacts in land data than vibroseis sources (e.g., Brocher and Hart, 1991). The similarities in crustal reflectivity for common geologic environments is an important observation that indicates common processes of tectonic evolution for these environments.

3.1. Terminology and overview of important contributions

As an introduction to the following sections, some frequently used expressions in the reflection literature are defined, after which we list some of the most important contributions made by seismic reflection studies.

a. *Laminated lower crust*: a lower crust that is full of densely packed, multiple sets of reflections referred to as seismic lamellae or laminae. These reflections are commonly subhorizontal, but occasionally dip, curve, and overlap, and often show a sudden onset at 4-6 s two-way traveltime (twt) (12-18 km depth) and an abrupt termination at the Moho.

b. Transparent upper (or lower) crust: a significant time interval of a seismic section that is relatively free of seismic reflections.

c. Structureless crust: reflections are predominantly short (1.5 km or less in length), the strength of reflections (i.e., reflectivity) generally decreases with depth, and the crust shows few reliable reflectivity patterns (see below); may sometimes be caused by data acquisition problems (section 2.3 Resolution and uncertainties).

*d. Diffractive crus*t: crust is dominated by long (3-5 km or more) hyperbola-shaped events, possibly generated by strong folds or small-scale inhomogeneities, such as vertical intrusives.

e. Seismic duplex: Seismic reflections in a rhomboid shape, similar to cross-sections of structural duplexes, and usually seen in shortened (compressional) environments (Fig. 2-2a).

f. Seismic wedge: Diverging, wedge-shaped seismic reflectors generally seen in shortened (compressive) environments; these are more reliably identified in migrated sections (Fig. 2-2b).

We illustrate and discuss the significance of these features within various geologic setting in the forthcoming sections. Some of the most important contributions of crustal reflection studies in the last 15 years are:

(1) The determination of the geometry at depth of prominent fault zones, some of which reach the lower crust. Examples are the Wind River fault in Wyoming (Smithson et al., 1979), the Hunsruck-fault in Europe (DEKORP Res. Group, 1991), and the Flannan fault north of Scotland (Peddy et al., 1986).

(2) The discovery of the seismically laminated lower crust, sometimes accompanied by a relatively transparent upper crust, in areas of extended crust (Allmendinger et al., 1986, 1987; BIRPS and ECORS, 1986; Matthews, 1986).

(3) The recognition of distinct structural developments for different ancient

orogens; for example, thin-skinned tectonics in the southern Appalachians (Clark et al., 1978; Cook et al., 1981) and the Variscan deformation in Europe (DEKORP Research Group, 1990), compared with thick-skinned compressional features in the New England Appalachians (Brown et al., 1983), the Grenville orogenic belt (Green et al., 1988) and the eastern Rhenish Massif (DEKORP Research Group, 1991).

(4) The confirmation of thick crustal roots under young orogenic belts and the absence of crustal roots in most Caledonian and Variscan belts (Meissner et al., 1987).

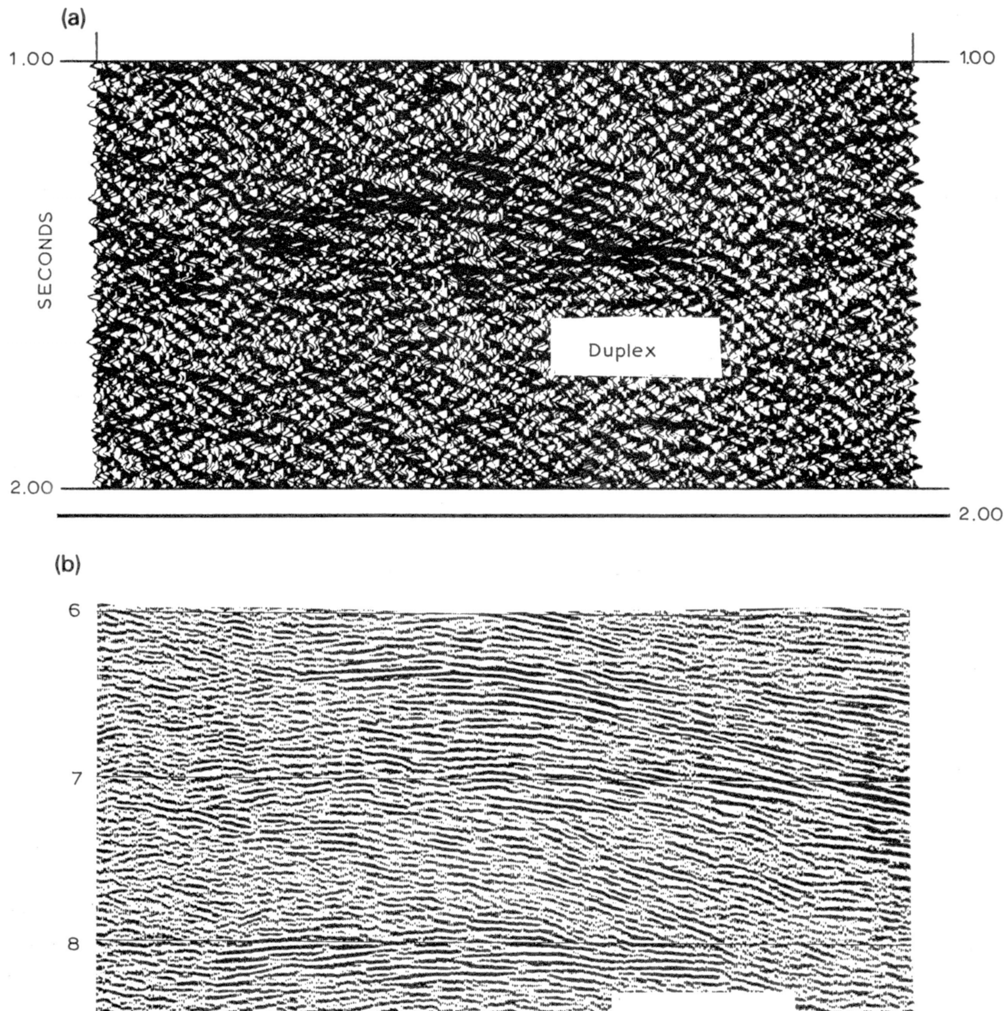

Fig. 2-2. Examples of two common structural features observed in deep seismic reflection data: (A) a seismic duplex (Hubbard et al., 1991) and (B) a seismic wedge (DEKORP Res. Group, 1991). These features are interpretated as resulting from crustal shortening (see sections 2.3.2-23.4).

3.2 Precambrian crust

Early seismic reflection profiles in Precambrian crust indicated a generally structureless lower crust, but revealed pronounced structural features in the upper and middle crust that appear to date from Precambrian times (Oliver et al., 1976; Smithson et al., 1980;

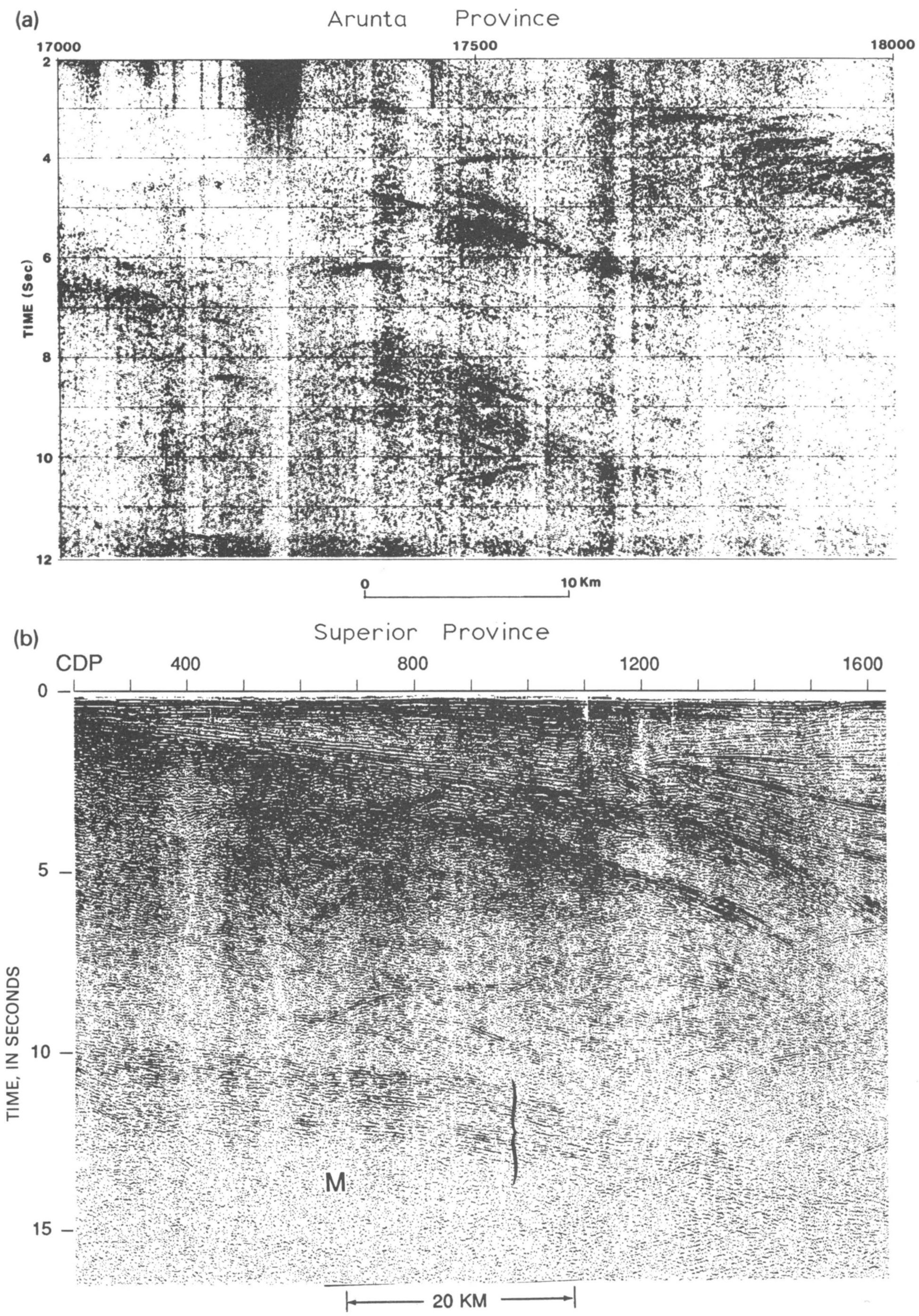

Fig. 2-3. Two examples of seismic reflection data from Precambrian crust: (A) the Arunta Province of central Australia (Goleby et al., 1990) and (B) the Superior Province of central North America (Hutchinson et al., 1991). These profiles were among the first to show strong lower crustal reflectivity in Precambrian crust, and therefore were significant in establishing the fact that a reflective lower crust and Moho (M) is not restricted to Phanerozoic crust. The dipping reflections beneath the Arunta block (A) appear to be related to a Precambrian collision that formed the block, and the reflectivity at 10-12 s TWT (above the Moho) beneath Lake Superior (B) may be related to rifting and igneous activity associated with the formation of the mid-continental rift.

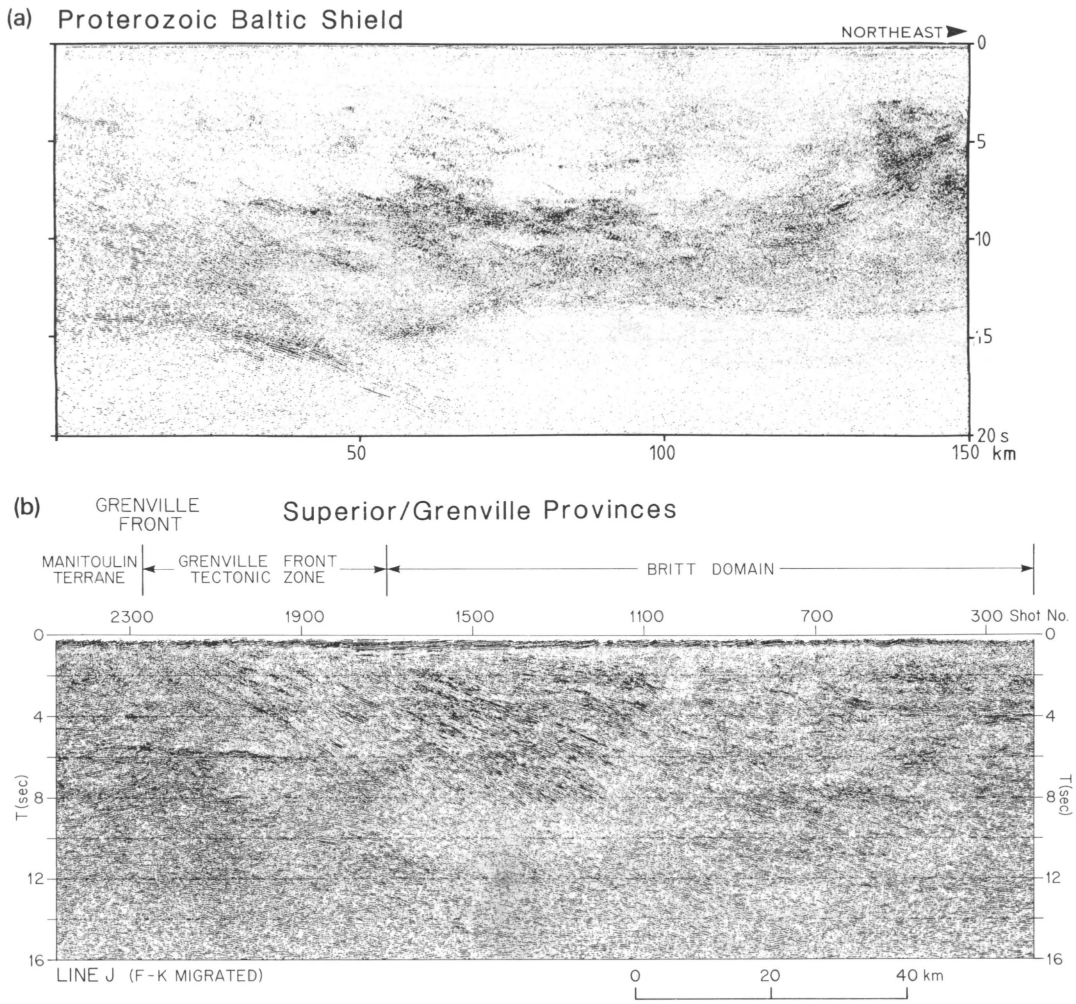

Fig. 2-4. Seismic reflection profiles from (A) the Precambrian crust of the Baltic shield (BABEL Working Group, 1990) and (B) across the Grenville Front beneath Lake Huron of North America (Green et al., 1988, 1989). The Baltic shield data show remarkably high reflectivity in the lower crust (10-15 s TWT) and a clear reflection Moho above a nearly transparent upper mantle. A Precambrian suture is evident at 50 km on the distance scale. Lower crustal reflectivity is less evident across the Grenville Front (B) due to the dominance of the distinctive dipping reflectivity pattern in the upper 8 s of the profiles. Nevertheless, sub-horizontal lower crustal and Moho reflectivity can be traced in a manner similar to other Precambrian reflection sections. The fact that the reflectivity patterns due to ancient tectonic events ("primary" patterns) are preserved beneath Precambrian crust (these examples and Fig. 2-3) attests to the long-term thermal and tectonic stability of the Precambrian crust; specifically, neither ductile flow nor magmatic intrusion from the mantle have occurred on a scale sufficient to disrupt primary lower crustal reflectivity patterns.

Gibbs, 1986; Dahl-Jensen, 1987; Smithson and Johnson, 1989). The most significant features are steeply-dipping zones of reflections that often can be correlated to depths greater than 15 km, although data quality generally is significantly better (and therefore interpretation more reliable) above this depth. The major inference is that observed structures are consistent with modern tectonic processes, such as collisional orogens, crustal extension, and basin development, operating in Precambrian time. Furthermore, the fact that many Precambrian structures have persisted through time (Gibbs, 1986) is indicative of the long-term thermal and tectonic stability of the Precambrian crust.

The lower crust imaged in these early reflection profiles is characterized by a decrease in the number of reflections with depth, and the Moho is indicated only by a few weak and discontinuous reflections. These observations led to the suggestion that the lower crust of Precambrian shields is cold and brittle and has lost its reflectivity due to brittle deformation (Meissner, 1986). The lack of a distinct Moho reflection appeared to be consistent with a transitional Moho beneath Precambrian crust, possibly due to a gradual transition from mafic lower crust to an ultramafic upper mantle (Pavlenkova, 1987, 1988).

These concepts of the seismic properties of the Precambrian lower crust have undergone a significant re-evaluation in the light of recently collected profiles that replaced conventional mechanical seismic sources (vibrators) with more powerful sources (large tuned airgun arrays and dynamite shots). Deep penetrating profiles over Precambrian crust have been recorded in central Australia (Fig. 2-3a; Finlayson et al., 1989; Goleby et al., 1990), the Superior Province of Canada and the United States (Figs. 2-3b and 2-4b; Behrendt et al., 1988, 1989, 1990; Cannon et at, 1989, 1991; Geis et al., 1990; Green et al., 1989; Jackson et at, 1990), Africa (Wright and Hall, 1990) and the Baltic shield (Behrens et al., 1989; BABEL Working Group, 1990; Fig. 2-4a). The common characteristics are strong lower crustal reflectivity and Moho reflections that can be traced in a piecewise continuous fashion across all records. These features have previously been ascribed only to Phanerozoic crust (e.g., Meissner, 1986). Precambrian lower crust, like Phanerozoic crust, appears to possess the full range of seismic reflectivity responses, ranging from transparent to highly reflective laminae. Deciphering the correlation of these Precambrian reflection patterns with tectonic processes is an area of forefront research.

The seismic reflection data of BABEL line 2 in the Gulf of Bothnia shows a strongly reflective lower crust and Moho throughout (BABEL Working Group, 1990; Fig. 2-4a). Compared to the lower crust of extensional regimes (see below), reflector lengths seem to be shorter, and reflectivity continues deeper. The termination of reflectivity at the Moho (at 40-50 km depth) agrees with the depth of refraction/wide-angle reflection Moho events. A Proterozoic compressional belt (suture zone) is interpreted in the middle of the profile (Fig. 2-4a), apparently representing an early version of plate tectonics (BABEL Working Group, 1990).

These recent Precambrian profiles are exciting new additions to our seismic constraints on crustal structure and evolution. Most Precambrian data collected to date are from Proterozoic terranes, and an outstanding question is whether Archean profiles will show similar reflectivity patterns, as suggested by Jackson et al. (1990), or will show a consistently thinner crust and some distinctive reflectivity patterns, as might be expected on the basis of seismic refraction data (Durrheim and Mooney, 1991).

3.3. Ancient orogens

We use the term ancient orogen to refer to Proterozoic to Mesozoic orogens, and therefore, there is some overlap with the previous category "Precambrian crust", for example, the Early Proterozoic orogen within the Baltic shield (Fig. 2-4a). In that example, and many other ancient orogens, it is significant that the evidence of a suture zone is still preserved in the lower crust and Moho. While the surface topographic expression of the orogen has long since been eroded, a remnant of the crustal root is still present. Apparently, lower crustal creep has not eliminated the crustal root, as is sometimes the case elsewhere (see below).

A distinct dipping reflectivity pattern is seen in the upper 8 s of the crust beneath the Grenville Front in Canada (Behrendt et al., 1988; Green et al., 1988, 1989), a late Proterozoic compressional belt. Here, the reflection elements are slightly longer and stronger, especially in the upper and middle crust where evidence for thrusting is observed (Fig. 2-4b). Lower crustal reflectivity, aside from the dipping pattern, is less well defined but appears to be gently dipping to sub-horizontal, like much of the lower crustal reflectivity beneath the other Precambrian areas (Figs. 2-3 and 2-4a).

Reflection profiles for Phanerozoic orogenic belts show significant differences in comparison with Proterozoic belts. The Caledonian and Variscan orogenic belts of Europe have apparently lost their crustal roots and show well developed, predominantly subhorizontal (rather than dipping) lamellae in the lower crust. These are features that appear to be due to post-orogenic crustal extension. Despite this overprinting by extensional processes, some structures remain that are presumably associated with Variscan crustal shortening. A prominent example is the Iapetus suture that is evidenced in the lower crust by a wedge-like reflectivity pattern (Fig. 2-5; Klemperer and Matthews, 1987; Freeman and Klemperer, 1988; Klemperer, 1988).

Fig. 2-5. Marine seismic reflection data recorded west of Ireland across the Iapetus suture, which is the result of the closure of the Iapetus Sea in the early Paleozoic Caledonian orogeny (Klemperer, 1989; Matthews and the BIRPS Group, 1990). The suture is marked by a series of prominent northward-dipping lower crustal reflectors that are truncated at the Moho (10 s TWT). The crustal root that presumably was formed during the orogeny (c.f., Fig. 2-4a) has been eliminated by post-orogenic extension and lower crustal ductile flow. We infer, therefore, that the observed reflectivity is multi-genetic in origin, and is due to processes associated with both lithospheric collision and later extension.

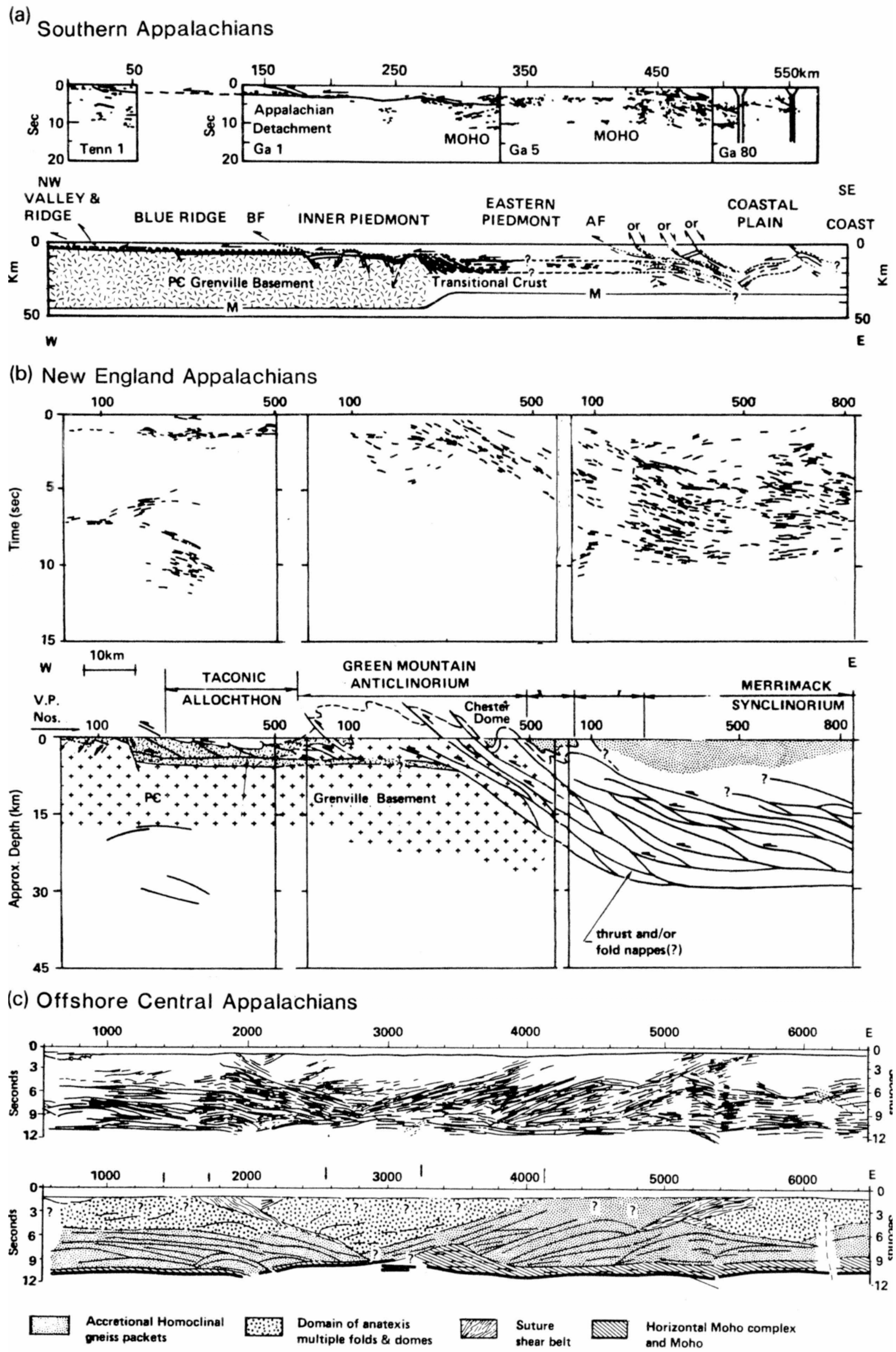

Fig. 2-6. Three pairs of seismic line drawings and interpretive crustal sections for the Appalachians of eastern North America. Information is limited regarding the lower crust in the top (a) and middle (b) panels due to sparse reflectivity within Grenville basement.

(a) Southern Appalachians profile showing thin-skinned tectonics in the upper crust, and dipping reflections from shear zones beneath the Coastal Plain where multiple thrusts are interpreted (Cook et al., 1981). Grenville basement appears to be seismically transparent in these data.

(b) New England, USA, Appalachians showng high reflectivity throughout the lower crust beneath the Merrimack Synclinorium (Brown et al., 1983). This reflectivity rises to the west and ramps over the Grenville basement beneath the Green Mountain Anticlinorium. Sporadic lower crustal reflectivity can be identified within Grenville basement, but coherent Moho reflections are not visible as are seen beneath the Grenville Province in the GLIMPCE data from Lake Huron (Fig. 2-4b) or beneath the Superior Province (Fig. 2-3b).

(c) Highly reflective lower crust is recorded beneath the offshore eastern Appalachians where superimposed orogenic shortening and later extension has deformed the lower crust and strongly enhanced reflectivity (Phinney, 1986).

The Paleozoic Appalachian orogen of North America has received much attention from reflection seismology since the first surveys reported evidence for thin-skinned megathrust tectonics (Clark et al., 1978; Cook et al., 1981). Later studies in the western Appalachians have confirmed this model and have traced a master decollement separating autochthonous Grenville basement from overlying allochthonous rocks of the Appalachian orogen (Fig. 2-6; Iverson and Smithson, 1983; Brown et al., 1986; Stewart et at, 1986; Spencer et at, 1989). In all of these interpretations the lower crust (Grenville basement) is seismically transparent and has, to date, been interpreted as relatively undeformed by the collision. In contrast, the lower crust of the eastern Appalachians (east end of Fig. 2-6a, b) is highly reflective due to extensive deformation and shearing.

The lower crust of the eastern Appalachians, which now underlies the eastern coastal plain and shelf (passive margin) of North America, was moderately extended during the Mesozoic- Tertiary opening of the Atlantic Ocean, and is highly reflective (Fig. 2-6c). The lower crustal reflectivity patterns of the eastern Appalachians have been interpreted to be the result of a tectonic cycle of superimposed shortening and extension (Fig. 2-6c; Phinney, 1986; Hutchinson et al., 1986, 1988; Hall et al., 1989, 1990; Marillier, 1989a,b). The superposition of successive, contrasting tectonic events is a common feature of seismic reflectivity patterns and presents a challenge to the interpreter of these patterns.

The seismic reflectivity patterns of the European North Variscan Deformation Front provide evidence for thin-skinned tectonics that is similar to the Appalachians (Matte and Hirn, 1988; Fig. 2- 7). Within the Variscan terranes there are abundant seismic wedges distributed over hundreds of kilometers. However, nearly all seismic profiles show a nearly flat Moho, indicating that the former crustal roots of these orogenic belts are "eroded", and the present-day Moho is apparently a young feature (Meissner et al., 1987; see Recently Extended Crust, Section 3.5). This flat Moho geometry is indicative of lower crustal creep during the post-orogenic development of these shortened orogenic belts. We assume that this crustal creep occurred during the long period of post-orogenic wrench tectonics and extension that affected the Variscan and Caledonian area (Ziegler, 1981, 1986; Bois et al., 1989).

3.4. Young (post-Mesozoic) orogens

Young orogenic belts have crustal roots that are approximately in isostatic

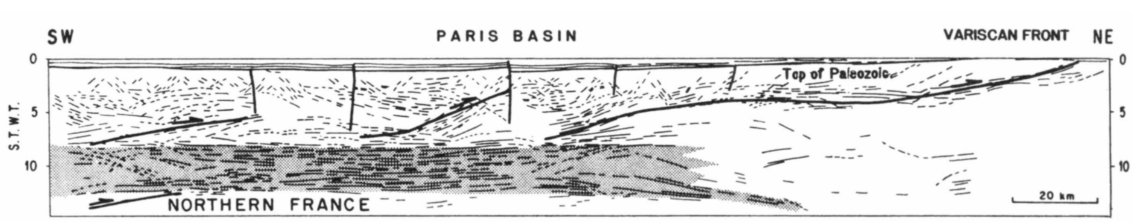

Fig. 2-7. Line drawing of seismic reflection data across the Paris Basin and Variscan Front in France (Bois et al., 1989) showing thin-skin tectonics similar to the Appalachians (Fig. 2-6), but with abundant sub-horizontal reflectivity in the lower crust below the Paris Basin. The Moho is nearly flat and a pre-existing crustal root has been eliminated by post-orogenic extension, as is also seen across the Iapetus Suture (Fig. 2-5) and numerous other regions that have experienced post-collisional extension. In contrast, a crustal root is identified beneath the younger orogens of the Alps and Pyrenees (Fig. 2-8), and the much older Baltic Shield (Fig. 2-43).

equilibrium with their high topography. In general the crustal root is 6-8 times thicker than the mean excess elevation (e.g., Woollard, 1962; Simpson et at, 1986). The crustal root beneath the Alps is about 20 km thick, and that of the Himalayas about 30-40 km thick.

A representative example of the seismic structure of recent orogens is provided by the Swiss Eastern Alps Traverse (Fig. 2-8). Here, the crustal root, as previously determined by a large number of wide-angle/refraction studies, is also imaged in the seismic reflection section. The lower crust dips toward the center of the root from both sides and shows seismic laminations terminating approximately at the refraction Moho. The reflectivity pattern of the lower crust may be the result of deformation along multiple shear zones. The pronounced crustal shortening suggests that at least some lower crustal material might be delaminated beneath the deepest part of the root (England and Houseman, 1989). The upper and middle crust, on the other hand, show a more complex reflectivity pattern that is suggestive of crustal wedging, with nappe displacements above these wedges (Pfiffner et al., 1988; Schweizerische Arbeitsgruppe fur Reflexionsseismik, 1988).

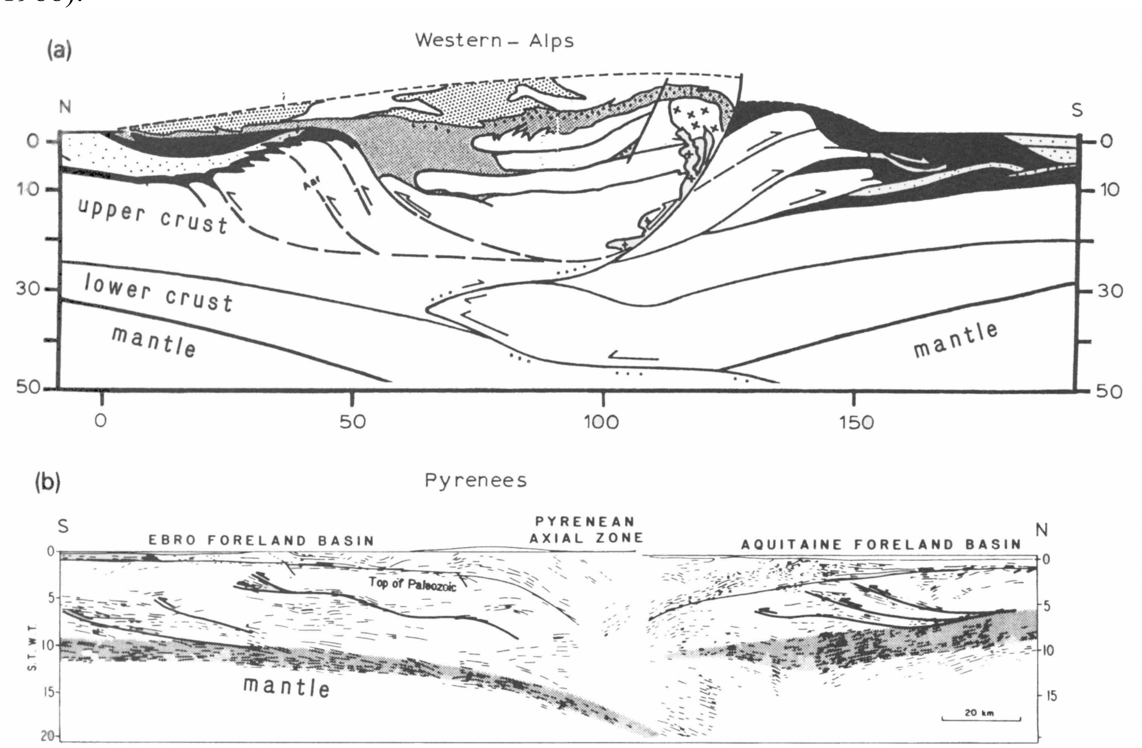

Fig. 2-8. Crustal structure of two young orogens from seismic reflection data.
(A) The Eastern Swiss Alps have a prominent crustal root as evidenced by lower crustal reflectivity (not shown) that dips strongly toward the center of the root (Pfiffner et al., 1988). Significantly, the smaller calculated volume of material within the crustal root as compared with the estimated amount of crustal shortening implies that some crustal material has been delaminated beneath the orogen.
(B) Line drawing of seismic reflection data across the Pyrenees showing northward lithospheric convergence at a moderate dip (Choukroune and the ECORS Team, 1989). Reflectivity can be followed to greater than 20 s TWT (more than 60 km), which suggests that lower crustal delamination also occurred beneath this orogen.

A deep seismic profile across the Pyrenees imaged, for the first time, the entire crust of this orogen, and provided the basis of a palinspastic restoration of the collision zone (Fig. 2-8b; ECORS pyrenees Team, 1988; Choukroune and the ECORS Team, 1989; Roure et al., 1989). The seismic data indicate about 100 km of north-south shortening between the Iberian and European plates and confirm a Moho offset (deeper to the south) of 15 km. The crust within the axial zone of the collision belt has been thickened by at least 15 km by thick-skinned, tacking of whole-crustal flakes (Fig. 2-8b). This profile is significant for its confirmation of both a large Moho offset and the thick-skinned deformation of the Iberian crust. Whereas thin-skinned tectonics is commonly observed in the Paleozoic orogens discussed above, there is excellent evidence for thick-skinned crustal shortening and crustal thickening for these late-Cenozoic orogenic belts. Overall, the style of deformation is very similar to the examples from the Precambrian (Section 3.2; Figs. 2-3 and 2-4).

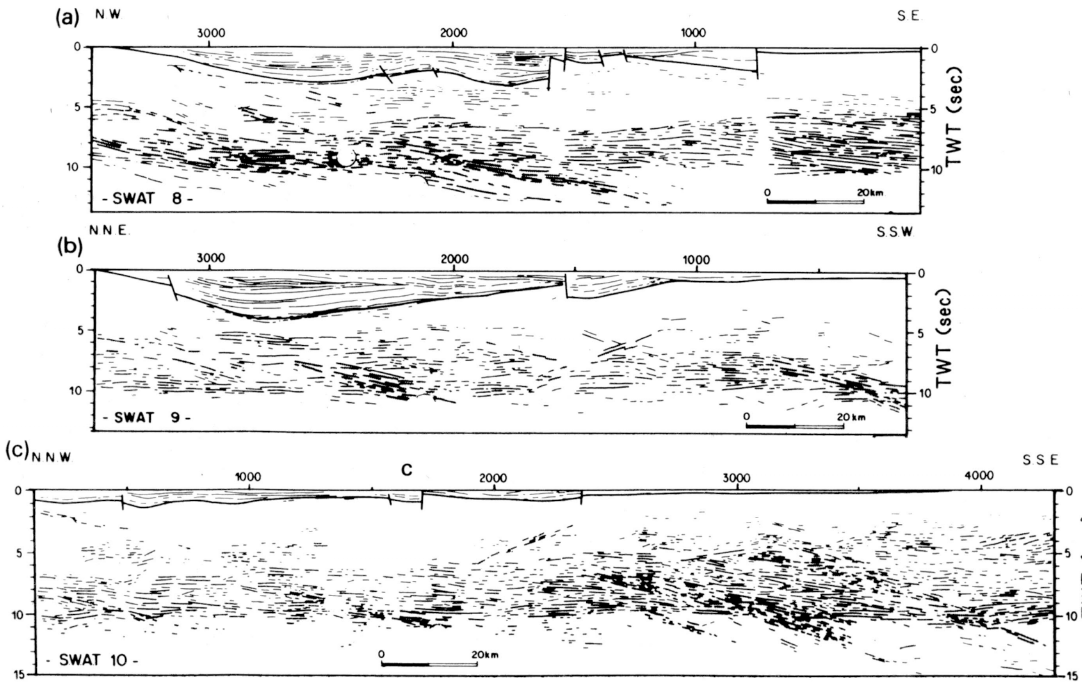

Fig. 2-9. line drawings of three offshore seismic reflection profiles over the extended continental crust between England and France (SWAT: Southwest Approaches Transect; Le Gall, 1990). These profiles illustrate the features that characterize seismic reflection profiles in extended crust: significant seismic reflectivity concentrated within the lower crust (and in supracrustal sedimentary basins); a predominantly sub-horizontal reflector geometry and a clear reflection Moho; some dipping upper mantle reflections. The reflectivity of the lower crust is enhanced by ductile stretching during extension (see section 2.5).

3.5. Recently extended crust

Perhaps no lower crustal seismic reflection data are more remarkable than the profiles recorded in areas where the last tectonic event was extensional, such as much of western Europe, the Basin and Range province of the United States, and many passive margins. In these areas, the lower crust (below a depth of about 10-14 km) is consistently highly reflective, in sharp contrast to the relatively transparent upper crust and upper mantle (e.g., Fig. 2-1). As mentioned above, the reflection Moho is usually nearly flat, because the extensional process eliminates crustal roots below pre-extensional orogenies. Upper crustal faults, where they can be traced into the lower crust, are listric at depth and merge into the reflective lower crust.

Some of the best examples of laminated lower crust and the other reflection patterns described above are from the extended crust around the British Isles (Matthews, 1986; BIRPS and ECORS, 1986; Le Gall, 1990; Fig. 2-9). The reflectivity pattern is variable, sometimes filling the lower crust nearly uniformly, and other times concentrating into two or more narrow zones, or forming a single narrow zone of reflections close to the Moho (Blundell, 1990). The Moho is remarkably flat, and there are several instances of upper mantle reflections, both dipping and flat-lying (Warner and McGeary, 1987).

The ECORS Nord de la France line (Fig. 2-7; Cazes et al., 1986) and DEKORP's lines in the Black Forest (Fig. 2-1; Lueschen et al., 1987) also show a pronounced lower crustal lamination. The four lines in the Black Forest are especially interesting because this region lies on the shoulders of the Tertiary Rhinegraben and has experienced recent uplift but little crustal extension. We speculate that both the lower crustal laminae and the uplift of the Rhinegraben were caused by minor extension, together with intrusion of magmas (from the mantle) into the lower crust. Hence, the laminae seem to represent a two-phase process: first, the intrusion of mafic magma and, second, an ordering process (lateral shearing) which makes reflectors near-horizontal and therefore readily observable on seismic reflection profiles. This process is discussed further in Section 5, below.

The Basin and Range province of the western United States underwent Cenozoic extension that generated listric and high-angle normal faults that are imposed on a crust that was shortened during a Mesozoic orogenic event. Similar to the examples cited above, this extended crust shows a reflective lower crust and a remarkably planar

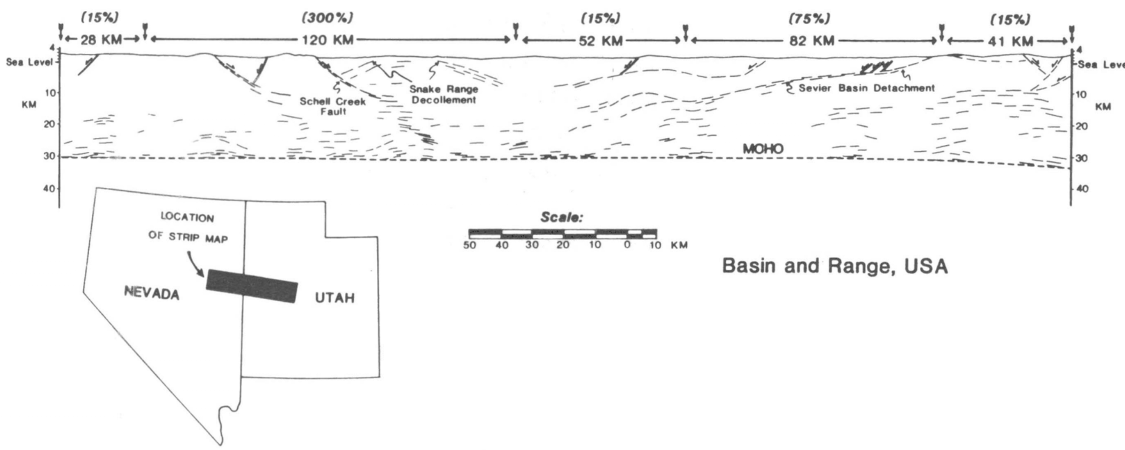

Fig. 2-10. Line drawing of seismic reflection data from the Basin and Range province of the western USA illustrating a flat Moho over a distance of 323 km, and beneath terrains with crustal extension estimated to vary from 15% to 300% (Gans, 1987). The important implication is that lower crustal flow has compensated for the laterally varying amounts of extension, with ductile flow into the areas with the largest amount of extension. The seismic sections from the Basin and Range show high lower crustal reflectivity and are similar to the data illustrated in Figures 1, 9 and 11, which were also recorded over extended crust.

reflection Moho (Fig. 2-10; Klemperer et al., 1986). It appears almost certain that the Moho in the Basin and Range, and other recently-extended crust, is a young feature that formed during extension. The processes forming the Moho would include the ductile stretching of recent igneous intrusions and the metamorphism and deformation of lower crustal rocks. Basin bounding normal faults can not be traced through the lower crust and do not appear to offset the Moho. A possible exception is a profile in Arizona which has been interpreted as showing a Moho offset (Hauser et al., 1987), although this interpretation has been disputed (Goodwin and Thompson, 1988).

3.6. The Moho

The Moho is defined as the boundary (or transition zone) where the compressional wave seismic velocity, as measured by seismic refraction data, increases to greater than or equal to 7.6 km/s (Steinhart, 1967). Sub-Moho seismic velocities of 8.0 ± 0.2 km/s are typical, and the upper mantle is generally interpreted to have an ultramafic bulk composition (cf., references in Holbrook et al., this volume, but see also Snyder and Flack, 1990, for a recent discussion of eclogite in the upper mantle). Seismic reflection observations over the past fifteen years have resulted in a major reevaluation of the nature of the Moho (cf. Jarchow and Thompson, 1989). Two observations were particularly critical in this reevaluation.

First, the Moho was expected to provide one of the strongest near-vertical reflections (after correcting for wave spreading losses) because seismic refraction models commonly show large seismic velocity steps (0.5-1.5 km/s) at the Moho. In practice, the reflection Moho has proved to be a highly variable and, at times, elusive feature. Mid-crustal reflections are often of higher amplitude than the Moho, and, as a result, the reflection Moho has come to be defined as the deepest set of reflections before the nearly-complete die out of reflectivity at the top of the upper mantle. In addition, since seismic refraction data established the Moho as a universal feature of the lithosphere, it was expected that Moho reflections would be observed nearly continuously across seismic reflection sections. In fact, the reflection Moho is generally a 1-2-s-wide zone of reflectivity that can only be traced laterally in a piecewise fashion, and less often as a very narrow, nearly continuous horizon.

Second, many crustal models derived from seismic refraction and gravity data showed considerable relief on the Moho across major physiographic features, and it was therefore expected that seismic reflection profiles would confirm and perhaps sharpen the definition of this high Moho relief. Unexpectedly, many reflection profiles showed the Moho to be nearly horizontal over large distances, despite the presence of significant togography and/or structure at the surface. Examples include profiles across either rugged topography and/or sedimentary basins, such as the Basin and Range, USA (Fig. 2-10), Variscan Europe (Fig. 2-10), and the continental shelf of the British Isles (Fig. 2-11). Profiles that confirmed high relief, and even offsets and/or overlapping of the Moho (double Mohos) include the Alps and pyrenees profiles (Fig. 2-8).

Several new insights have been derived from these observations, some of which have been alluded to in the previous sections. The characteristic of the reflection Moho as a gradual die out in reflectivity that can usually only be followed in a piecewise continuous manner is best explained by a model of the crust in which the Moho is a 3-5-km-wide transition zone consisting, like the lower crust, of alternating thin, random high-

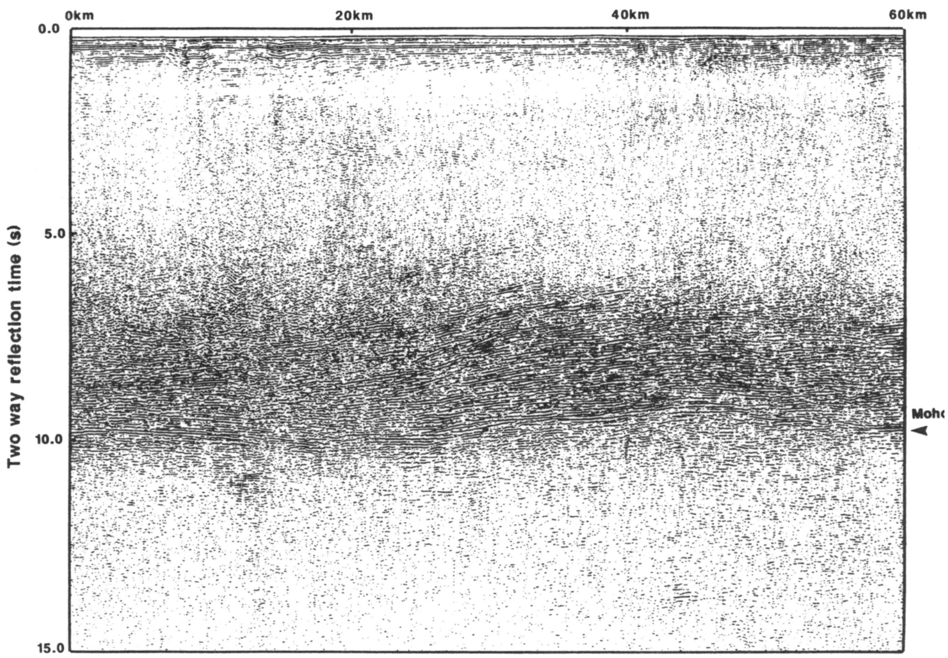

Fig. 2-11. Seismic reflection profile recorded in shallow water offshore of Britain showing a highly reflective lower crust and relatively transparent upper crust and mantle (Warner, 1990). This remarkable seismic section illustrates the 10 km-and-longer continuity of individual reflections, as well as the strength of the reflections (spherical divergence and trace equalization have been applied). Similar seismic sections have been obtained in other regions of extended continental crust. The origin of this reflectivity is discussed in sections 2.4 and 2.5, and in Figs. 2-12, 2-14 and 2-15.

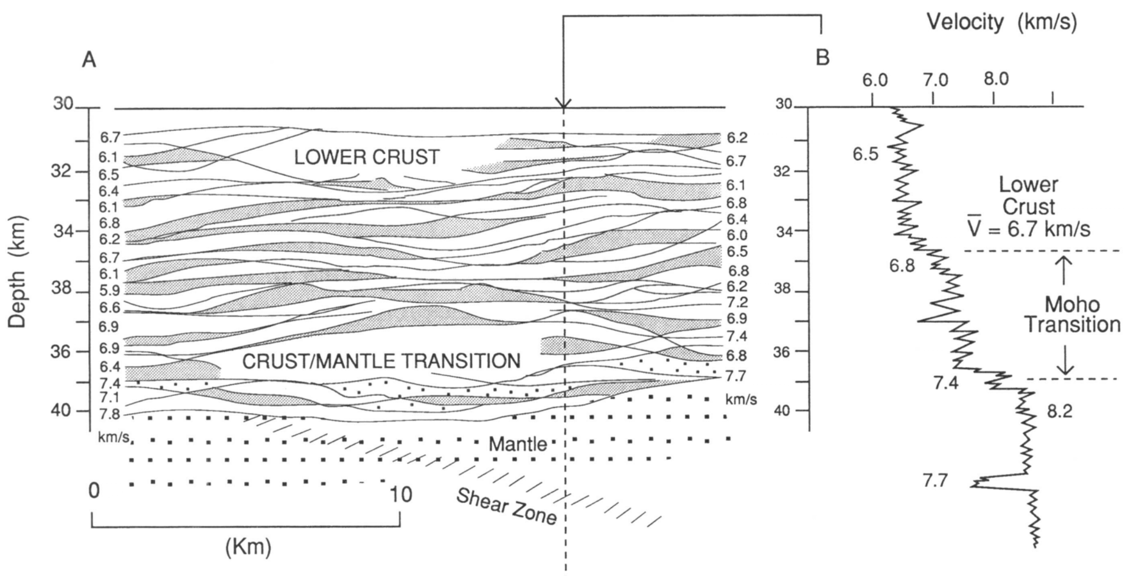

Fig. 2-12. (A) Model for the lower crust and Moho that consists of random, anastomosing, high- and low-velocity lamellae. These lamellae (or laminae) are variously estimated to be 400 ± 300 m thick and show random velocity variations of ±0.05-0.5 km/s with respect to the mean lower crustal velocity (e.g., Wenzel et al., 1987; Sandmeier et al., 1987). Seismic lamellae most probably have their primary origin in lithologic and metamorphic layering, including that caused by igneous intrusions and shear zones. In many cases, the reflectivity of lamellae (e.g., Figs. 2-1 and 2-11) are enhanced by ductile stretching of the lower crust (section 2.5).
(B) Sketch of a representative one-dimensional velocity-depth profile through model (A).

and low-velocity layers (seismic laminae; Fig. 2-12). These laminae are anastomosing rather than continuous and flat-lying, with the result that there is both constructive and destructive interference that increases and decreases reflection amplitudes (Braile and Chang, 1986). Since the entire lower crust, including the Moho, is laminated with thin, random velocity layers, the Moho layer may not be the highest amplitude reflection, and often can only be identified by the transition to the transparent upper mantle.

The flat Moho reflection geometry was unexpected in several areas where previous seismic reflection results indicated significant Moho relief. As discussed above, the most likely explanation for the lack of relief is that the Moho has been reformed by igneous intrusions and ductile deformation in the lower crust (Klemperer et al., 1986). This explanation implies that the Moho transition zone is often the youngest "layer" within the continental crust, a concept that is contrary to conventional stratigraphic succession (younger over older).

Where older seismic refraction interpretations were incompatible with modern seismic reflection observations, newer coincident seismic refraction data confirm that the Moho is in fact measured at the same depth on the two data sets (e.g., Mooney and Brocher, 1987; Holbrook, 1990; Catchings and Mooney, 1991). In all cases where modern data are available, the seismic refraction and reflection Mohos appear to correspond in depth to within the resolution of the techniques. However, a few discrepancies remain. The Moho depth beneath the Colorado Plateau, U.S., is reported as 42 km from older seismic reflection data and as 50 km from modern seismic reflection data (Hauser and Lundy, 1989). Modern seismic refraction measurements could clarify this apparent discrepancy.

4. Multi-genetic origin of lower crustal reflections

The examples described in the previous sections document the diversity in lower crustal reflectivity patterns for different tectonic settings and age provinces. This diversity is indicative of a similar variability in the origin of crustal reflectivity (Smithson, 1986, 1989). In this section we discuss the major hypotheses for the origin of lower crustal reflectivity in the context of these and other examples, and in the context of laboratory measurements of lower crustal properties. We separate this discussion into instances for which there exists direct borehole or outcrop evidence for the origin of seismic reflectors and those instances that lack this direct evidence.

4.1. Origin of reflectivity: concepts with direct borehole or outcrop evidence

In several instances, the origin of upper crustal reflectivity has been established either by drilling into a reflection horizon or by following it to outcrop. Since these upper crustal examples are also relevant to the lower crust, we discuss these well-determined cases first.

Reflectivity caused by igneous intrusives has been documented by drill-hole and field mapping studies. A seismic reflection profile across the Siljan meteorite impact structure, Sweden, revealed a series of discrete, near-horizontal, upper crustal reflections (Juhlin and Pedersen, 1987). Subsequent deep drilling established that these reflections were due to dolerite sheets (V_p = 6.6 km/s) within granitic country rock (V_p = 6.0 km/s)

(Juhlin, 1988). The individual sheets ranged in thickness from about 4 to 60 m. Similar thin igneous sheets are possible in the lower crust and would cause high reflectivity.

The reflectivity of a high-grade metamorphic terrane in the southern Appalachians, U.S. (Coruh et al., 1987) was investigated with laboratory measurements on cores from a bore hole (Christensen 1989a). The lithologies in the study area are significant in that they were once part of the middle crust and are typical of many high-grade metamorphic terranes throughout the world. Computer modelling of compressional-wave velocity and density data determined in the laboratory demonstrate that fine-scale layering of these metamorphic rocks can produce reflection amplitudes typical of lower crustal events. The results reported by Christensen (1989a) emphasize three additional observations: (1) anisotropy in metamorphic rocks can decrease vertical-incidence impedance contrasts at lithologic contacts and hence, reflection amplitudes; (2) lower crustal seismic bright spots, which are commonly attributed to high pore pressure or partial melting (e.g., Brown et al, 1980, 1986), can also be induced by fine tuning of thin metamorphic layers (Fig. 2-13); and (3) the seismic properties of laminated, thin layers within geologic units may be more important for the generation of deep crustal near-vertical reflections than the contacts between major mapped geologic units. The latter conclusion is contrary to the common notion that seismic reflection data provide an approximate cross section of deep-crustal geology.

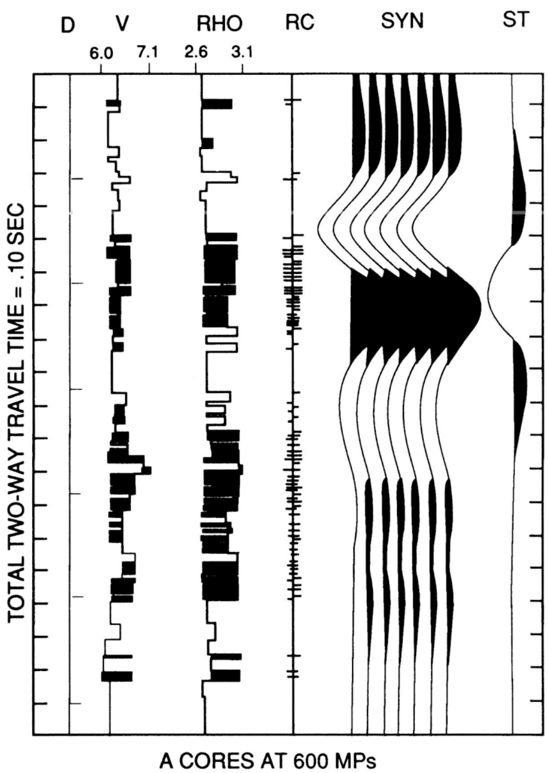

Fig. 2-13. Reflection originating from a zone of alternating high- and low-velocities that demonstrates that multiple thin layers can produce significant reflections (Christensen, 1989). These thin layers may have multiple causes, such as layering of lithology, metamorphic grade, or seismic anisotropy. Seismic "bright spots" need not be due to thick crustal layers. Key: *D* = depth scale in 50 meter intervals; *V* = compressional-wave velocity in km/s; *RHO* = density in gm/cc; *RC* = reflection coefficient; *SYN* = synthetic seismogram; *ST* = standard wavelet corresponding to a reflection coefficient of 0.1, is shown for amplitude comparison.

In a few instances, seismic reflections from fault zones have been traced to outcrops. The high reflectivity observed for some deeply penetrating fault zones (e.g., Smithson et al., 1979; Brewer et al., 1983) can be caused either by the juxtaposition of different rock types across the fault or by the physical characteristics within the fault zone itself. The latter, crustal reflections from mylonitic shear zones, have been investigated by several workers. Laboratory measurements and computer modelling demonstrate that preferred mineral orientations and the presence of retrograde mineral assemblages can result in lower seismic velocities (for propagation directions normal to mylonite foliation within fault zones) (Jones and Nur, 1984; Fountain et al., 1984). The correlation of high upper crustal reflectivity with mylonite zones was established by tracing this reflectivity to the outcrop of mylonites (Hurich et al., 1985; McDonough and Fountain, 1988; Christensen and Szymanski, 1988; Wang et al., 1989; Calvert and Clowes, 1990). The structural and petrologic characteristics of these mylonite zones indicate that they are representative of mid- to lower crustal mylonite zones. These studies demonstrate that fault zone reflectivity does not result simply from reflections from the top and bottom of the fault, but that reflections with the highest amplitudes can originate from within the fault zone. These reflections are caused by internal fault zone structure and a complex interaction of lithologic layering. These layers may often have varying amounts of seismic anisotropy resulting from ductile strain.

4.2 Origin of reflectivity: concepts without direct borehole or outcrop evidence

Upper crustal reflections that are similar in character to those beneath the Siljan structure that were confirmed to be due to dolerite sheets (see above) were recorded in southwestern Arizona, USA (Hauser et al., 1987) and may also be due to igneous sheets that are mapped nearby. Although no drill-hole confirmation is available, these reflectors have been successfully modeled by Goodwin et al. (1989) as thin, high-velocity layers based on the phase (positive polarity) and amplitudes observed on three-component reflection data. The modeling results of Goodwin et al. (1989) are incompatible with an origin of the reflections from water-saturated shear zones, since these would be associated with a negative reflection polarity. These and other similar studies emphasize the importance of igneous intrusions as one origin for crustal reflectivity (Dohr and Meissner, 1975; Oliver et at, 1976; McCarthy and Thompson, 1988; Warner, 1990).

The possible origin of crustal reflections by fluid layers in the deep crust has found support from geologic (e.g., Fyfe et al., 1978) and geoelectrical conductivity data (e.g., Shankland and Ander, 1983). The evidence for deep crustal fluids is summarized by Wickham (this volume) and Jones (this volume). More recently, a number of workers have suggested that lower crustal reflectivity is related to layers with stratified porosity contrasts (Matthews, 1986; Gough, 1986; Jones, 1987; Klemperer, 1987; Hyndman, 1988; Hyndman and Shearer, 1989). Hyndman and Shearer (1989) argue that reasonable physical conditions permit the existence of 1-4% pore water with a salinity close to that of sea-water in the lower crust. They note the common coincidence of high lower-crustal electrical conductivity with high seismic reflectivity, and suggest that this correspondence is indicative of a common origin. In their model, layers with porosity contrasts occur primarily in greenschist and lower amphibolite facies rocks where resistivites of 20-30 Omega m and high reflectivity are observed. They correlate low reflectivity to the absence of pore water; examples include much of the upper crust (where pore water has escaped by flow through fractures), and the lower crust of shields (presumed to be composed of dry granulite facies rocks). Hyndman and Shearer (1989) cite the BIRPS data from the continental shelf around the British Isles as an example of high lower crustal reflectivity that may be associated with high porosity.

An alternate view of the origin of the high lower-crustal reflectivity is expressed by Warner (1990). He summarizes the geologic and geophysical evidence for and against layered lower crustal porosity, and concludes that there are two factors that are difficult to reconcile with a fluid origin for crustal reflectivity: (1) large quantities (2-4%) of pore fluids needed, and (2) long fluid residence times in the lower crust seem unreasonable given the prevailing physical conditions. He concludes that igneous intrusions and underplating of the lower crust with mantle-derived mafic rocks provide a more suitable general explanation for lower crustal reflectivity. Non-fluid causes of high crustal conductivity are summarized in Jones (this volume).

In the model of Hyndman and Shearer (1989), layered lower crustal porosity contrasts result in layers with high and low seismic velocities. However, we note that either high porosity and moderate pore pressure or, in the case of lower (2-4%) porosity, very high pore pressure is needed to produce a significant reduction in seismic velocity in the lower crust (Christensen, 1989b). Geologic evidence would appear to contradict the hypothesis of high porosity within the high-grade metamorphic and igneous rocks that probably are common within the lower crust. Furthermore, very high pore pressures would be difficult to maintain for long times. Therefore, large velocity contrasts associated with layers with stratified porosity contrasts seem physically implausible.

An additional piece of evidence that weighs against abundant stratified porosity contrasts are estimates of the seismic attenuation (inverse *Q*) in the lower crust Qualitatively, the recording of high-frequency (60 Hz and higher) lower crustal and Moho reflections argues for high *Q*. Quantitative *Q* estimates have large error bars, but nearly all indicate lower crustal compressional-wave *Q* values that are moderately high (400-1000; Hwang and Mooney, 1986; Hobbs et at, 1990; Benz et at, 1990). These observations would appear to preclude porosity layers of significant cumulative thickness (greater than several kilometers), assuming that high-porosity layers would have low *Q* values (10-50).

The recent seismic reflection profiles reported from Proterozoic shields provide a test of the hypothesized correlation between high crustal reflectivity and conductivity. While the lower crustal reflectivity of recently-reported shield profiles is comparable to that reported in Phanerozoic areas (Section 3.2), the conductivity of shields is commonly (but not always) several orders of magnitude lower than the regions cited by Hyndman and Shearer (1989). Thus, there does not appear to be a widespread correlation between lower crustal reflectivity and conductivity in Proterozoic crust Furthermore, some of the reflectivity patterns in Proterozoic crust appear to date from ancient collisions; a fluid origin for this reflectivity would require either very long residence times, or fluid recharge along pre-existing shear zones, neither of which seems plausible. Therefore, it is unlikely that lower crustal fluids contribute to the reflectivity of Proterozoic crust

The high lower crustal reflectivity observed in passive margins, in the Basin and Range of the United States, and in Variscan Europe led to the suggestion that reflectivity is related to the process of extension (Phinney and Jurdy, 1979; Matthews and Cheadle, 1986; Smithson, 1986; Allmendinger et al., 1987; McCarthy and Thompson, 1988; Reston, 1988). Although the proposed mechanisms differ in detail, they all incorporate ductile flow in the lower crust to align minerals and stretch lower crustal bodies into a subhorizontal geometry. The resultant structure is reflective due to compositional, metamorphic, and mineralogical layering. The laboratory and borehole data summarized in the previous section support these ideas. The additional possibility that high strain zones are overpressured with fluids (Matthews and Cheadle, 1986) is a special and probably rare case, as discussed above. We expand on the evidence for the enhancement of lower crustal reflectivity by ductile flow in Section 5, below.

In a number of studies, realistic models of the lower crust have been constructed from geologic cross-sections, and velocities and densities have been estimated for these cross-sections based on laboratory and/or borehole measurements (Hale and Thompson, 1982; Fountain et at, 1984; McDonough and Fountain, 1987; Sandmeier et al., 1987; Christensen, 1989a; Fountain and Christensen, 1989). These studies demonstrate that compositional layering by common crustal rocks is an important cause of lower crustal reflectivity.

In summary, several causes of crustal reflectivity have been directly confirmed by drill-hole or outcrop evidence. These causes are igneous intrusions, lithologic and metamorphic layering, and reflectivity from mylonite zones. Additional evidence shows that compositional layering by common rock types contributes to lower crustal reflectivity, and that lower crustal ductile flow enhances reflectivity by aligning minerals and stretching lower crustal bodies into a subhorizontal geometry. For the reasons outlined above, stratified layers with porosity contrasts are likely to be the exception, rather than the rule, as a source of crustal reflectivity.

5. Reflectivity and viscosity in the lower crust

The previous sections presented evidence that lower crustal reflectivity is multi-genetic, and illustrated that the tectonic history of the crust is strongly correlated with the observed reflectivity patterns. In this section we summarize the evidence that ductile deformation in the lower crust acts to enhance lower crustal reflectivity. This ductile deformation may be ancient, as in Precambrian suture zones, or young, as in recently extended crust. We seek to explain the widespread occurrence of a reflective lower crust by considering three lines of evidence for the existence (and nature) of a lower crustal ordering process that enhances crustal reflectivity. The three primary constraints are: (1) the geometry of lower crustal reflections; (2) the evidence for a global stress system; and (3) the rheology of the lower crust. We consider each of these in turn.

5.1. Geometry of lower crustal reflections

Crustal reflectors (or closely-spaced reflector segments) must have a minimum length in order to achieve constructive interference at the surface. This length is about 5 km in the lower crust, provided the interface is planar and not curved. In warm and young lower crusts, mostly rather long, planar, and possibly slightly anastomosing reflectors are present in the lower crust (Section 3.4; Figs. 2-1, 2-9, and 2-11). The widespread, regional appearance of the laminated lower crust and the lack of a consistent correlation with higher seismic velocities (greater than 6.8 km/s) appears to rule out an origin of reflectivity solely from extensive magmatic intrusions. For example, the lower crust of volcanic areas like the Eifel, Germany, do not show a significant increase in seismic lamination, as would be expected if igneous intrusion were the primary cause (DEKORP Res. Group, 1991). Rifted and unrifted areas in the North Sea have the same seismic signatures (Klemperer, 1988). Furthermore, magmatic intrusions should produce many diffractions, as are recorded across the Mid-Continental Rift system (Brown et al., 1982), rather than the planar reflections that are commonly observed elsewhere. While magmatic intrusions certainly play a role in causing lower crustal reflections (cf., section 4.1), away to explain the wide-spread presence of lower crustal lamellae is an ordering process due to regional or global stress system (Zoback et al., 1990). This ordering process would include an ordering of magmatic intrusions.

5.2 Stress-strain evidence for major displacements in the lower crust

Several observations argue for major displacements in a ductile lower crust. The first observation is the existence of large regional stress systems in the upper crust of the continents, as determined by a number of measurements in North America (Zoback and Zoback, 1989), Europe and other parts of the world (Zoback et al., 1990). These large-scale stress systems are related to plate boundary forces that must also affect the lower crust, and can apparently be transferred over distances of more than 1000 km.

The second observation relates to movements of the upper crust relative to the lower crust and subcrustal lithosphere. The mapping of crustal detachment faults that often become listric in the middle or lower crust is among the outstanding achievements of crustal reflection studies (Cook et al, 1981; Allmendinger et al., 1983; Matthews, 1986). The presence of a flat Moho beneath an upper crust that has been extended by an amount that varies locally (ranging from 15 to 300% in the Basin and Range, U.S.A.) implies that brittle upper crustal deformation is decoupled from a more uniform deformation in the middle and lower crust. Another relevant observation is that of tectonic rotation of upper crustal blocks (Matte, 1986; Nur et al., 1989). These block rotations can be modeled by stress concentrations in the rigid upper crust above a weak, low viscosity lower crust (Kusznir, 1982; Kusznir and Park, 1987; Strehlau and Meissner, 1987).

5.3. Rheological control of seismic reflectors

There is a growing body of evidence that the lower continental crust must be a rheologically weak layer between the upper, rigid crust and the sub crustal lithosphere (Sibson, 1982; Meissner and Strehlau, 1982; Kirby and Kronenberg, 1987). Maximum rock strength in the upper crust as calculated from Byerlee's general relationship (Byerlee, 1987) increases with depth (z), but is drastically reduced by the onset of creep, generally at the base of the upper crust between 8 and 15 km depth. Creep may give rise to regional scale flow of material within the lower crust beneath the rigid mid-crust, thereby providing isostatic compensation for domal core complexes (exposed mid-crustal rocks) and maintaining a flat Moho beneath these complexes (Gans, 1987; Block and Royden, 1990). Figure 2-14 shows a calculated viscosity- depth curve for Phanerozoic areas with moderate heat flow. This curve is idealized, and abrupt changes can only be real in the case of an equally abrupt change of rock composition.

The minima in viscosity-depth curves are of particular importance to studies of seismic reflectivity. In the case of young crust these minima coincide with the seismic lamellae of the lower crust, especially for the broad viscosity-minimum of the lowermost crust (Meissner and Kusznir, 1987). Figure 2-14 shows a conceptual picture of this correlation between crustal viscosity and reflectivity. This correlation suggests that lamellae are enhanced by ductile shear in the lower crust (Mooney and Brocher, 1987) have shown that the onset of lower crust lamination does not correlate globally with any specific intracrustal velocity boundaries. Thus, crustal velocities (and by inference, composition) seem to be secondary to crustal viscosity for the formation of seismic lamellae.

While global plate stresses may be important in accounting for the nearly ubiquitous observation of lower crustal reflectivity, stresses at collisional and extensional boundaries are clearly equally important, given that nearly all continental crust has experienced some convergent and/or extensional tectonism at some point. Thus, lower crustal reflectivity may be classified as primary if it is associated with early crust forming processes, or secondary if it is associated with later ductile deformation or metamorphism.

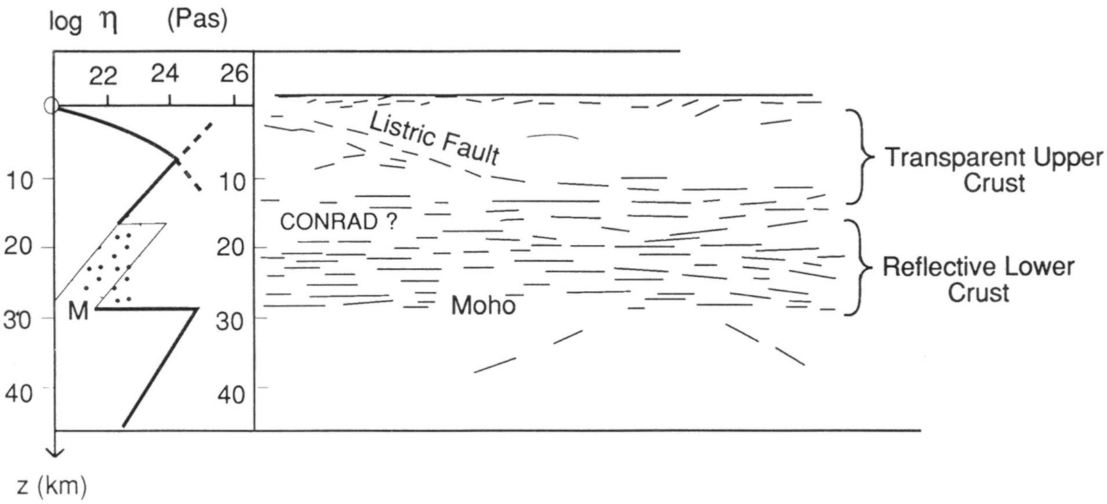

Fig. 2-14. Conceptual picture of the relationship between crustal viscosity and reflectivity that emphasizes the enhancement of lower crustal lamellae (e.g., Figs. 2-1 and 2-11) by ductile shear. Key: *M* = Moho. See discussion in section 2.5.

5.4. Seismically transparent lower crust and mixed reflectivity patterns

We have argued that the global stress system will enhance lower crustal reflectivity nearly everywhere by inducing lower crustal ductile flow that produces a subhorizontal ordering. One may therefore ask why planar seismic lamellae are not present everywhere in the lower crust, that is, what is the origin of seismically transparent lower crust, where it occurs? We first note that crustal stress does not necessarily imply crustal strain. The fact that some Precambrian sutures preserve their reflectivity patterns (Section 3.2) indicates that the crust cooled and has not undergone lower crustal ductile deformation for almost 2 Ga. Much of the observed reflectivity is therefore primary, that is, it is related to deformation and shear that occurred during the suturing of micro-plates to form the Precambrian shields. Second, depending on composition and mineralogy, the lower crust may be more or less likely to form thin metamorphic layers and mylonitic shear zones that enhance reflectivity. Third, either vertically oriented igneous intrusions, or deep faulting could disrupt lamallae after they have formed. A final factor that has already been mentioned (section 2.3) are data acquisition problems that can degrade the quality of deep reflection data, at times to the point that there are few coherent reflectors.

Mixed reflectivity patterns were discussed above (section 3.5) and are indicative of a rheologically stratified crust. Some pre-Cenozoic shortened orogenic belts display both upper crustal wedging and interfingering patterns, and lower crustal lamellae that indicate ductile stretching. Examples include the Basin and Range province (Allmendinger et al., 1983), the eastern Appalachians (Keen et al., 1986; Marillier, 1989a,b; Hall et al., 1989, 1990), and the Variscan crust of western Europe (DEKORP Res. Group, 1990). A Mesozoic age for Variscan lower crustallaminations can be estimated by noting that they have obliterated Paleozoic thrusts and faults at depth, but have themselves been offset by Tertiary deformation (Bois et al., 1989). The determination of the chronology of seismic reflectivity patterns provides a major constraint on the structural evolution of the crust (e.g., Klemperer et al., 1990).

6. Conclusions

Seismic reflection profiles collected over the past 15 years range in tectonic setting from stable Precambrian shields to actively extending areas and young collision zones. These profiles have provided a variety of new insights into the structure and evolution of the continental lower crust and Moho. Here we summarize some of the key insights; the appropriate primary references are cited in the previous sections. We arrange this discussion into three principle topics: the structural development of the lower crust in various tectonic settings, the nature of the Moho, and the origin of lower crustal reflectivity.

Prominent insights regarding the structural development of the crust include the demonstration of the mechanics of crustal shortening in either a thick- or thin-skinned fashion. The southern Appalachians (Fig. 2-6a) and parts of the Variscan front (Fig. 2-7) illustrate thin-skinned thrusting that extends for 100 km and more. Thick-skinned tectonics, which involves deformation of the lower crust, is seen beneath the New England Appalachians (Fig. 2-6b) and is implied by the depth extent of the Iapetus suture (Fig. 2-5). Seismic profiles across the Alps and the Pyrenees (Fig. 2-8) both show significant crustal shortening, with the lower crust deforming and thickening beneath the high topography. Significantly, the lower crust may delaminate beneath some collision belts, thereby creating a less mafic orogenic crust.

Recent profiles within Precambrian crust have identified lower crustal reflectivity patterns that are consistent with modern tectonic processes, including collisional orogenies (Figs. 2-3 and 2-4). The data permit the extension of plate tectonic models into the Early Proterozoic with greater confidence. The fact that ancient reflectivity patterns remain "frozen in" the lower crust attests to the tectonic and thermal stability of the Precambrian crust over the past 1-2 Ga. Expanded deep reflection profiling of Archean crust remains a scientific frontier.

Perhaps the most well known seismic reflection profiles are those recorded over extended continental crust that show striking sub-horizontal lamellae throughout the lower crust (Figs. 2-1, 2-9 and 2-11). Well-studied examples include the crust around the British Isles, the Basin and Range, USA, and much of western Europe. Normal faults generally become listric with depth, and anastomosing shear zones appear to predominate in the lower crust (Fig. 2-11). The stretching of the lower crust has eliminated pre-extensional crustal roots beneath these terranes, and the flat Moho relief, despite laterally varying amounts of extension (15-300%) is indicative of lower crustal flow to equalize crustal thickness.

Mixed reflectivity patterns are also common, and are indicative of successive tectonic episodes that have superposed their structural signatures on the crust. For example, the eastern Appalachians of North American and Variscan Europe preserve evidence for Paleozoic crustal shortening in the upper crust and Mesozoic ductile stretching in the lower crust. The superposition of crustal reflectivity patterns can lead to interpretational ambiguities, but also provides a rich source of information regarding the tectonic history of the crust. The dating of diverse seismic reflectivity patterns, generally by geometrical relationships, is an important goal for seismic reflection studies.

Seismic reflection profiles show great diversity in the structure of the Moho. Recently recorded airgun and explosion source profiles have provided the best images to date of the Moho beneath Precambrian shields. In these profiles the lower crust and Moho are moderately-to-highly reflective, with horizontal to gently dipping reflections predominating, especially away from ancient collision zones (Figs. 2-3 and 2-4). In young orogenic zones, the Moho dips steeply downward from either side of the mountain belt (e.g., the Alps and the Pyrenees; Fig. 2-8), and the Mohos of opposing convergent lithospheric plates are offset, with either a Moho step, or a "double Moho" beneath a limited portion of the mountain belt.

The fine-scale properties of the Moho have been investigated as well, and seismic reflection data indicate that the Moho is commonly a 3-5 km wide transition that, like the lower crust above it, consists of anastomosing thin layers with randomly alternating high and low velocities (Fig. 2-12). Despite the significant velocity increase (0.5-1.5 km/s) from the lower crust to the mantle, the Moho reflection is often not the strongest event in a seismic section because the fine structure of reflective layers is more important than the bulk velocity change. Thus, middle or lower crustal laminations may produce stronger reflections than the crust/mantle transition.

The Moho is a young, newly formed feature in areas of highly extended crust, and thus the principle of stratigraphic succession, younger over older, is violated in the crystalline continental crust and Moho. As mentioned above, the nearly flat geometry of the Moho beneath highly variably-extended crust implies that there is lateral flow in the lower crust to preserve a nearly uniform crustal thickness.

Various hypotheses have been advanced for the origin of lower crustal reflectivity, and we have argued that this reflectivity is multi-genetic, a point that is unambiguously demonstrated by the correlation of reflection profiles with boreholes and by tracing reflection horizons to outcrop. Causes of crustal reflectivity that have been demonstrated by such direct evidence includes igneous intrusions (sills), fine-scaled metamorphic layering, mylonite zones, and lithologic layering.

Recent reflection profiles in Precambrian shields reveal strong crustal reflectivity, including sub-horizontal laminae in the lower crust, thereby upsetting an apparent correlation between lower crustal lamination and crustal extension. Many of the Precambrian reflectivity patterns appear to preserve structures that date from ancient collisional events, which implies that this reflectivity is not due to post-orogenic ductile shear, igneous intrusions, or fluids. Rather, reflectivity is due to primary lithologic and metamorphic layering, and Precambrian shear zones that were formed during the compressional orogenies that formed the Precambrian crust (Fig. 2-15). Similar mechanisms can account for reflective lower crust in young orogens.

In extended crust we favor a model that is depth-dependent, with ductile shear enhancing reflectivity that has its primary origin in lithologic and metamorphic layering in the lower crust, and igneous layering within the 3 – 5-km-thick Moho transition zone (Fig. 2-15). Layered zones of high pore pressure (up to several kilometers thick and due to metamorphic dewatering of the lower crust) may sometimes be present at the top of the lower crust, as indicated by low seismic refraction velocities and high electrical conductivity.

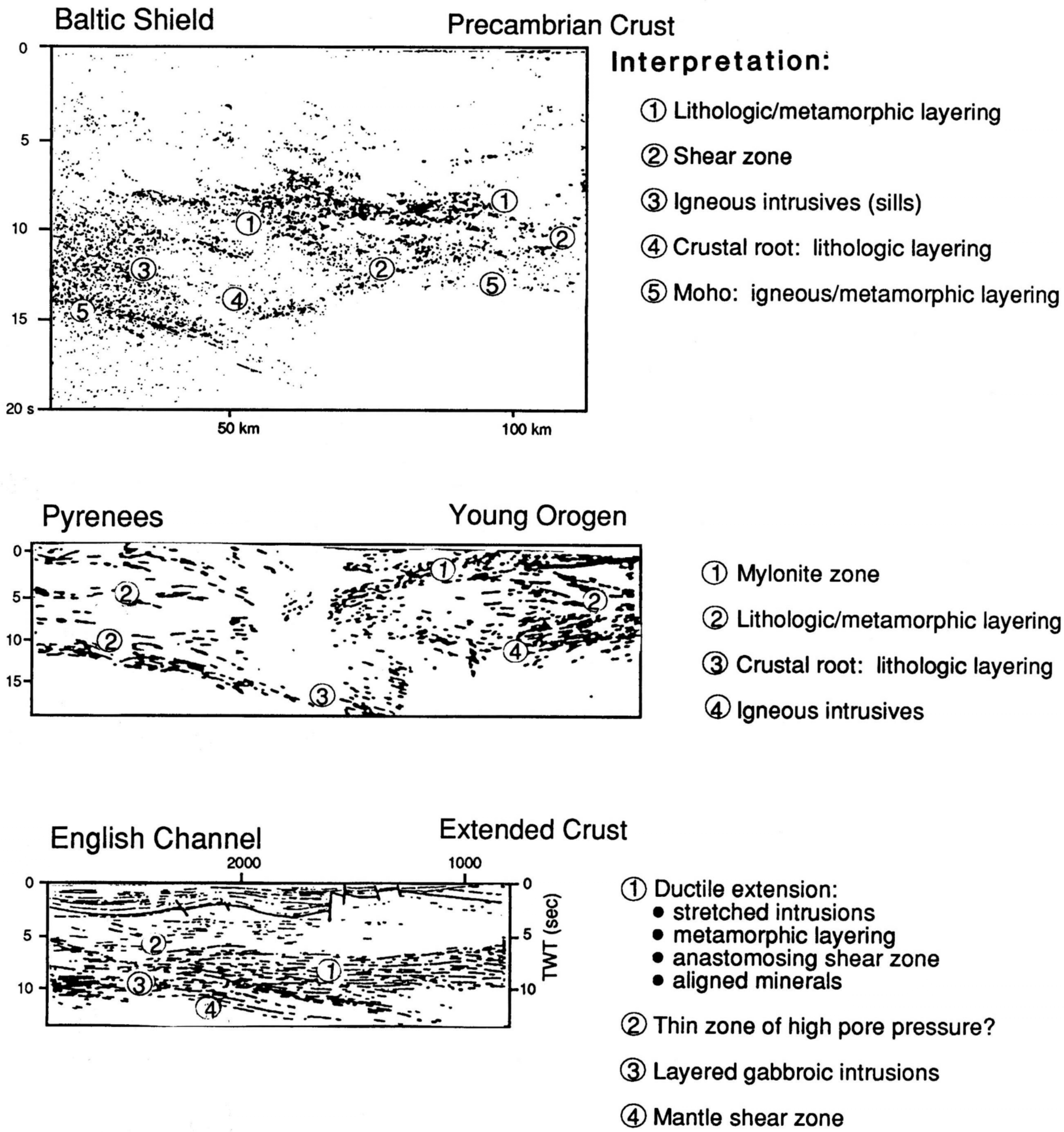

Fig. 2-15. Examples of reflective lower crust from three diverse geologic settings that emphasize the multi-genetic origin of crustal reflectivity. Interpretations of selected reflective zones (numbers within circles) are based on the evidence presented in sections 2.4 and 2.5. However, these interpretations are only working hypotheses, there being no drill-hole confirmation available for these deep-crustal examples. The most common causes of crustal reflectivity are indicated in the top and middle examples (lithologic/metamorphic layering, shear zones, etc.), and we refer to these as “primary” causes; extended crust nearly always shows enhanced lower crustal reflectivity due to ductile flow, which we term a "secondary" cause. See text for discussion.

The general observation of lower crustal reflectivity leads to the suggestion that a common process acts to promote reflectivity. We favor a model whereby regional and global plate stresses act to enhance reflectivity by inducing lower crustal ductile flow that produces subhorizontal lamination. This ordering process would enhance reflectivity that has its primary cause in igneous intrusions, or compositional, metamorphic, or mineralogic layering (Fig. 2- 15). This ordering process requires elevated temperatures to promote ductile flow, and the dipping reflection patterns in Precambrian terranes appear to be primary patterns associated with Proterozoic micro-plate collisions, rather than secondary reflectivity associated with ductile flow. Thus, the Precambrian crust appears to possess considerable long-term thermal and tectonic stability.

Deep seismic reflection exploration of the crystalline crust has yielded an abundance of insights into the continental lower crust and Moho. Future insights will come from expanded geographic coverage, particularly in Proterozoic and Archean crust, the application to shear-wave profiling, and from continued correlation with related geologic/geophysical studies, and borehole and laboratory studies of the physical properties of rocks.

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