J Geophys (1987) 61:1–11

Journal of Geophysics

A relation between continental heat flow and the seismic reflectivity of the lower crust

Simon L. Klemperer

British Institutions' Reflection Profiling Syndicate (BIRPS), Bullard Laboratories, Madingley Rise, Madingley Road, Cambridge, CB3 0EZ, U.K.

Abstract. Deep seismic reflection profiling of the continental crust has found world-wide evidence for prominent reflections from the middle and lower crust. This paper presents evidence for a correlation between the depth to the zone of pronounced reflectivity (the 'reflective lower crust') and the surface heat flow. The highly reflective zone appears to be shallower beneath regions with higher heat flow, suggesting that one condition for the development of the reflective zone is the existence of a sufficiently high temperature in the crust. The data presented in this paper suggest that the highly reflective zone is generally developed only at temperatures higher than 300°–400 °C. This correlation of reflectivity with heat flow implies that crustal reflectivity must be variable on the same time scales on which crustal heat flow is variable. This constraint appears to favour an origin for the lower crustal reflectivity related either to ductile strain banding or to free fluids that may be transient on geologic time scales, rather than to compositional layering or multiple igneous intrusions that are relatively permanent features of the crust.

Key words: Reflection seismology – Deep-crustal reflection profiling – Heat flow – Continental lower crust

Introduction

Among the most fundamental discoveries of crustal reflection profiling has been the finding that the lower part of the continental crust typically exhibits a pronounced reflectivity that is unmatched in the upper crust above or in the upper mantle below. Typical deep seismic sections show reflections from the sedimentary cover, if present, below which lies an upper crust that is rather non-reflective apart from occasional through-going fault reflections, and a strongly reflective middle and lower crust above an upper mantle that is frequently non-reflective (Matthews and Cheadle, 1986). Data fitting this model have been recorded in North America (e.g. Brown et al., 1986; Green et al., 1986), Europe (e.g. Matthews and Cheadle, 1986; Bois et al., 1986; Meissner and Wever, 1986) and Australia (e.g. Moss and Mathur, 1986) and from areas as diverse as active tectonic regions (Basin and Range province, Allmendinger et al., 1987a) and Archaean cratons (Mathur et al., 1977). Numerous hypotheses have been advanced to explain the reflectivity of the lower crust. These hypotheses include multiple intrusion of mafic sills (Meissner, 1973; McKenzie, 1984), layered igneous intrusions (Lynn et al., 1981), sediments thrust beneath the upper crust (Brown et al., 1983), extensional or compressional strain banding developed in the 'ductile' lower crust (Phinney and Jurdy, 1979), laminae of hydrated rocks (Hall, 1986) and open, fluid-filled cracks (Matthews, 1986). Though these hypotheses are not necessarily mutually exclusive, it seems probable that such a widespread phenomenon as the reflectivity of the lower crust has relatively few dominant underlying causes.

There are also some areas of the world where the reflective lower crust is not recognisable as a separate entity because reflective zones are visible throughout the crust. Such areas are frequently in the interior of compressional orogens (e.g. Appalachians: Ando et al., 1983; North American Cordillera: Potter et al., 1987a; Cook et al., 1987) and the reflective zones at high levels in the crust are commonly interpreted as fault zones or highly faulted rocks. These areas are not considered further in this paper, because the reflective lower crust cannot be easily defined on profiles from these areas.

Different approaches to the study of the cause of the reflectivity of the lower crust include both attempts to trace the lower crustal layering to surface outcrop or to drillable subcrop in areas of deep erosion, and also attempts to relate the distribution of reflectivity to geologic history, tectonic age or present-day crustal state. The latter approach is taken in this paper. Crustal reflection profiles from three continents have been selected on which the reflective lower crust can be clearly defined over length scales at least comparable with crustal thickness, and for which the regional surface heat flow is known for the area in which the profile was collected. Deep seismic reflection profiles provide an image of the physical properties of rocks present at the time the profile is collected, and provide a measure of the present-day state of the crust. Similarly, because the thermal time constant of the continental crust is of the order of only 10 Ma (Lachenbruch and Sass, 1977), surface heat flow is a measure of the present-day, or, at least, of the Neogene, state of the crust in a way that surface exposures of rocks dating from former epochs are not.

Previous attempts to analyse the distribution of reflectivity on crustal reflection profiles have included both histogram analyses of the number of reflections observed in successive depth intervals (Meissner et al., 1983; Meissner and Wever, 1986; Wever et al., 1986) and more subjective estimates of reflection amplitudes and patterns (Allmendinger et al., 1987b). In this paper a simpler parameter is mea-

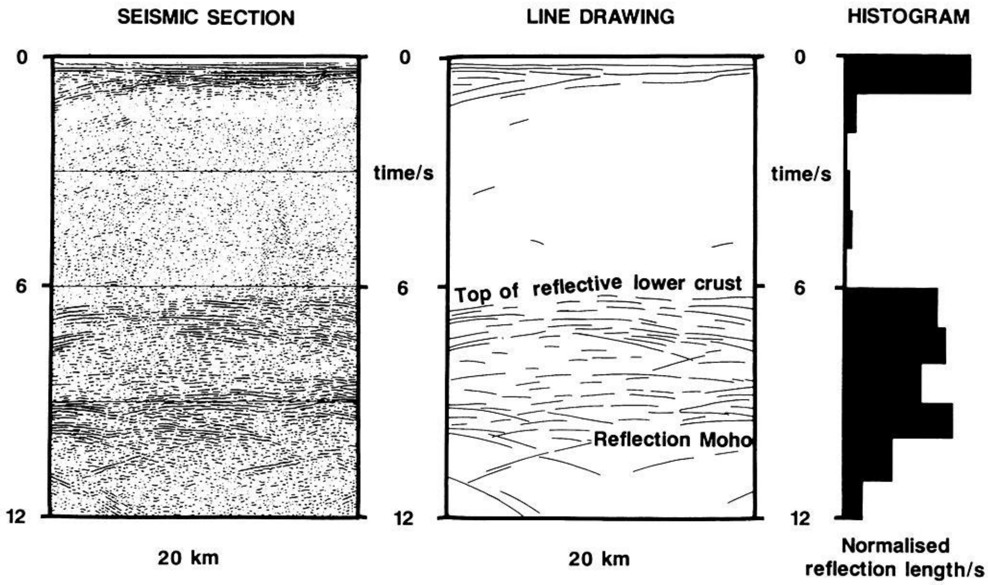

Fig. 1. a Segment of crustal reflection data (BIRPS SWAT 7: BIRPS and ECORS, 1986) to show the 'reflective lower crust' between the poorly reflective upper crust and upper mantle. **b** Line drawing of data in Fig. 1a to show the interpretation of the top of the reflective lower crust and of the reflection Moho. Note the slight travel-time pull-down ($\simeq$0.5 s) of the top of the reflective lower crust beneath the basin on the left of the data segment; the top of the reflective lower crust may be at a constant depth. **c** Histogram of reflection length summed over 1 s travel-time intervals, showing the abrupt increase in reflectivity beneath 6 s

sured; the two-way travel time from the surface to the top of the reflective lower crust. Where the seismic velocity of the crust is known, this travel time may be converted to an absolute depth. It is this depth to the top of the reflective layering for which a correlation with the surface heat flow is presented in this paper.

Presentation of data

Figure 1a shows a sample of deep seismic data recorded southwest of Britain by BIRPS. A dramatic increase in reflectivity occurs at 6.0–6.5 s (all times are given as two-way travel times). Similar jumps in reflection density can be observed on a line drawing of the seismic data (Fig. 1b) and on a histogram showing reflection density as a function of travel time (Fig. 1c). In order to illustrate as much data as possible in this paper line-drawings are used, but references are given to other papers which show the actual data. The data in Fig. 1 and throughout this paper are unmigrated but, because the reflections in the lower crust are typically sub-horizontal, migration does not in general greatly alter the distribution of these reflections with travel time. The travel time of onset of the reflective layering varies from as little as 3.5 s in parts of the Basin and Range Province, Nevada, and beneath the Cornubian batholith, south-west Britain, to as much as 8 s beneath the Colorado Plateau, Utah, the northern Paris Basin, France, and parts of the Australian craton. In the area of Fig. 1 the onset of the reflectivity is sharp and is observed at a rather uniform travel time across a lateral distance of several tens of kilometres. In some areas the distinction between non-reflective upper crust and highly reflective lower crust is less clearly defined, perhaps due to higher noise levels, but may be estimated to within 0.5–1.0 s. The rather larger range of travel times (0.5–2.0 s) given for the data points compiled in Table 1 is accounted for by local, lateral variation in travel time to the top of the reflective lower crust. Lateral variation can be extreme where major faults cut through the crust. In these cases the top of the reflective layering dips systematically over a range of several seconds [e.g. Appalachian suture, Georgia (Nelson et al., 1985)] and may be bounded by a fault-plane reflection [e.g. Outer Isles fault zone (Brewer et al., 1983)]. These cases have not been included in the present compilation because it is meaningless to pick a single travel time to the top of the reflective lower crust when the top is steeply dipping.

Although the observed position of the top of the reflective lower crust may be locally dependent on signal-to-noise ratios which may be expected to vary both laterally and vertically, care has been taken to average the observed depth of the reflective lower crust over distances of several tens of kilometres thus circumventing any local noise problems in these data sets. Indirect evidence that it is possible to pick the top of the reflective lower crust comes from gravity modelling. Successful gravity models along seismic profiles around Britain and in Germany have been prepared using the top of the reflective lower crust as a significant density boundary (Setto and Meissner, 1986), implying that the reflective lower crust is a distinct geological entity.

Western USA

The western USA is an area for which the heat flow is well determined (over 120 heat-flow determinations are incorporated in Fig. 2a), and across which the heat flow shows considerable variability (Fig. 2a) from greater than 100 $mW \cdot m^{-2}$ in the Battle Mountain heat-flow high to less than 60 $mW \cdot m^{-2}$ in the Colorado Plateau and the Sierra Nevada (Lachenbruch and Sass, 1977). It is also an area across which continuous deep reflection profiles have been run from the Sierra Nevada to the Colorado Plateau (Allmendinger et al., 1987a) by COCORP (Consortium for Continental Reflection Profiling). Figure 2b shows parts of line drawings of COCORP reflection data acquired across the Battle Mountain heat-flow high and across the Colorado Plateau, and shows a clear variation in the travel time to the lower crustal reflectivity which ranges from 3.5 to 8 s. It is important to note that the COCORP profiles represent an essentially continuous transect across the Basin and Range province and the western Colorado Plateau. The deep reflective layering is observed across the entire 750-km transect, except in areas where signal penetration was limited to the upper crust only (Mayer and Brown, 1986). This is a powerful argument that the reflective layer-

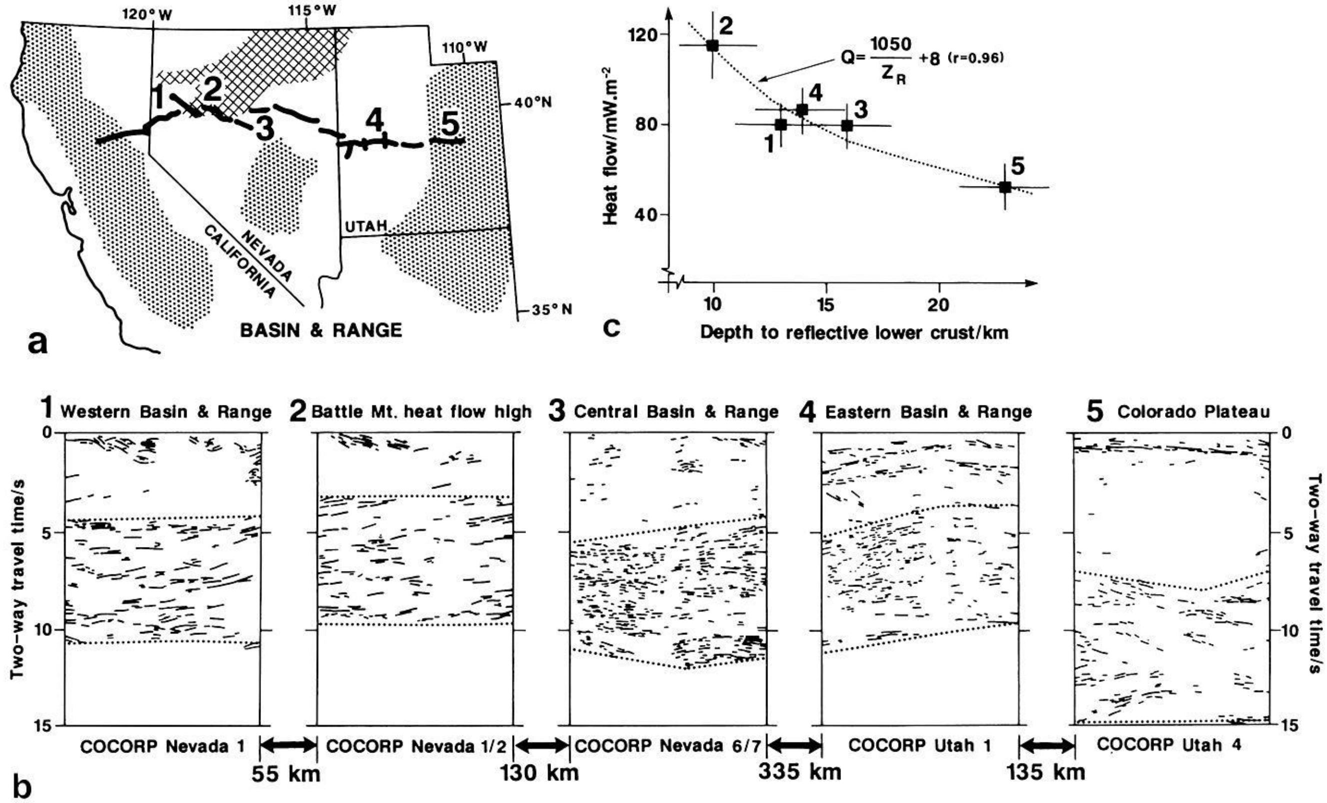

Fig. 2. a Map of western USA to show COCORP lines in relation to heat flow provinces. *Cross-hatching*: heat flow $\geq 105\ \mathrm{mW \cdot m^{-2}}$; *dots*: heat flow $\leq 63\ \mathrm{mW \cdot m^{-2}}$; *blank area*: intermediate heat flow values (Lachenbruch and Sass, 1977). Numbers 1–5 show locations of data segments illustrated in Fig. 2b. **b** Data segments along the COCORP Basin and Range survey. The reflective lower crust has been shown to be essentially continuous between the data segments illustrated. Each data segment is 25 km long, and is located by its number in Fig. 2a. Data redrawn from Hauge et al. (1987), Potter et al. (1987b), Allmendinger et al. (1983) and Mayer and Brown (1986). *Dotted lines* mark the interpreted top of the reflective lower crust and the interpreted reflection Moho. **c** Plot of heat flow vs. depth to the top of the reflective lower crust. Values plotted (see Table 1) are derived from the full dataset located in Fig. 2a rather than the 25-km data segments illustrated in Fig. 2b

ing has a related cause in both the Basin and Range (a Neogene extensional province) and in the Colorado Plateau (where the most recent deformation was Eocene compression) despite their different Cenozoic histories (though of course tectonic events in the lower crust may not have parallelled the observed upper crustal events in character or timing). Figure 2c shows a plot of surface heat flow against depth to the reflective layering for the COCORP transect, with depths being calculated from refraction velocities given by Prodehl (1979). Note that regional estimates averaged over 50–100 km of both heat flow and depth to reflective layering have been used. A simple inverse relationship can be fitted to the data.

Offshore British Isles

The British Institutions' Reflection Profiling Syndicate (BIRPS) has collected over 8000 km of deep reflection data over the continental shelf around Great Britain, and most of these data show a typically transparent upper crust and strongly reflective lower crust (Fig. 1). No measurements of heat flow were made along the lines of the BIRPS profiles, but because there are many heat-flow determinations in the British Isles (over 180 measurements in the area of Fig. 3a) and because the profiles were collected close to shore (within 20–30 km) it is possible to extrapolate generalised heat-flow provinces offshore from England and Wales (Fig. 3a). The extensive BIRPS dataset allows the reflective layering in the lower crust to be traced continuously across all these heat-flow provinces. Unusually high heat flow occurs over the Hercynian Cornubian batholith in southwest Britain (Bloomer et al., 1979), and where the BIRPS profiles cross the batholith the reflective layering is observed to shallow by 3 s to about 3.5 s (BIRPS and ECORS, 1986) (Fig. 3b). A similar shallowing of the reflective layering was observed beneath the Cotentin granites of northern France (BIRPS and ECORS, 1986). Although the unusually shallow reflective layering beneath the Cornubian batholith could be due to lithologic layering related to the granites, the continuity of the reflective lower crustal layer both north and south of the Cornubian batholith argues that the layering has a more general cause that is widespread over hundreds of kilometres. In the Celtic Sea the regional heat flow is less than $60\ \mathrm{mW \cdot m^{-2}}$ (Bloomer et al., 1979) and the top of the reflective layering is generally beneath 5 s (BIRPS and ECORS, 1986) (Fig. 3a and b; see also Fig. 1 from offshore Brittany). Further north in the Irish Channel the heat flow is $60–80\ \mathrm{mW \cdot m^{-2}}$ (Bloomer et al., 1979) and the top of the reflective layering is at about 5 s (Brewer et al., 1983). Travel times were converted to depths using velocity models derived from crossing refraction profiles by Blundell and Parks (1969), Holder and Bott

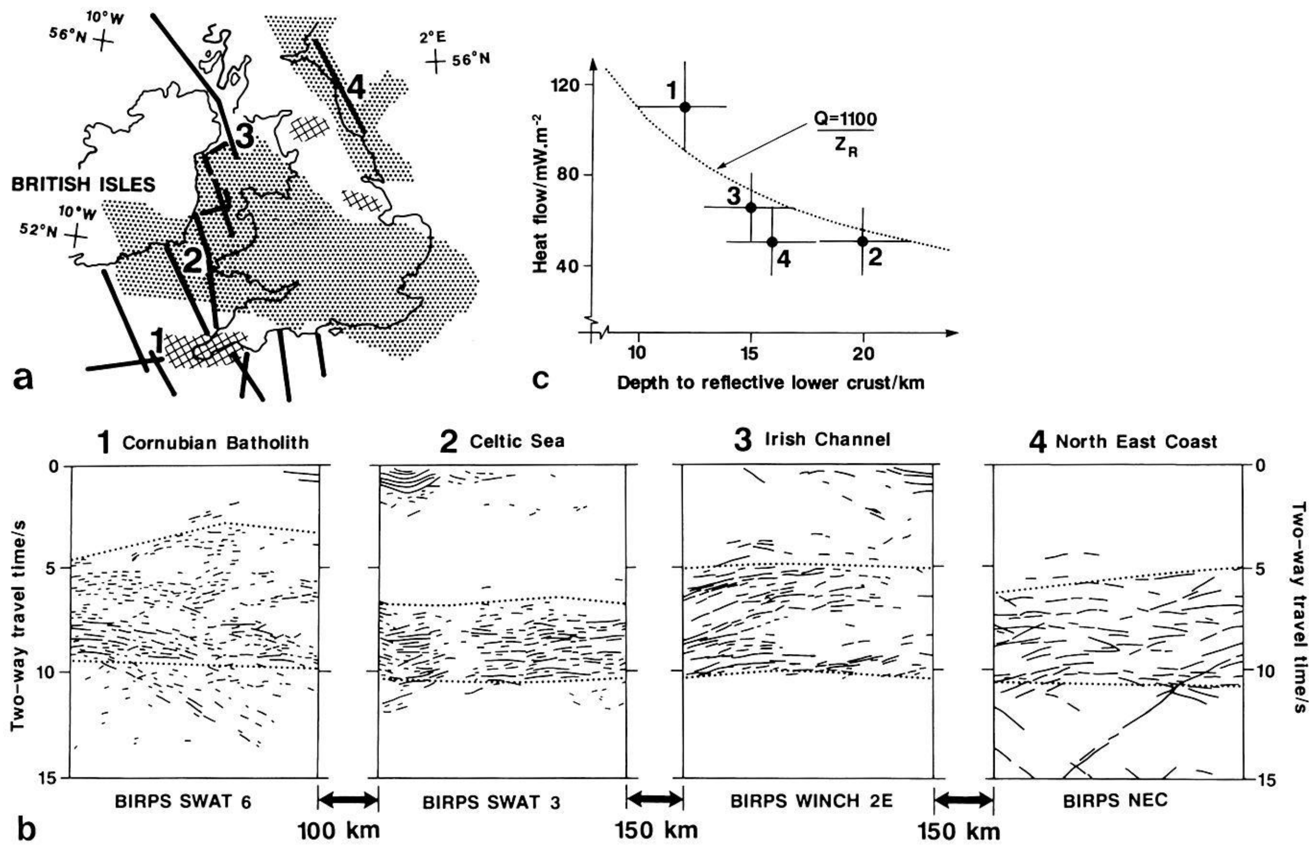

Fig. 3. a Map of British Isles to show BIRPS lines in relation to heat flow provinces. *Cross-hatching*: heat flow 80 mW·m^{-2}; *dots*: heat flow ≤ 60 mW·m^{-2}; *blank area*: intermediate heat flow values (Čermák, 1979; Gale, 1986). Numbers 1–4 show locations of data segments illustrated in Fig. 3b. **b** Data segments from BIRPS surveys around Britain. The reflective lower crust has been shown to be essentially continuous between the data segments shown from the west side of Britain. Each data segment is 30 km long, and is located by its number in Fig. 3a. Data redrawn from BIRPS and ECORS (1986), Brewer et al. (1983) and Klemperer and Matthews (1987). *Dotted lines* mark the interpreted top of the reflective lower crust and the interpreted reflection Moho. **c** Plot of heat flow vs. depth to the top of the reflective lower crust. Values plotted (see Table 1) are derived from the full dataset located in Fig. 3a rather than the 30-km data segments illustrated in Fig. 3b

(1971), Bamford (1972) and Lewis (1986). Figure 3c shows a plot of average depth to the reflective layering for each of these three areas, between which the reflective lower crust may be traced continuously, and also for the western North Sea.

World-wide compilation

Both the western USA and the continental shelf around Great Britain show similar relationships between heat flow and depth to the reflective layering. In order to obtain a larger number of data points it has been necessary to compile all available deep crustal data from the whole world, including data from other areas of North America, Europe and Australia (Table 1 and Fig. 4). The criteria for selection of these areas were that the reflective lower crust be seismically well defined over length scales of at least several tens of kilometres, and that the regional heat flow be known. The presentation of these diverse data sets on a single plot implicitly assumes that the reflective layering has the same cause in all these areas. Although this assumption may well be justified in areas of continuous profiling such as the COCORP or the BIRPS data sets presented above, this assumption cannot be easily tested world-wide. It is possible that there exist different causes for lower crustal reflectivity in different areas, and that this is the reason for the greater scatter in the data in the world-wide compilation (Fig. 4) as compared to the regional datasets (Figs. 2c and 3c). Although there is a considerable spread of data points in Fig. 4, points tend not to plot in the upper right (high heat flow/deep reflective lower crust) nor in the lower left (low heat flow/shallow reflective lower crust) of the graphs in Fig. 4. More data points, in particular from areas of very high heat flow, are needed to adequately constrain any relationship between heat flow and depth to the reflective lower crust. Within this limitation, the world-wide ensemble of data visually fits a single pattern, despite being drawn from wide-ranging geographic locations and highly variable tectonic environments.

Correlation between temperature and depth to reflectivity

In the previous paragraphs regional and global data sets were used to substantiate a correlation between surface heat flow and depth to the seismically reflective lower crust. Statistical correlations have been calculated both for depth to the reflective lower crust proportional to heat flow, and for the more physically meaningful relation of depth to the lower crust inversely proportional to heat flow. If the depth to the reflective lower crust is inversely proportional

Table 1. Heat flow, and travel time and depth to the reflective lower crust and Moho

Location and line	Q/mW·m^{-2}	t_R/s	z_R/km	t_M/s	z_M/km	References
USA						
Colorado Plateau COCORP Utah 4	52±10	7.0 –8.5	23	14.5 –15.0	48	[1, 2, 3, 4, 5]
Eastern Basin and Range COCORP Utah 1	86±10	4.5 –5.5	14	10.0 –10.5	29	[1, 2, 5, 6, 7]
Central Basin and Range COCORP Nevada 6 and 7	79±10	5.0 –6.0	16	10.0 –12.0	33	[1, 2, 5, 7, 8]
Battle Mt. Heat Flow High COCORP Nevada 1east and 2	115±15	3.5 –4.0	10	9.5 –10.0	29	[1, 2, 5, 7, 9]
Western Basin and Range COCORP Nevada 1west	79±10	4.5 –5.0	13	10.0 –10.5	31	[1, 2, 5, 7, 9]
Rio Grande Rift New Mexico COCORP Socorro 2A	90±15	3.5 –4.0	10	11.5 –12.5	34	[10, 11, 12]
North Texas COCORP Hardeman Cty 1, 2, 3	52±10	5.5 –6.5	16	15.5	50	[1, 2, 13, 14]
Long Island New York U.S.G.S lines 36 and 23	50±10	4.5 –6.0	14	9.5 –11.0	30	[1, 2, 15]
Adirondacks New York COCORP New York 7	42±10	5.5 –7.0	20	(11)	35	[1, 2, 16, 17]
Europe						
Cornubian Batholith, U.K. BIRPS SWAT 6	110±20	3.5 –4.5	12	9.5 –10.0	27	[18, 19, 20, 21]
Celtic Sea, U.K. BIRPS SWAT 2, 3north, 4north, 5north	50±15	6.5 –7.25	20	9.5 –10.0	30	[18, 19, 20, 22]
Irish Channel, U.K. BIRPS WINCH 2E, 3, 3A, 4north	65±15	5.0 –5.5	15	9.5 –10.0	30	[18, 19, 23, 24]
North East Coast, U.K. BIRPS NEC	50±15	5.0 –7.0	16	10.0 –11.0	32	[18, 19, 25, 26]
Paris Basin, France ECORS Nord du France	70±10	6.0 –8.0	20	12.0 –13.0	38	[27, 28]
Aachen, West Germany	55±10	5.75–6.25	18	10.0	30	[29, 30]
Northwest German Basin	55±10	6.0 –7.5	19	10.0–10.5	32	[29, 31]
Urach, West Germany U1 and U2	85±15	4.5 –6.0	13	8.5 – 9.0	26	[32, 33, 34]
Dinkelsbühl, West Germany DEKORP 2 South	75±10	5.0 –6.0	17	9.0 –10.0	29	[29, 35]
Rastatt, Rhine Graben West Germany	100±10	6.75–7.25	18	9.25– 9.75	26	[29, 36, 37, 38]
Haslach, Schwarzwald West Germany KTB line 8401	70±10	5.0 –5.5	15	8.5 – 9.0	26	[29, 38, 39, 40]
Australia						
Yilgarn Block Western Aus. Hines Hill traverses Q, R, S	52±10	4.0 –4.5	13	11.0 –12.0	40	[41, 42]
Fraser Ranges Western Aus. Traverses D, E, F	33±10	6.0 –8.0	22	11.0 –12.0	36	[41, 42]
Bowen Basin Queensland Sites F and G	65±15	4.25–5.75	15	11.5 –13.0	36	[41, 43, 44]
Georgina Basin Queensland Sites C and D	80±15	4.0 –5.0	12	(17)	53	[41, 43, 45]
Eromanga Basin Queensland	80±15	7.0 –8.0	20	12.5 –13.5	38	[41, 46, 47, 48]

Q: regional surface heat flow, with an uncertainty subjectively estimated from variability and density of local heat-flow measurements;
t_R and t_M: ranges of two-way travel times to top of reflective lower crust and to reflection Moho, respectively (numbers in brackets: travel times estimated from refraction data);
z_R and z_M: mean estimated depths to top of reflective lower crust and to Moho, respectively. Typical uncertainty in converting a crustal travel time to depth is ±2 km, which therefore represents the minimum error bar on each estimate

References:

1. Sass et al., 1976
2. Lachenbruch and Sass, 1977
3. Farmer et al., 1987
4. Mayer and Brown, 1986
5. Prodehl, 1979
6. Allmendinger et al., 1983
7. Klemperer et al., 1986
8. Potter et al., 1987b
9. Hauge et al., 1987
10. Reiter et al., 1986
11. Brown et al., 1979
12. Olsen et al., 1979
13. Oliver and Kaufman, 1977
14. Tryggvason and Qualls, 1967
15. Phinney, 1986
16. Klemperer et al., 1985
17. Taylor and Toksöz, 1979
18. Bloomer et al., 1979
19. Gale, 1986
20. BIRPS and ECORS, 1986
21. Holder and Bott, 1971
22. Bamford, 1972
23. Brewer et al., 1983
24. Blundell and Parks, 1969

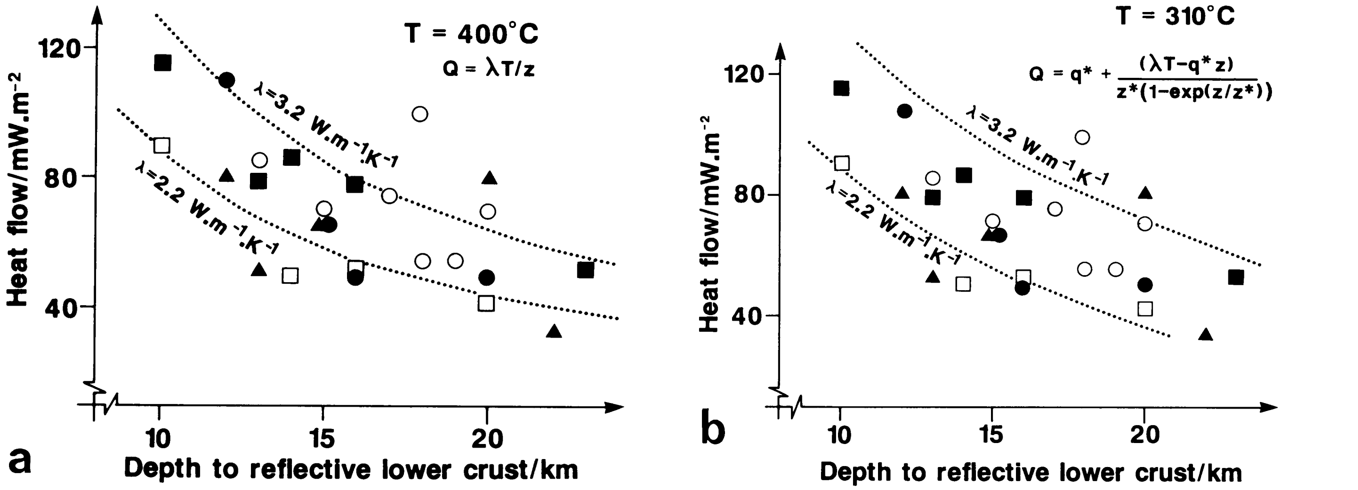

Fig. 4a, b. Plots of heat flow vs. depth to the top of the reflective lower crust for the data in Table 1. **a** *Filled squares*: western USA (Fig. 2); *open squares*: other USA data; *filled circles*: British Isles (Fig. 3); *open circles*: other European data; *triangles*: Australian data. Error bars are omitted for clarity but are comparable with those in Figs. 2c and 3c. Superimposed *dotted lines* are isothermal curves for a temperature at the top of the reflective lower crust, T_z, of 400° C, for a model of a constant geothermal gradient, calculated for plausible values of crustal thermal conductivity. **b** *Dotted lines* are isothermal curves for a temperature at the top of the reflective lower crust, T_z, of 310° C, for the Lachenbruch (1970) model with a reduced heat flow of 33 mW·m^{-2} and a scale length for the depletion of radioactive isotopes of 10 km, calculated for plausible values of crustal thermal conductivity

to heat flow, then the top of the reflective lower crust may be an isothermal boundary in regions of constant thermal conductivity. For the global data and the western USA data the correlation coefficients obtained for the model in which heat flow is inversely proportional to depth are higher than for the model in which heat flow is directly proportional to depth.

Different thermal models of the crust, including both a constant geothermal gradient and the more complex model due to Lachenbruch (1970) in which the concentration of radioactive isotopes decreases exponentially with depth, can be used to calculate isothermal curves relating heat flow to depth for chosen values of thermal conductivity. In the model with constant geothermal gradient, the depth (z) to an isotherm (T_z) is related to the thermal conductivity of the crust (λ) and the surface heat flow (Q) by the equation

$$Q=\lambda T_z/z.$$

A least-squares fit of the equation

$$Q=\lambda T_z/z+\alpha$$

to the data in Table 1 (25 points) gives values for the constants

$$\lambda T_z=866\ \mathrm{W\cdot m^{-1}};\ \alpha=12.8\ \mathrm{mW\cdot m^{-2}}$$

with a correlation coefficient $r=0.64$ (1σ limits: $0.49\le r\le 0.75$). For an assumed upper-crustal conductivity of 2.7 W·m^{-1}·K^{-1} (Smithson and Decker, 1974), these data suggest that the reflective layering preferentially appears in regions in the crust with temperatures above about 320 °C. Isothermal curves are plotted in Fig. 4a for a plausible range of upper-crustal thermal conductivities ($2.2\le\lambda\le 3.2$). Using only the data from the western USA (Fig. 2c; 5 points) values of

$$\lambda T_z=1051\ \mathrm{W\cdot m^{-1}},\ \alpha=7.7\ \mathrm{mW\cdot m^{-2}},$$

and

$$r=0.96\ (1\sigma\ \text{limits}: 0.86\le r\le 0.99)$$

are obtained. Taking $\lambda=2.7$ W·m^{-1}·K^{-1}, a critical isotherm of $T_z\simeq 390$ °C is found (Fig. 2c).

In the Lachenbruch (1970) model of exponential distribution of heat production in the crust, with z^* representing the characteristic depth of the heat-production distribution and q^* the reduced heat flow,

$$\lambda T_z=q^*z+(Q-q^*)z^*\{1-\exp(-z/z^*)\}.$$

Making the simplest possible assumption that this equation is correct world-wide, with constant, global-average, values of $z^*=10$ km and $q^*=33$ mW·m^{-2} (e.g. Turcotte and Schubert, 1982), a best fit to the equation (a minimum value of chi-squared) is obtained with $\lambda T_z=830$ W·m^{-1}. Thus for an assumed crustal conductivity of 2.7 W·m^{-1}·K^{-1} the data of Table 1, when fit to the exponential heat-production model, suggest that the reflective layering tends to be restricted to parts of the crust with temperatures above about 310 °C (Fig. 4b). Given the simplistic assump-

---

25. Klemperer and Matthews, 1987
26. Lewis, 1986
27. Vasseur and Lucazeau, 1982
28. Bois et al., 1986
29. Bram, 1979
30. Meissner et al., 1984
31. Dohr et al., 1983
32. Haenel and Zoth, 1982
33. Zoth, 1982
34. Bartelsen et al., 1982
35. DEKORP Research Group, 1985
36. Dohr, 1970
37. Demnati and Dohr, 1965
38. Ansorge et al., 1970
39. Lüschen et al., 1985
40. Gajewski and Prodehl, 1985
41. Cull, 1982
42. Mathur et al., 1977
43. Mathur, 1983
44. Collins, 1978
45. Finlayson, 1982
46. Cull and Conley, 1983
47. Moss and Mathur, 1986
48. Finlayson et al., 1984

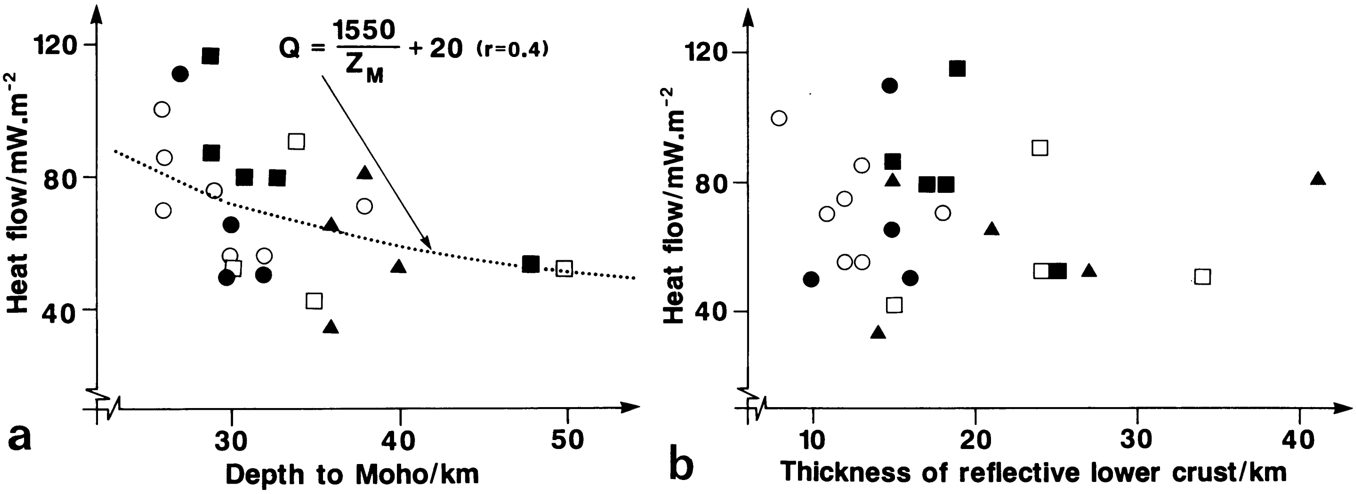

Fig. 5. **a** Plot of heat flow vs. Moho depth for the data in Table 1, with symbols as in Fig. 4. Error bars are omitted for clarity but are comparable with those in Figs. 2c and 3c. The *dotted line* is the best-fitting curve of the form $Q = \lambda T_z/z + \alpha$. **b** Plot of heat flow vs. thickness of reflective lower crust (depth of Moho minus depth to top of the reflective lower crust) for the data in Table 1, with symbols as in Fig. 4. Error bars are omitted for clarity but are comparable with those in Figs. 2c and 3c. The data are essentially uncorrelated

tion in this model of a globally constant reduced heat flow and scale length for the distribution of heat production, it is perhaps remarkable that any correlation of the data with the model is visible.

The data in Table 1 have also been used to make similar correlations between Moho depth and calculated temperature at the Moho (Fig. 5a). Weak negative correlations between heat flow and depth to the Moho have been reported by other authors (e.g. Vyskočil, 1979; Čermák and Zahradník, 1982). However, for the present heat flow data set (severely restricted by the requirement for coincident deep-crustal seismic data) the negative correlation obtained between heat flow and Moho depth and the positive correlation obtained between heat flow and reciprocal Moho depth ($r = 0.41$; 1σ limits: $0.22 \leq r \leq 0.57$ for the global dataset) are clearly worse than the correlations obtained between depth to the reflective lower crust or reciprocal depth to the lower crust. The depth to the Moho is also only very weakly correlated with the depth to the reflective lower crust ($r = 0.2 \pm 0.2$). The reciprocal of the thickness of the lower crust, calculated simply as the reciprocal of the difference between the depth to the Moho and the depth to the top of the reflective lower crust, is uncorrelated with heat flow ($r = 0.1 \pm 0.2$) for the global dataset (Fig. 5b), but may be correlated for regional subsets of the data (Wever et al., 1987).

Discussion

Data presented regionally for the western USA (Fig. 2) and the U.K. (Fig. 3), and as a global compilation (Fig. 4) suggest that the seismically reflective lower crust may be restricted to regions with temperatures greater than 300°–400 °C. The very definition of the 'reflective lower crust' as existing beneath a largely non-reflective upper crust requires that the reflectivity of the lower crust occurs only at elevated temperatures. Nonetheless, the demonstrable inverse correlation between heat flow and depth to the top of the reflective lower crust suggests that a necessary, though not sufficient, condition for the development of the lower crustal reflectivity is the existence of a suitably high temperature. However, temperature cannot be the only controlling variable, and to some extent multiple processes must be operating, since the correlations obtained are imperfect (correlation coefficient < 1.0). In several areas the top of the reflective crust has a substantial dip (30° or more) where it follows a tectonic boundary [e.g. Outer Isles fault zone (Brewer et al., 1983); Appalachian suture, Georgia (Nelson et al., 1985)]. Because the thermal time constant of the crust is only about 10 Ma (Lachenbruch and Sass, 1977), such dipping boundaries are most unlikely to be isotherms at the present day. Thus the occasional existence of steeply dipping boundaries to the top of the reflective lower crust also implies that development of the reflectivity requires both elevated temperatures and additional factors, presumably the existence in the lower crust of suitable lithologic contrasts.

The depth to the reflective lower crust is more strongly correlated with reciprocal heat flow than is the Moho depth (Figs. 4 and 5a), and the thickness of the reflective lower crust is neither constant nor is it correlated with heat flow or reciprocal heat flow (Fig. 5b). Moho depth is only very weakly correlated with depth to the reflective lower crust. Therefore, the relation between heat flow and reflectivity of the lower crust is most probably due to the nature of the reflectors, rather than being controlled by independent constraints on crustal thickness or lower crustal thickness *per se*.

Is lower crustal reflectivity transient?

The correlation of the top of the reflective lower crust with an isotherm of about 300°–400 °C is based on data from crustal provinces ranging in age from Archaean to Neogene. If the correlation is correct, and has been maintained over geological time, then as surface heat flow varies through time (due either to changing mantle heat flux or to erosion of the uppermost levels of the crust that are enriched in radioisotopes), then the position of the top of the reflective lower crust must vary with respect to a fixed marker horizon in the crust. Hypotheses for the origin of the reflectivity of the lower crust that depend only on the presence of different rock types, such as multiple mafic sills or layered igneous intrusions, do not easily permit the layering to move both up and down with respect to a marker horizon in the crust. In contrast, hypotheses for the origin of the reflectivity of the lower crust that rely on time- or temperature-dependent properties of rocks may satisfy this requirement. Two hypotheses that are time- and temperature-dependent relate to the possible existence of ductile strain banding or fluid-filled cracks in the lower crust (e.g. Matthews and Cheadle, 1986).

Further evidence to suggest that many seismic reflectors in the lower crust are transient and are not related solely to lithologic boundaries comes from reflection profiles over terranes exhumed from the lower crust. Crustal reflection profiles have been recorded over exposed granulites of the Fraser Range, Western Australia (Mathur et al., 1977), the Erzgebirge, East Germany (Bölsche and Kresser, 1978), the Lewisian, northwest Scotland (Brewer et al., 1983) and the Adirondacks, New York State (Klemperer et al., 1985). In all four cases the granulitic basement forming the upper crust, which must have at one time been at lower crustal depths, is relatively non-reflective. In contrast the present-day lower crust, probably at granulite facies at some time during its formation or evolution, is highly reflective in all these areas. If these exposed granulite terranes are lithological analogues of the present-day lower crust, then the lack of pronounced reflectivity in the upper crust must be due to changes that occurred as a result of uplift. These changes might include changing rheology as a result of reduced temperature and consequent brittle deformation near the surface, or loss of trapped fluids and opening of cracks due to reduced confining pressure near the surface.

Possible causes of transient reflectivity

At least some ductile shear zones have been shown to be reflective where they outcrop near the surface (Hurich et al., 1985). Ductile banding may be widely developed below the 'brittle-ductile transition' which occurs at about 350 °C (but is highly dependent on lithology, strain rate, fluid activity, etc.) (Meissner and Strehlau, 1982), a temperature commensurate with the range of temperatures identified in this paper as marking the upper boundary of the reflective lower crust. If the temperature of the crust increases, ductile banding could be developed at successively higher levels in the crust, producing a corresponding rise in top of the reflective lower crust. If the crustal temperature declines and the brittle-ductile transition moves to a deeper level in the crust, then ductile banding might be caused to disappear, or at least to cease to become reflective, by long-term creep or chemical processes, or brittle deformation (Wever et al., 1987).

Another possible cause of the reflectivity of the lower crust is the presence of free fluids in the lower crust (Matthews, 1986). Free fluids in the lower crust have also been invoked in recent years to explain the widespread observation of extremely high electrical conductivity in the lower crust (e.g. Shankland and Ander, 1983). The generally very low conductivity of the upper crust is usually taken to imply an absence of fluids, or at least a very different distribution of fluids, in the upper crust. Although little is known about the ways in which fluids might reside in the lower crust, the confining mechanism of lower-crustal fluids must depend on the rheology of the lower crust and hence on the temperature. It has been suggested that lower-crustal conductive zones may be confined to regions where metamorphic dehydration reactions are able to provide free fluids, at temperatures warmer than about 400 °C (Jones, 1987) or above 500 °C (Ádám, 1976), and thus to the thermal region identified in this paper as the reflective lower crust. In this model, reflectors would represent local zones of high fluid pore-pressures, probably but not necessarily constrained by pre-existing lithological boundaries. The large-scale distribution of such fluids might also be controlled by the presence of major crustal faults, giving rise to the dipping boundaries to the reflective lower crust discussed above. Lower-crustal fluids might be expected to be mobile on geological time scales, migrating relatively rapidly in response to changes in crustal temperature and rheology.

Although this model of widespread free fluids in the lower crust can explain the observations presented in this paper there remain many questions, not least the source of the fluids and problem of how to maintain high fluid pressures in discrete layers separated by layers with much lower fluid pressure. Measured rock permeabilities are sufficiently high that fluids should diffuse out of high-pressure zones very rapidly on geologic time scales (Jones and Nur, 1984). If free fluids are the cause of the observed lower-crustal reflectivity, then an efficient mechanism is required to supply very large quantities of fluid into the lower crust to maintain locally high fluid pressures: the volume of fluid contained by 1% porosity in a 15-km layer of the continental crust of the whole Earth is equivalent to the entire volume of water ($\simeq 30 \times 10^6$ km^3) subducted every 20 Ma (Fyfe and Kerrich, 1985). However, the exposure at the surface of high-grade anhydrous metamorphic terranes (granulite terranes) is commonly taken to imply that the lower crust contains negligible amounts of free water. If these objections can be overcome, it is well known that strong reflections can be generated by fluid-filled porosity, including hydrocarbon accumulations marked by 'bright spots' (e.g. Dobrin, 1976). Changes in fluid-filled porosity or fluid pore-pressure, measured to 12 km depth in the Kola borehole in crystalline basement, have been directly linked to changes in seismic properties and to seismic reflections (Karus et al., 1982; Kozlovsky, 1984).

Conclusions

There exists a moderate inverse correlation between the depth to the reflective lower crust and the surface heat flow, when averaged over tectonic provinces, that can be demonstrated both regionally and globally. This correlation implies that one requirement for the development of the reflectivity of the lower crust is the existence of a sufficiently high temperature in the lower crust. This hypothesis will be tested further as additional seismic profiles are run in areas of laterally varying heat flow. In particular, more experiments are needed in areas of very high heat flow to confirm that the correlation proposed in this paper is a consequence of the nature of the reflective lower crust rather than an effect of the way in which the reflective lower crust is defined and recognised. A further implication of the proposed correlation is that the lower-crustal reflectivity must be variable on the same time scales as crustal temperatures are believed to vary, i.e. tens of millions of years.

This result favours, as a cause of the lower-crustal reflectivity, mechanisms that may be transient features of the crust, notably free fluids or ductile strain banding. Indeed, these possibilities are not mutually exclusive. The lower-crustal reflectivity may be best developed where fluids are present to enhance reflectivity due to lithologic contrast in ductile regions of the crust.

Acknowledgements. It is a pleasure to acknowledge the help of many colleagues at COCORP, USGS Branch of Seismology, BIRPS, and Institut für Geophysik Kiel, who have provided data, ideas and discussion. In particular, T. Brocher provided a detailed and critical review. The seismic section in Fig. 1 was prepared by R.V. Hobbs using Merlin Profilers' Seismic Kernel System on the VAX 11/750 computer system operated by Bullard Laboratories. BIRPS is funded by the Natural Environment Research Council. Cambridge Earth Sciences Contribution no. 824.

References

Ádám, A.: Quantitative connections between regional heat flow and the depth of conductive layers in the Earth's crust and upper mantle. Acta. Geodaet., Geophys. et Montanist. Acad. Sci. Hung. **11**, 503–509, 1976

Allmendinger, R.W., Sharp, J.W., Von Tish, D., Serpa, L., Brown, L., Kaufman, S., Oliver, J., Smith, R.B.: Cenozoic and Mesozoic structure of the eastern Basin and Range province, Utah, from COCORP seismic-reflection data. Geology **11**, 532–536, 1983

Allmendinger, R.W., Hauge, T.A., Hauser, E.C., Potter, C.J., Klemperer, S.L., Nelson, K.D., Knuepfer, P., Oliver, J.: Overview of the COCORP 40° N transect, western USA: the fabric of an orogenic belt. Geol. Soc. Am. Bull. **98**, (3), 1987a (in press)

Allmendinger, R.W., Hauge, T.A., Hauser, E.C., Potter, C.J., Oliver, J.: Tectonic heredity and the layered lower crust in the Basin and Range province, western United States. In: Continental extensional tectonics, M.P. Coward, J.F. Dewey, eds., Spec. Publ. Geol. Soc. Lond., **28**, 223–246, 1987b

Ando, C.J., Cook, F.A., Oliver, J.E., Brown, L.D., Kaufman, S.: Crustal geometry of the Appalachian orogen from seismic reflection studies. Geol. Soc. Am. Mem. **158**, 83–101, 1983

Ansorge, J., Emter, D., Fuchs, K., Lauer, J.P., Müller, St., Peterschmitt, E.: Structure of the crust and upper mantle in the rift system around the Rhine graben. In: Graben problems, J.H. Illies, St. Mueller, eds.: pp. 190–197. Stuttgart: Schweizerbart, 1970

Bamford, S.A.D.: Evidence for a low-velocity zone in the crust beneath the western British Isles. Geophys. J. R. Astron. Soc. **30**, 101–105, 1972

Bartelsen, H., Lueschen, E., Krey, Th., Meissner, R., Schmoll, H., Walter, Ch.: The combined seismic reflection-refraction investigation of the Urach geothermal anomaly. In: The Urach geothermal project, R. Haenel, ed.: pp. 247–262. Stuttgart: Schweizerbart, 1982

BIRPS, ECORS: Deep seismic reflection profiling between England, France and Ireland. J. Geol. Soc. Lond. **143**, 45–52, 1986

Bloomer, J.R., Richardson, S.W., Oxburgh, E.R.: Heat flow in Britain: an assessment of the values and their reliability. In: Terrestrial heat-flow in Europe, V. Čermák, L. Rybach, eds.: pp. 293–300. Berlin: Springer-Verlag, 1979

Blundell, D.J., Parks, R.: A study of the crustal structure beneath the Irish Sea. Geophys. J. R. Astron. Soc. **17**, 45–62, 1969

Bois, C., Cazes, M., Damote, B., Galdéano, A., Hirn, A., Mascle, A., Matte, P., Raoult, J.F., Torreilles, G.: Deep seismic profiling of the crust in northern France: the ECORS project. In: Reflection seismology: A global perspective, M. Barazangi, L. Brown, eds.: A. G. U. Geodynamics Series **13**, 21–29, 1986

Bölsche, J., Kresser, K.D.: First results of depth-seismic exploration in the southern part of the G.D.R. In: Proceedings of the XXIInd International Geophysics Symposium, Prague (1977), 457–471, 1978

Bram, K.: Heat flow measurements in the Federal Republic of Germany. In: Terrestrial heat-flow in Europe, V. Čermák, L. Rybach, eds.: pp. 191–196. Berlin: Springer-Verlag, 1979

Brewer, J.A., Matthews, D.H., Warner, M.R., Hall, J., Smythe, D.K., Whittington, R.J.: BIRPS deep seismic reflection studies of the British Caledonides. Nature **305**, 206–210, 1983

Brown, L.D., Krumhansl, P.A., Chapin, C.E., Sanford, A.R., Cook, F.A., Kaufman, S., Oliver, J.E., Schilt, F.S.: COCORP seismic reflection studies of the Rio Grande Rift. In: Rio Grande Rift: tectonics and magmatism, R.E. Riecker, ed.: A. G. U., 1979

Brown, L., Ando, C., Klemperer, S., Oliver, J., Kaufman, S., Czuchra, B., Walsh, T., Isachsen, Y.W.: Adirondack-Appalachian crustal structure: the COCORP northeast traverse. Geol. Soc. Am. Bull. **94**, 1173–1184, 1983

Brown, L., Barazangi, M., Kaufman, S., Oliver, J.: The first decade of COCORP: 1974–1984. In: Reflection seismology: A global perspective, M. Barazangi, L. Brown, eds.: A. G. U. Geodynamics Series **13**, 107–120, 1986

Čermák, V.: Heat flow map of Europe. In: Terrestrial heat-flow in Europe, V. Čermák, L. Rybach, eds.: pp. 3–40. Berlin: Springer, 1979

Čermák, V., Zahradník, J.: Two-dimensional correlation of heat flow and crustal thickness in Europe. In: Geothermics and geothermal energy, V. Čermák, R. Haenel, eds.: pp. 17–25. Stuttgart: Schweizerbart 1982

Collins, C.D.N.: Crustal structure of the central Bowen Basin, Queensland. BMR J. Austral. Geol. Geophys. **3**, 203–209, 1978

Cook, F.A., Simony, P.S., Coflin, K.C., Green, A.G., Milkereit, B., Price, R.A., Parrish, R., Patenaude, C., Gordy, P.L., Brown, R.L.: Lithoprobe southern Canadian Cordilleran transect: Rocky Mountain thrust belt to Valhalla gneiss complex. Geophys. J. R. Astron. Soc. **89**, 91–98, 1987

Cull, J.P.: An appraisal of Australian heat-flow data. BMR J. Austral. Geol. Geophys. **7**, 11–21, 1982

Cull, J.P., Conley, D.: Geothermal gradients and heat flow in Australian sedimentary basins. BMR J. Austral. Geol. Geophys. **8**, 329–337, 1983

DEKORP Research Group: First results and preliminary interpretation of deep-reflection seismic recordings along profile DEKORP 2-South. J. Geophys. **57**, 137–163, 1985

Demnati, A., Dohr, G.: Reflexionsseismische Tiefensondierungen im Bereich des Oberrheintalgrabens und des Kraichgaues. J. Geophys. **31**, 229–245, 1965

Dobrin, M.B.: Introduction to geophysical prospecting, 3rd edn. New York: McGraw-Hill, pp. 630, 1976

Dohr, G.: Reflexionsseismische Messungen im Oberrheingraben mit digitaler Aufzeichnungstechnik und Bearbeitung. In: Graben problems, J.H. Illies, St. Mueller, eds.: pp. 207–218. Stuttgart: Schweizerbart, 1970

Dohr, G., Lukić, P., Bachmann, G.H.: Deep crustal reflections in the Northwest-German Basin. In: Seismic expression of structural styles, A.W. Bally, ed.: Am. Ass. Petrol. Geol. Studies in Geology **15**, (1), 1.5–1 – 1.5–5, 1983

Farmer, H.G., Allmendinger, R.W., Peddy, C., Oliver, J., Kaufman, S., Brown, L.: Crustal structure of the Basin and Range-Colorado Plateau transition in central Utah from COCORP seismic reflection data. Bull. Am. Assn. Petrol. Geol. 1987 (in press)

Finlayson, D.M.: Seismic crustal structure of the Proterozoic North Australian craton between Tennant Creek and Mount Isa. J. Geophys. Res. **87**, 10569–10578, 1982

Finlayson, D.M., Collins, C.D.N., Lock, J.: P-wave velocity features of the lithosphere under the Eromanga Basin, easten Australia, including a prominent mid-crustal (Conrad?) discontinuity. Tectonophysics **101**, 267–291, 1984

Fyfe, W.S., Kerrich, R.: Fluids and thrusting. Chem. Geol. **49**, 353–362, 1985

Gajewski, D., Prodehl, C.: Refraction seismic investigation, Schwarzwald. Abstracts, 2nd International Symposium on observation of the continental crust through drilling. Bonn: Alfred-Wegener-Stiftung, 68, 1985

Gale, I.N. (compiler): Geothermal map of the United Kingdom, 1:1,500,000. In: Geothermal energy – the potential in the United Kingdom, R.A. Downing, D.A. Gray, eds.: Keyworth: British Geological Survey 1986

Green, A.G., Berry, M.J., Spencer, C.P., Kanesewich, E.R., Chiu, S., Clowes, R.M., Yorath, C.J., Stewart, D.B., Unger, J.D., Poole, W.H.: Recent seismic reflection studies in Canada. In: Reflection seismology: A global perspective, M. Barazangi, L. Brown, eds.: A. G. U. Geodynamics Series **13**, 85–97, 1986

Haenel, R., Zoth, G.: Temperature measurements and determination of heat flow density. In: The Urach geothermal project, R. Haenel, ed.: pp. 81–88. Stuttgart: Schweizerbart, 1982

Hall, J.: Nature of the lower continental crust: evidence from BIRPS work on the Caledonides. In: Reflection seismology: The continental crust, M. Barazangi, L. Brown, eds.: A.G.U. Geodynamics Series **14**, 223–231, 1986

Hauge, T., Allmendinger, R., Caruso, C., Hauser, E., Klemperer, S., Opdyke, S., Potter, C., Sanford, W., Brown, L., Kaufman, S., Oliver, J.: Crustal structure of western Nevada from COCORP deep seismic reflection data. Geol. Soc. Am. Bull. **98**, (3), 1987 (in press)

Holder, A.P., Bott, M.H.P.: Crustal structure in the vicinity of south-west England. Geophys. J. R. Astron. Soc. **23**, 465–489, 1971

Hurich, C.A., Smithson, S.B., Fountain, D.M., Humphreys, M.C.: Seismic evidence of mylonite reflectivity and deep structure in the Kettle Dome metamorphic complex, Washington. Geology **13**, 577–580, 1985

Jones, A.G.: MT and reflection: an essential combination Geophys. J. R. Astron. Soc. **89**, 7–18, 1987

Jones, T.D., Nur, A.: The nature of seismic reflections from deep crustal fault zones. J. Geophys. Res. **89**, 3153–3171, 1984

Karus, Ye.V., Kuznetsov, O.L., Kuznetsov, Yu.I., Nartikoyev, V.D.: Possible nature of seismic boundaries at depth. Akademiia Nauk SSSR Doklady Earth Science Sections **265**, (Jul–Aug), 14–16, 1982

Klemperer, S.L., Matthews, D.H.: Iapetus suture located beneath the North Sea by BIRPS deep seismic reflection profiling. Geology **15**, (3), 1987 (in press)

Klemperer, S.L., Brown, L.D., Oliver, J.E., Ando, C.J., Czuchra, B.L., Kaufman, S.: Some results of COCORP seismic reflection profiling in the Grenville-age Adirondack mountains. New York State. Can. J. Earth Sci. **22**, 141–153, 1985

Klemperer, S.L., Hauge, T.A., Hauser, E.C., Oliver, J.E., Potter, C.J.: The Moho in the northern Basin and Range province along the COCORP 40° N seismic reflection transect. Geol. Soc. Am. Bull. **97**, 603–618, 1986

Kozlovsky, Ye.A.: The world's deepest well. Scientific American **251**, (6), 98–104, 1984

Lachenbruch, A.H.: Crustal temperature and heat production: implications of the linear heat-flow relation. J. Geophys. Res. **75**, 3291–3300, 1970

Lachenbruch, A.H., Sass, J.H.: Heat flow in the United States and the thermal regime of the crust. In: The Earth's Crust, J.G. Heacock, ed.: A. G. U. Geophysical Monogr. **20**, 626–675, 1977

Lewis, A.H.J.: The deep seismic structure of northern England and adjacent marine areas from the Caledonian suture seismic project. University of Durham Ph.D. thesis, pp. 167, 1986

Lüschen, E., Menges, D., Rühl, Th., Sandmeier, K.-J., Wenzel, F., Gowin, J., Janoth, W., Keller, F., Stiller, M., Söllner, W., Trappe, H.: Presite seismic reflection survey, Schwarzwald. Abstracts, 2nd International Symposium on Observation of the Continental Crust through Drilling. Bonn: Alfred-Wegener-Stiftung, 69, 1985

Lynn, H.B., Hale, L.D., Thompson, G.P.: Seismic reflections from the basal contacts of batholiths. J. Geophys. Res. **86**, 10633–10638, 1981

Mathur, S.P.: Deep reflection experiments in northeastern Australia, 1976–1978. Geophysics **48**, 1588–1597, 1983

Mathur, S.P., Moss, F.J., Branson, J.C.: Seismic and gravity investigations along the Geotraverse, Western Australia, 1969. BMR Bulletin **191**, pp. 63, 1977

Matthews, D.H.: Seismic reflections from the lower crust around Britain. In: The nature of the lower continental crust, J.B. Dawson, D.A. Carswell, J. Hall, K.H. Wedepohl, eds., Spec. Publ. Geol. Soc. Lond. **24**, 11–22, 1986

Matthews, D.H., Cheadle, M.J.: Deep reflections from the Caledonides and Variscides west of Britain and comparison with the Himalayas. In: Reflection seismology: A global perspective, M. Barazangi, L. Brown, eds.: A. G. U. Geodynamics Series **13**, 5–19, 1986

Mayer, J.R., Brown, L.D.: Signal penetration in the COCORP Basin and Range-Colorado Plateau survey. Geophysics **51**, 1050–1055, 1986

McKenzie, D.P.: A possible mechanism for epeirogenic uplift. Nature **307**, 616–618, 1984

Meissner, R.: The Moho as a transition zone. Geophys. Surveys **1**, 195–216, 1973

Meissner, R., Strehlau, J.: Limits of stresses in continental crusts and their relation to the depth-frequency distribution of shallow earthquakes. Tectonics **1**, 73–89, 1982

Meissner, R., Wever, Th.: Nature and development of the crust according to deep reflection data from the German Variscides. In: Reflection seismology: A global perspective, M. Barazangi, L. Brown, eds.: A. G. U. Geodynamics Series **13**, 31–42, 1986

Meissner, R., Lüschen, E., Flüh, E.R.: Studies of the continental crust by near-vertical reflection methods: a review. Phys. Earth Planet. Inter. **31**, 363–376, 1983

Meissner, R., Springer, M., Flüh, E.: Tectonics of the Variscides in North-Western Germany based on seismic reflection measurements. In: Variscan tectonics of the North Atlantic region, D.H.W. Hutton, D.J. Sanderson, eds.: Spec. Publ. Geol. Soc. Lond. **14**, 23–32, 1984

Moss, F.J., Mathur, S.P.: A review of continental reflection profiling in Australia. In: Reflection seismology: A global perspective, M. Barazangi, L. Brown, eds.: A. G. U. Geodynamics Series **13**, 67–76, 1986

Nelson, K.D., Arnow, J.A., McBride, J.H., Willemin, J.H., Huang, J., Zheng, L., Oliver, J.E., Brown, L.D., Kaufman, S.: New COCORP profiling in the southeastern United States. Part 1: Late Paleozoic suture and Mesozoic rift basin. Geology **13**, 714–718, 1985

Oliver, J., Kaufman, S.: Complexities of the deep basement from seismic reflection profiling. In: The Earth's Crust, J.G. Heacock, ed.: A. G. U. Geophysical Monogr. **20**, 243–253, 1977

Olsen, K.H., Keller, G.R., Stewart, J.N.: Crustal structure along the Rio Grande Rift from seismic refraction profiles. In: Rio Grande Rift: tectonics and magmatism, R.E. Riecker, ed.: pp. 127–143. A. G. U. 1979

Phinney, R.A.: A seismic cross-section of the New England Appalachians: the orogen exposed. In: Reflection seismology: The continental crust, M. Barazangi, L. Brown, eds.: A. G. U. Geodynamics Series **14**, 157–172, 1986

Phinney, R.A., Jurdy, D.M.: Seismic imaging of the deep crust. Geophysics **44**, 1637–1660, 1979

Potter, C.J., Allmendinger, R.W., Hauser, E.C., Oliver, J.E.: COCORP deep seismic reflection traverses of the U.S. Cordillera. Geophys. J. R. Astron. Soc. **89**, 99–104, 1987a

Potter, C.J., Liu, C-S, Huang, J., Zheng, L., Hauge, T.A., Hauser, E.C., Allmendinger, R.W., Oliver, J.E., Kaufman, S., Brown, L.: Crustal structure of north-central Nevada: results from COCORP deep seismic profiling. Geol. Soc. Am. Bull. **98**, (3) 1987b (in press)

Prodehl, C.: Crustal structure of the western United States. United States Geological Survey Professional Paper **1034**, pp. 74, 1979

Reiter, M., Eggleston, R.E., Broadwell, B.R., Minier, J.: Estimates of terrestrial heat flow from deep petroleum tests along the Rio Grande Rift in central and southern New Mexico. J. Geophys. Res. **91**, 6225–6245, 1986

Sass, J.H., Diment, W.H., Lachenbruch, A.H., Marshall, B.V., Munroe, R.J., Moses, T.H., Jr., Urban, T.C.: A new heat-flow contour map of the conterminous United States. United States Geological Survey Open File Report **76–756**, pp. 24, 1976

Setto, I., Meissner, R.: Support of gravity modelling for seismic interpretation. Terra Cognita **6**, 346, 1986

Shankland, T.J., Ander, M.E.: Electrical conductivity, temperatures and fluids in the lower crust. J. Geophys. Res. **88**, 9475–9484, 1983

Smithson, S.B., Decker, E.R.: A continental crustal model and its geothermal implications. Earth Planet. Sci. Letts. **22**, 215–225, 1974

Taylor, S.R., Toksöz, M.N.: Three-dimensional crust and upper mantle structure of the northeastern United States. J. Geophys. Res. **84**, 7627–7644, 1979

Turcotte, D.L., Schubert, G.: Geodynamics: Applications of continuum physics to geological problems. New York: Wiley and Sons, pp. 450, 1982

Tryggvason, E., Qualls, B.R.: Seismic refraction measurements of crustal structure in Oklahoma. J. Geophys. Res. **72**, 3738–3740, 1967

Vasseur, G., Lucazeau, F.: Some aspects of heat flow in France. In: Geothermics and geothermal energy, V. Čermák, R. Haenel, eds.: pp. 79–89. Stuttgart: Schweizerbart, 1982

Vyskočil, P.: Heat flow, crustal thickness and recent vertical movements. In: Terrestrial heat-flow in Europe, V. Čermák, L. Rybach, eds.: pp. 119–125. Berlin: Springer-Verlag, 1979

Wever, Th., Trappe, H., Meissner, R.: Crustal reflectivity pattern and its relation to geological provinces. European Association of Exploration Geophysicists, 48th Annual Meeting, Ostend, Beglium. Technical programme and abstracts of papers, 88, 1986

Wever, Th., Trappe, H., Meissner, R.: Possible relations between crustal reflectivity, crustal age, heatflow, and viscosity of the continents. Ann. Geophys. 1987 (in press)

Zoth, G.: The temperature field of the Urach area. In: The Urach geothermal project, R. Haenel, ed.: pp. 205–221. Stuttgart: Schweizerbart, 1982

Received July 17, 1986; revised version November 24, 1986
Accepted November 26, 1986