LISPB DELTA, a lithospheric seismic profile in Britain: analysis and interpretation of the Wales and southern England section

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Abstract: The Lithospheric Seismic Profile in Britain (LISPB), shot in 1974, included a 310 km profile LISPB DELTA, crossing the Palaeozoic Welsh Basin, the western extent of the Midland microcraton and the Cornubian zone of southern England. This first comprehensive analysis of these data has produced a sub-horizontally layered seismic and associated gravity model that correlates well with surface geology. A north–south decrease in crustal velocity and density corresponds to the change from Avalonian crust into the Rheno-Hercynian zone at the south end of the profile. High velocities and densities in the lowest crustal layer beneath north Wales are proposed to result from Cenozoic and possibly Ordovician igneous intrusive rocks, the former derived from an upwelling plume associated with the opening of the North Atlantic. Examination of the load distribution throughout the model shows that it is strongly correlated with the earthquake distribution along LISPB DELTA. Earthquake focal depth also correlates with heat flow. A simple heat-flow profile has been derived, and the seismic velocity model used to constrain crustal heat production values. A long-wavelength excursion from published data can be explained in terms of an increase in mantle heat flow resulting from a previously identified deep thermal anomaly beneath the Irish Sea.

Supplementary material: normalised, synthetic and true amplitude seismic record sections including picks and modelled phases for shots SP04, SP05, SPS1, SPS2 are available at http://www.geolsoc.org.uk/SUP18450.

Regional geology and tectonics

LISPB DELTA lies across a landmass that has experienced the closure of two oceans. The geology is complex and has been extensively described elsewhere (e.g. Woodcock & Strachan 2002a). A brief summary relevant to the present study is set out here.

In the Late Neoproterozoic much of Wales and central England formed part of an Andean-type orogen on the northern margin of the Vendian supercontinent (Holdsworth et al. 2002). The region is formed of calc-alkaline igneous complexes in the form of gabbro, diorite, granodiorite and granite and/or their extrusive volcanic equivalents (Pharaoh & Gibbons 1994; Strachan 2002). The oldest plutonic rocks (c. 702 Ma) occur close to LISPB DELTA in the Welsh Borderland Fault System (Fig. 1b).

The break-up of the Vendian supercontinent between 600 and 580 Ma resulted in the separation of both Laurentia and Baltica from Gondwana and the creation of the Iapetus Ocean. There followed a long period of quiescence on the northern margin of Gondwana before the subsequent closure of this ocean. It was during this period in the Early to Mid-Cambrian that the Welsh Basin (Fig. 1b), crossed by the seismic profile and including deep marine sediments and turbidites, formed in-board from the continent–ocean boundary. In the Late Cambrian, complementary uplift appears to have occurred in the region of present south Wales, the regional ‘basin and range’ morphology suggesting a developing transform-faulted rifted margin presaging the break-up of Gondwana on the southern margin of Iapetus. The southern margin of the Welsh Basin was controlled by the Welsh Borderland Fault System, which underwent significant strike-slip displacement when rifting of the western Midland Platform.
Fig. 1. (a) Location map for the Lithospheric Seismic Profile in Britain (LISPB), showing sections ALPHA, BETA, GAMMA (see e.g. Bamford et al. 1978; Barton 1992) and DELTA (reported here). (b) Tectonic map of western UK, showing LISPB DELTA from north Wales to the English Channel (after Soper 1986; Bluck et al. 1992; Warr 2002; Woodcock & Strachan 2002b; Woodcock et al. 2007). MSFS, Menai Strait Fault System; BF, Bala Fault; WBFS, Welsh Borderland Fault System; BCFZ, Bristol Channel Fault Zone. Palaeozoic faults and cleavage traces after Soper (1986). Variscan faults and cleavage traces after British Geological Survey (1996). (c) Geology map of western UK (after British Geological Survey 2001) and section showing LISPB DELTA from north Wales to the English Channel. Principal geological units and those relevant to LISPB DELTA are identified as, for example, ‘Ord’ for ‘Ordovician’. Shotpoints are identified (SP04, SP05, SPS1, SPS2) and stations are as for tectonic map. Section (modified from Dunning 1985) scaled to geological map. BF, Bala Fault; PL, Pontesford Lineament; CSF, Church Stretton Fault; BCFZ, Bristol Channel Fault Zone.
occurred (Woodcock 2002). Eastern Avalonia was one of the principal continental fragments detached by subduction-related rifting with associated magmatism from the northern margin of Gondwana in the early Ordovician (c. 475 Ma). The subduction led to closure of Iapetus during the Silurian with the collision of Avalonia with the margin of Laurentia, and opening of the Rheic Ocean along the trailing edge of this microcontinent.

Following this collision (c. 425 Ma) extensional basin-forming events were, in the southern part of Avalonia, made more complex by the gradual closure of the Rheic Ocean. In south Wales, the non-marine Early Devonian succession (the Lower Old Red Sandstone) crossed by LISPB DELTA resulted from the vigorous supply of predominantly river-borne sediment from the maturing uplands of northern Wales and England. In SW England, offshore deep marine and marine shelf basins were supplied with marine and alluvial sediments off the southern margin of Avalonia. The Early to Mid-Devonian Acadian event in the Welsh Basin records uplift and crustal thickening as a result of encroachment of the Armorican microcontinent (c. 400 Ma) from the south (Warr 2002), with closure resulting in a major unconformity between the Lower and Upper Old Red Sandstone. Subsequent episodic sea-level rise in Late Devonian times provided widespread alluvial and marginal marine sedimentation to generate the Upper Old Red Sandstone in south Wales and across the Cornubian basins of SW England (Fig. 1c).

Shallow marine deposition continued during the Carboniferous in south Wales and SW England, and as Variscan deformation migrated northward the depocentres eventually developed into flexural foreland basins. Continuing Variscan deformation resulted in northward-directed thrusting in south Wales and widespread basin inversion during the last stages of the assembly of the Gondwana-derived microplates into the supercontinent of Pangaea (Warr 2002). The LISPB DELTA profile crosses the Variscan Front in the vicinity of the Mendip Hills to the south of the Bristol Channel, and to the south passes across the Avalonian foreland Rheico-Hercynian Cornubian zone affected by Variscan folding and thrusting. It has been suggested (Woodcock et al. 2007) that the lack of Acadian deformation in this Cornubian zone between the Bristol Channel Fault Zone (Fig. 1b) and the Armorican Rhiec suture immediately off the south coast of SW England occurs as a result of major dextral strike-slip movement along the crustal-scale Bristol Channel Fault Zone and its eastward extension to the Bray Fault in NW Europe, sometime between the Acadian and late Variscan. This transferred Rheico-Hercynian crust originally adjacent to that in the present Rhenish Massif some 400 km westward into its present location in SW England.

Post-orogenic collapse of the Variscan mountains and Proto-Atlantic rifting dominated the subsequent development of southern Britain (Ruffell & Shelton 2002), a complex network of basins being created in the zone between two major rift systems developing north from the central Atlantic and south from the Norwegian–Greenland Rift into the North Sea (Peacock 2004). Reactivated Caledonian, Devonian and Carboniferous structures controlled the basin trend and sediment distribution during the Mesozoic and Tertiary (Coward 1993). The southern section of LISPB DELTA crosses this region covered by Late Cretaceous sediments, these at the time blanketing most of the NW European continental shelf.

During the Cenozoic, Britain was affected by major regional exhumation of up to about 2000 m, caused by uplift and resulting denudation. The causes of this uplift are debated, and have been suggested to result from underplating at or near the base of the crust (e.g. Brodie & White 1994; Clift 1999), the presence of an anomalously hot, low-density region in the lithosphere and asthenosphere down to at least 200 km (Bott & Bott 2004) or continental shortening resulting from compressional stress transmitted to the plate interior from the plate boundary (Hillis et al. 2008). The northern part of LISPB DELTA lies close to the area of maximum Cenozoic denudation through the East Irish Sea and northern England as described by Jones et al. (2002). The southern part of the profile crosses a region where there is less evidence of this uplift but Hillis et al. (2008) have demonstrated stratigraphic evidence of tectonic shortening.

**Data acquisition and processing**

LISPB data acquisition details have been described elsewhere (Bamford et al. 1976). Briefly, 60 mobile three-component seismic recording stations occupied four segments (ALPHA, BETA, GAMMA and DELTA) of the profile at different times, recording the various shots (Fig. 1a). Seismic energy for this study, recorded on segment DELTA originated from the four shotpoints (north to south) SP04, SP05, SPS1 and SPS2. Reduced time ($V_R = 6 \text{ km s}^{-1}$) record sections from the four shots (filtered between 2 and 20 Hz) are shown in Figure 2. There were two deployments, both having a nominal station spacing of c. 5 km, with the second being shifted by half this amount with respect to the first. This produced a final record spacing of c. 2.5 km over a total distance of 310 km. For shotpoint SP05 only one shot was recorded, resulting in a less dense section with a station spacing of c. 5 km (Fig. 2b). The data and time signal (MSF, HBG or DCF) were recorded on analogue recording systems (the German MARS (Berkhemer 1970) and the UK Geostore). Instrument locations were identified from field maps to a nominal accuracy of $\pm 10$ m. The observed ranges for shots into DELTA (see Table 1) were sufficient for seismic penetration of the Moho. The two land shots SP04 (5000 lb) and SP05 (2000 lb) were fired in boreholes, and provided signals with dominant frequencies between about 3–7 and 4–10 Hz respectively (Fig. 2a and b). Shot SPS1 (0.4 tons) was fired on the sea-bed, providing a signal with dominant frequencies between 4–10 Hz (Fig. 2c) whereas shot SPS2 (4 $\times$ 0.2 tons) was fired at optimum depth, the signal having observed dominant frequencies between about 2–5 Hz (Fig. 2d).

Immediately after the LISPB field project a substantial effort was made to provide high-quality digital files of the recorded data (Bamford et al. 1976). The files obtained for this study include reduced time vertical component record sections of limited time length. As a result of the short record lengths, no S-wave analysis has been undertaken. The files have been read into the ZP processing package (B. Zelt, pers. comm.), which allowed trace killing and various filtering and plotting procedures to enhance the recorded seismic phases and their trace-to-trace correlation for subsequent picking and analysis.

| Shotpoint | Observed range (km) |
|-----------|---------------------|
| SP04 (to north) | 0–20 |
| SP04 (to south) | 0–280 |
| SP05 (to north and south) | 0–150 |
| SPS1 (to north) | 0–310 |
| SPS2 (to north) | 80–390 |
Data modelling

Phase identification and picking

Barton (1992) has stated that 'the experiment ... result[ed] in outstandingly good data'. Those from DELTA are no exception. Figure 2 shows record sections (both normalized and true amplitude) together with modelled phase correlations and associated synthetic seismogram sections. These were produced using the ZP package and the RAYINVR suite of programs (Zelt & Smith 1992) respectively. The phase correlations appear similar to those described by Bamford et al. (1978) for northern Britain. However, owing to DELTA being the only LISPB segment crossing the Palaeozoic Welsh Basin and the Variscides of SW England, they are noticeably different. Although it is not possible to relate a particular phase to an arrival from a particular horizon without a priori information, the major phases are here so described owing to their identification from the subsequent modelling. Using Bamford et al.'s (1978) terminology (see Fig. 3) the following phases have been identified.

The 'a' curves are first arrival refracted diving waves out to distances of 125–152 km from the four shotpoints. Phase a, with an apparent velocity of c. 4 km s\(^{-1}\), is a refracted (or direct) arrival through the uppermost layers, interpreted as being near-surface, low-velocity sediments. Phase a\(_0\), with an average apparent velocity of c. 5.9 km s\(^{-1}\), is a diving wave arrival from the underlying layer, within the Palaeozoic Welsh Basin between SP04 and SP05 to the north and from beneath the near-surface sedimentary layer between SP05 to the south and SPS1. Phase a\(_1\), with an average apparent velocity of c. 6.4 km s\(^{-1}\), being of higher velocity than a\(_0\) is a diving wave from beneath a deeper interface, either within or beneath the Palaeozoic Welsh
Basin from SP04 and SP05 (to the north) and most reasonably from within the crystalline basement from SP05 (to the south), SPS1 and SPS2.

Beyond these arrivals, the phases can be described as by Bamford et al. (1978).

Phase d, consisting of sub-Moho arrivals has been split into two, $d_1$ and $d_2$. $d_1$ is the Moho diving wave (conventionally Pn) which is seen to c. 220–245 km from SP04, SPS1 and SPS2.

Thereafter from the latter two shots, a slightly faster arrival $d_2$ is seen to ranges of just under 300 km from SPS1 and 380 km from SPS2. With SP04 being a land borehole shot, the energy appears insufficient to excite this phase.

Phase c, the high-amplitude Moho reflection is seen from all shots, its maximum amplitude occurring at c. 130 km for SP04 and SP05 (N), c. 120 km for SP05 (S), c. 110 km for SPS1 and c. 90 km for SPS2. These values provide evidence for thicker crust beneath the northern section of the profile beneath the Palaeozoic Welsh Basin.

As for LISPB ALPHA, BETA and GAMMA across northern Britain (Bamford et al. 1978), there is a further well-defined intracrustal diving wave and its associated reflection, and also weak arrivals from mid-crustal discontinuities.

Phase e appearing before the larger amplitude phase c, can be identified on all four record sections. It continues at longer offsets into phase f. These phases are identified as a reflection and associated long offset diving wave from a layer in the lowest part of the crust.

Phase b can be identified on the record sections from SP04 and SP05. It is a phase observed over relatively short ranges between about 30 and 95 km, occurring between the a and the e or c phase travel times on these sections, and is considered to be caused by energy reflected either from the layer defined by the diving wave $a_1$ or from a further mid-crustal discontinuity that has little horizontal continuity.

Four further phases (or bursts of enhanced energy) were identified: (1) two significantly curved phases immediately behind the first arrival (the mantle diving wave) from SPS1 and SPS2 between distances of about 100 and 120 km from SP04 (Fig. 2c and d); (2) an apparently high-frequency (c. 10 Hz) near-linear second arrival phase from SP04 from 2 s reduced
time at 110 km to 1 s reduced time at 180 km from SP04 (Fig. 2a); (3) a short retrograde phase from SP05 to the south from 4 s reduced time at 225 km to 2 s reduced time at 260 km from SP04 (Fig. 2b); (4) a high-amplitude reverberative phase originating from SPS2 at 3–4 s reduced time between 30 and 110 km from SP04 (Fig. 2d). The analysis and interpretation of these phases is discussed below.

Where possible, phase identification was confirmed by checking the reciprocal travel times. This proved useful for phases a1, c and e between SP04, SP05 and SPS1, and phases d1 and d2 between SPS1 and SP04. Traveltime picks were obtained using the ZP package and were assigned the program’s default picking error (Table 2) based on the relative signal to noise ratio within a 0.25 s time window before and after each pick.

Travel-time modelling

Travel-time modelling was conducted using a combination of forward modelling and inversion using the RAYINVR code of Zelt & Smith (1992). The distance scale of the final model was obtained by fitting a least-squares best-fit straight line through the shot and station coordinates, choosing the origin at SP04. The picked record sections provided true time–distance values, which were used in the modelling. Because of the near linearity of the recording profile, the model errors resulting from this procedure are considered negligible.

The recording line length of 310 km with the greatest shot offset (SPS2) of 80 km results in a maximum error of c. 60 m in distance owing to curvature of the Earth. This would result in a travel-time error of c. 7.5 ms for those shots providing energy that travelled the full extent of the profile. This is also considered negligible in relation to the pick errors.

An initial model was obtained using trial and error forward modelling in a top-down layer stripping approach. This model was then refined via damped least-squares inversion to minimize the travel-time residuals. The subsequent model was again refined slightly, mainly regarding the vertical velocity gradients, based on a qualitative comparison of the record sections with synthetic seismograms (Fig. 2) calculated using the TRAMP code of Zelt & Forsyth (1994).

The model

The final velocity model (Fig. 4a) was derived from the major phases a–f described above. The model extends from −30 to 370 km including all four shots. The surface layers are constrained by both seismic modelling and a priori information. The depth of water in the English Channel beneath shotpoints SPS1 and SPS2 was obtained from Bamford et al. (1976). The layer beneath this, of average velocity 3.9 km s⁻¹, has a base of severe topography. It has been interpreted as the pre-Variscan basement surface, initially defined from a priori information (Kelly 2006), and subsequently modified to optimize the model travel times of rays passing through it. A shallow layer of average velocity 4.0 km s⁻¹ has been included beneath the Bristol Channel. A 1 km thick low-velocity layer of average velocity 4.15 km s⁻¹ is required beneath SP05. Beneath this to a depth of about 4.5 km below sea level, a layer occurs with a steep northern margin and average velocity 5.1 km s⁻¹. The surface extent of this layer conforms to the Late Palaeozoic and Mesozoic rocks of south Wales and immediately south of the Bristol Channel. Further to the north, another narrow low-velocity (average 4.1 km s⁻¹) unit is required about 25 km south of SP04, coinciding with the Berwyn Hills and lying just to the south of the Bala Fault. At the northern end of the profile a surface layer of average velocity 5.2 km s⁻¹ is required, although this velocity is unreversed and defined by only a small number of raypaths.

Beneath the surface layers is a single horizon extending across the whole of the model. It reaches the surface beneath the Palaeozoic Welsh Basin, where its velocity is 5.7 km s⁻¹. It has an average velocity of 5.8 km s⁻¹, the values decreasing from north to south from a maximum of about 70 km south of SP04. The base of this layer has been defined by a 1 km thick gradient layer to allow a reasonable match between real and synthetic amplitudes of arrivals either reflected or refracted across this boundary. Reflections have been identified from both SP04 and SP05 to the south. The layer base appears to define a synform

Table 2. Pick errors and modelling results

| Phase | Error (s) | n  | Per cent | $\tau_{RMS}$ (s) | $\chi^2$ |
|-------|-----------|----|----------|-----------------|--------|
| a₄    | 0.02      | 5  | 83       | 0.020           | 1.311  |
| a₀    | 0.03      | 71 | 97       | 0.078           | 10.156 |
| a₁    | 0.06      | 57 | 100      | 0.089           | 4.974  |
| b     | 0.06      | 14 | 88       | 0.073           | 2.081  |
| c     | 0.08      | 98 | 97       | 0.082           | 1.559  |
| d₁    | 0.08      | 80 | 99       | 0.100           | 3.683  |
| d₂    | 0.08      | 42 | 98       | 0.090           | 2.080  |
| e     | 0.09      | 61 | 87       | 0.102           | 3.863  |
| f     | 0.09      | 58 | 98       | 0.089           | 1.786  |
| All phases | 0.07 | 496 | 94 | 0.089 | 3.856 |

Error, average assigned traveltime error for that phase; n, number of picks used in the modelling; Per cent, percentage of picks used in the modelling; $\tau_{RMS}$ and $\chi^2$ are as defined by Zelt & Smith (1992).
beneath the Palaeozoic Welsh Basin, the southern margin of which, at about 100 km, is brought up steeply beneath the location at which the overlying layer has a steep northern margin. The layer base remains approximately constant to about 140 km, after which it deepens to about 13.5 km at 260 km, rising southwards beneath the English Channel.

A second intracrustal reflector identified across the whole model at an average depth of 23 km defines the base of a mid-crustal layer. The layer has an average velocity of 6.35 km s$^{-1}$, which decreases markedly from north to south from about 6.5 km s$^{-1}$ beneath the Palaeozoic Welsh Basin to about 6.1 km s$^{-1}$ beneath the English Channel. Its base is deepest beneath the region of severe layer boundary gradients in the upper crust; however, this is in the region of least reflected ray coverage. It shallows to the south beneath the English Channel.

The Moho reflection is observed from all shots. Once again, the coverage is least in the central part of the profile, between about 105 and 180 km. It defines the base of the lowest crustal layer at a depth of about 36 km below sea level beneath SP04, deepening to 38 km beneath SP05, before shallowing to the south to about 30 km beneath SPS1. The Moho topography follows that of the overlying horizon. The velocity in the lowest crustal layer decreases severely from north to south from 7.25 km s$^{-1}$ beneath the Palaeozoic Welsh Basin to 6.65 km s$^{-1}$ beneath the southern part of the profile, with an average value of 6.9 km s$^{-1}$.

The mantle velocity immediately beneath the Moho has an average value of 8.0 km s$^{-1}$, once again decreasing slightly from north to south. A further sub-crustal layer has been modelled about 10 km beneath the Moho. It has been identified from an apparent higher velocity long offset arrival seen from shots SPS1 and SPS2. A possible reflection from the top of this layer originates from SP05 (see section on anomalous phases below). From SPS2 the equivalent reflection would occur almost exactly where the anomalous arrival between about 100 and 120 km from SP04 occurs, whereas from SPS1, it would occur close to the end of the record section where the signal to noise ratio is small. The average velocity at the top of this layer is 8.2 km s$^{-1}$.

**Model resolution**

Although the data quality is good, the distribution of shots and recording stations is sparse in comparison with present practice. Typically an equivalent onshore survey today (e.g. Mackenzie et al. 2005) has shots at a nominal 50 km and recording stations at a nominal 1 km spacing. Because of this the model resolution on
LISP DELTA has been examined in various ways and its effect on critical final model parameter values evaluated.

(1) The parameter values output by RAYINVR (Zelt & Smith 1992), n the number of rays traced, $T_{RMS}$ the RMS travel-time residual, and $\chi^2$ the normalized form of the misfit parameter, have been derived for each phase and the model as a whole (Table 2). In most cases the percentage of picks modelled is greater than 90% of the total number of travel-time picks, and the final $T_{RMS}$ value is nearly equivalent to the average phase uncertainty, indicating that the data have not been over-modelled. The $\chi^2$ values are relatively high. However, to reduce them, it would be necessary to increase the number of nodes, which would then result in a reduction in the parameter resolution. The very large $\chi^2$ value for the a0 phase is not unexpected. High values are often observed for this phase (e.g. Zelt & Smith 1992; Mackenzie et al. 2005) and result most probably from small-scale heterogeneities in the near surface. Zelt & Smith (1992) also suggested that other factors may contribute.

(2) The ray-path distribution including reflection point ranges on each horizon sampled by the profile has been plotted to provide a graphic image of the subsurface coverage (Fig. 5). As for all such surveys, the extremities of the model are unsampled, as is most of the unrecorded region between shots SPS1 and SPS2. There is also a poorly sampled region beneath the central shot SP05. As a result the model between SP04 and SPS1 is that taken forward to the gravity modelling.

(3) The diagonal elements of the resolution matrix have been overplotted on the final model (Fig. 4b). Zelt & Smith (1992) suggested that parameter values associated with resolution matrix diagonal elements greater than 0.5 are well resolved and reliable, although an ‘arbitrary cut-off of satisfactory resolution [may be] inappropriate’. In the present study the values are constrained owing to the sparse data coverage. They are dependent amongst other constraints on the range of change in velocity and depth node values allowed during inversion. These were set at 0.1 km s$^{-1}$ and 1 km respectively. The depth node values then average at 0.66 in the crust and the velocity node values average at 0.42. Although these values provide an indication of the relative confidence with which various parts of the model can be viewed, it was felt that examination of the absolute uncertainty of the crustal layer depths and the velocity values in particular locations was also necessary.

(4) Absolute uncertainty of the parameter values has been assessed by perturbing the values until the point at which the model so obtained is unable to fit the observed data as well as the final model. The stopping criteria involve (1) the ability to trace approximately the same number of rays through that part of the model and (2) the result of an F-test at the 95% confidence level comparing the $\chi^2$ values of the perturbed and original final model (Zelt & Smith 1992). Layer boundary and velocity nodes have been perturbed separately.

Layer boundary nodes. For the layer boundary nodes, the relative variation in node depth along a boundary is kept constant while the absolute depth of the layer boundary is perturbed. Results for the four principal layer boundaries (base upper crust, base middle crust, Moho and sub-Moho) show that the average depths for the four layers lie within an upper to lower bound range of 2.9, 1.6, 2.1 and 6.1 km respectively. A critical section of the 5.7 km s$^{-1}$ boundary (see Fig. 4a) beneath the structural low between 260 and 290 km was also perturbed to determine the minimum depth within the defined criteria. The average depth is 3.7 km; the minimum depth is 2.8 km.

Velocity nodes. Velocity values were examined for the principal crustal layers and in areas crucial to both the geological interpretation of the model and the conversion of the seismic velocity values to density for the modelling of available gravity data. The average layer velocities are well constrained, the uncertainties increasing with increasing depth from $\pm 0.03$ km s$^{-1}$ for the upper crustal layer to $\pm 0.11$ km s$^{-1}$ for the model base mantle layer. Six further regions were also examined comprising: (1) the high-velocity region in the lower crust between c. –30 and 130 km; (2) the upper crust immediately beneath the surface layer between 245 and 300 km; and the following four regions, which were examined for the upper bound on velocity, the derived value in each case being less than expected from subsequent gravity modelling: (3) the surface layer beneath SP05 between 120 and 155 km; (4) the surface unit between 30 and 37 km adjoining the Bala Fault; (5) the synformal layer beneath SP05 between 100 and 205 km; (6) the surface layer between 215 and 305 km.

The high-velocity lower crust section (1) beneath the northern end of the profile is well constrained (the upper to lower bound
range ($\Delta V$) is 0.22 km s$^{-1}$, as are the deeper regions (2) and (5). For (2) the range $\Delta V$ is 0.12 km s$^{-1}$ whereas for (5) the upper bound range ($\Delta V^+$) is 0.2 km s$^{-1}$. The velocity for the surface layer beneath the southern end of the profile (6) is also reasonably well constrained, $\Delta V^+$ being 0.2 km s$^{-1}$. However, the velocities for the surface layers (3) and (4) are very poorly constrained, $\Delta V^+$ being 2.1 and 2.2 km s$^{-1}$ respectively, as expected from the small number of raypaths sampling these two regions. The correspondence between these velocity bounds and the contoured diagonal elements of the resolution matrix (Fig. 4b) suggests that the latter can be accepted as a satisfactory guide to the overall resolution distribution of the model.

Anomalous phases

Four anomalous phases described above provide further information about the deep structure without affecting the parameter values of the final velocity model.

(1) Faults identified from SPS1 and SPS2. Significantly curved phases are observed at or immediately behind the first arrival between 100 and 120 km from SP04. The phases are observed most clearly from SPS1, but also seen from SPS2. Two vertical faults have been modelled at 104 and 116 km with diffractions ray-traced from 5 and 10 km depth. The correspondence to the observed data from SPS1 is good (Fig. 6). To provide the optimum image, the diffraction arrival times have been delayed by 0.4 s behind the first arrival. The fact that an exactly equivalent but fainter arrival is fitted by exactly the same model for the shot from SPS2 strongly supports this interpretation. The diffraction delay is not well constrained. Should it exist, it could result from a number of causes; for example, the non-2D nature of the problem, a fault intersecting the profile at some angle, small-scale velocity heterogeneity around the fault, or velocity anisotropy in the near-surface layers. The identification of two sources for the curved phase is also not well constrained. There may be a band of diffraction sources between these two distances. Finally, the maximum depth of the diffraction source is also uncertain. It has arbitrarily been taken to be 10 km, as this seems to provide a reasonable fit between the observed and modelled data. From the strength of the anomalous curved phases, it would seem to be a major crustal feature and therefore may well penetrate to even greater depths. Nevertheless, it may also be limited to the depth of the major sub-horizontal boundary occurring at about 8 km in the vicinity of the fault.

(2) High-frequency near-linear second arrival from SP04. This phase emerges from the Moho reflection PmP (phase c) at about 110 km. As this distance lies within the two faults identified in (1) above, it is thought likely that they may also be the cause of this phase. Two diffractions were produced again from 5 and 10 km depth, but originating from the PmP phase from SP04 at a distance once again of 104 km (the equivalent arrival at 116 km has not been modelled, to reduce the ‘complexity’ of Fig. 7). At the same time, the diffraction was modelled originating from the head wave (phase a$_1$) from SP04, produced in this case from 5 and 8 km depth, the latter value owing to the fact that the modelled headwave could not penetrate deeper than this. It can be seen that there is very good correlation between the PmP diffraction from 10 km depth and the high-frequency near-linear second arrival propagating to 1 s reduced time at 180 km. Similarly, there is a surprisingly good correlation between the equivalent a$_1$ diffraction from 8 km depth and a phase emerging at about 130 km immediately behind the first arrival. It is considered that these observations enhance the argument that, although the precise values of depth and position cannot be defined from the present data, there is a major sub-vertical crustal fault zone at about 110 km model distance, with the conjecture being that two major faults exist penetrating to at least 8–10 km at about 104 and 116 km from SP04.

Fig. 6. Ray trace modelling of vertical ‘fault planes’ at 104 and 116 km model distance for first arrival phases from SPS1 and SPS2. (a) Raytrace plot for rays emerging from both SPS1 and SPS2, diffraction points at 5 and 10 km depth. Only the diffractor at 104 km is shown, to reduce figure complexity. (b) Trace normalized reduced time section (bandpass filter 2–20 Hz; reducing velocity 6 km s$^{-1}$) and associated modelled travel-time curves for arrivals from SPS1. (c) Trace normalized reduced time section (bandpass filter 2–20 Hz; reducing velocity 6 km s$^{-1}$) and associated modelled travel-time curves for arrivals from SPS2. In (b) and (c) diff 5 and diff 10 refer to diffraction point depths for phases shown.
Short retrograde phase from SP05 to the south. This phase occurs behind the modelled PmP (phase c) from SP05 between 100 and 130 km modelled distance. It is not of large amplitude, but neither is the modelled PmP (Fig. 2b). Considerable modelling was undertaken to determine if this phase could be PmP itself. However, the model fit was significantly reduced if it was so interpreted, owing primarily to the well-defined velocities obtained for the lower crust and Moho derived from the long offset dividing phases f and d1 seen from SP04, SPS1 and SPS2. Interpreting it as a reflection from the mantle layer beneath the Moho (identified from phase d1) provided a reasonably good fit to the phase arrival times (Fig. 8).

Long offset reverberation from SPS2. The long offset (260–340 km from the shotpoint), high-amplitude reverberation from SPS2 is not seen from any of the other shots, but none of them are observed at such great distances. A surprisingly good arrival time fit is obtained for this phase from a Moho reflection multiple from this shot (Fig. 9). The fact that the highest amplitude appears to coincide with the multiple of the critical distance Moho reflection is strongly supportive of this interpretation. Not only that, but this interpretation would also be supportive of the lateral variation in crustal thickness in the final model. It is also of interest that this anomalou phase once again emerges at about 110 km, and it could therefore be enhanced by the superimposition of diffracted energy from the faults identified between 116 and 104 km modelled distance.

Gravity modelling

Gravity data along DELTA have been obtained from the British Geological Survey UK database (British Geological Survey 2006). Bouguer anomaly values together with station elevations were obtained at 1.5 km spacing along the profile. The data show a general decrease in value from north to south, the overall change over the model being c. 40 mGal (Fig. 10). This trend is surprising in the light of the seismic model showing a decrease in crustal thickness from north to south. The profile, as for that in northern Britain (Barton 1992) is also characterized by a number of small-amplitude (±10 mGal) anomalies, here with wavelengths of 10–50 km. The lateral resolution of the seismic data is limited by the small number of shots and relatively large recording station spacing, and it is not the intention here to attempt to model all such short-wavelength anomalies.

Gravity modelling was conducted using the 2.5D gravity and magnetic modelling software GRAVMAG (Pedley et al. 1993). Because of the very poor resolution between S1 and S2, a density model was derived from the seismic model between 30 and 300 km and linearly extended to ±5000 km, the model depth being set to 70 km. The gravity model surface layers were derived from the interfaces defined by the seismic model, whereas those below the first continuous horizon across the extent of the seismic model were obtained from a contour plot of the velocities below this level. These were contoured at a grid spacing of 2 km and velocity interval of 0.2 km s^{-1}. Densities were then assigned to each 0.2 km s^{-1} interval polygon. The contouring resulted in each of the seismic model crustal interfaces being replicated exactly in the gravity model.

The ‘Initial’ model (see Fig. 10) was derived using densities converted from velocity using mean values from the Nafe–Drake relationship (Ludwig et al. 1970) as specified by Barton (1986).
There are two significant mismatches between the calculated and observed values (Fig. 10), the mean difference between the two profiles over the model range being 2.20 ± 16.51 mGal. First, the obvious short-wavelength discrepancy between 120 and 200 km, together with those at c. 220 and 270 km appear to correlate almost exactly with the presence of near-surface low-velocity structures, which have initially been assigned low densities. There are other minor short-wavelength differences between the two curves. Second, whereas the fit of the calculated curve is reasonably close to the observed data at the north end of the profile, it is higher at the south end. To improve this longer wavelength discrepancy the densities of the mid- and lower crustal layer block at the southern end of the line have been decreased by an amount lying within the 95% confidence limit derived from the seismic velocities of these two layers (i.e. −0.03 and −0.02 Mg m$^{-3}$ respectively). Also, analysis of a crustal refraction–wide-angle reflection investigation between northern Brittany and the Channel Islands 85 km to the south of SPS2 (Grandjean et al. 2001) showed that the crust thickens to the south of LISPB DELTA. The Moho thickness was therefore allowed to increase to 30 km at a model distance of 455 km before being extended to 5000 km. The resulting ‘Intermediate’ model (Fig. 10) was then analysed in relation to the short-wavelength discrepancies discussed above.

To derive a satisfactory model providing a ‘good fit’ between observed and calculated Bouguer anomaly values it is necessary to modify the densities of the surface layers above the first continuous horizon across the model. These layers do not correspond exactly to different lithologies, but are amalgamations that provide a single bulk velocity (and therefore density) unit resolved by the seismic model. Given the correlation of these units with surface geology, it has been possible to identify density blocks associated primarily with: (1) Silurian lithologies at the northern end of the profile; (2) Ordovician–Silurian lithologies in the neighbourhood of the Bala Fault; (3) Cambro-Ordovician–Silurian lithologies between 40 and 100 km in the model; (4) Devonian and late Palaeozoic lithologies between 105 and 205 km; (5) Mesozoic lithologies from 225 km to the English Channel.

Griffiths (pers. comm.) in an unpublished study of the gravity field in Wales used densities for the sedimentary formations derived from Brooks (1960) and Gibb (1961). Examination of previous data for densities of Palaeozoic and Mesozoic rocks in Wales (including the results provided by Griffiths (pers. comm.))
Fig. 10. (a) Observed and calculated Bouguer anomaly values along the LISPB DELTA profile between ~30 and 300 km model distance, the calculated values being derived from the final seismic model. (b) Gravity model derived from the final seismic model (Fig. 4a). Model surface layers were derived from the interfaces defined by the seismic model, whereas those below the first continuous horizon across the model were obtained from a contour plot of the velocities below this level. These were contoured at a grid spacing of 2 km and velocity interval of 0.2 km s$^{-1}$. Densities in Mg m$^{-3}$. Model displayed to 60 km depth only.

Fig. 11. P-wave seismic velocity v. density.
and SW England and beyond suggests that preferred densities for these units lie in the ranges (Mg m$^{-3}$): (1) 2.6–2.75 (Griffiths, pers. comm.; Cook & Thirlaway 1955; Ford et al. 1991; Carruthers et al. 1992); (2) 2.6–2.8 (Griffiths, pers. comm.; Cook & Thirlaway 1955; Carruthers et al. 1992); (3) 2.6–2.8 (Griffiths, pers. comm.; Carruthers et al. 1992); (4) 2.5–2.73 (Griffiths, pers. comm.; Thomas & Brooks 1973; Brooks et al. 1977; Ford et al. 1991; Carruthers et al. 1992); (5) 2.3–2.65 (Brooks et al. 1977; Donato 1988; Carruthers et al. 1992).

We accept that these densities should be used with caution owing to their being derived both from rock samples (in some cases uncorrected for burial) and from the modelling of anomalies. In the first case they lack correction, and in the second they are model dependent. However, our model was modified allowing the densities of the relevant units to alter within these ranges. In every case, the revised density was greater than that derived from the seismic model.

Considering the velocity–density relation (Fig. 11), the revised values all lie within the bounds identified by Barton (1986), except those associated with the Cambrian–Silurian lithologies in the neighbourhood of the Bala Fault, and also those associated with the surface low-velocity unit beneath SP05. For both of these units the seismic velocity is very poorly resolved, the upper bound at the 95% confidence limit (see section on ‘Model resolution’) for the former being 6.3 km s$^{-1}$ and for the latter being 6.25 km s$^{-1}$, which would place them well within the bounds defined by Barton (1986).

With these modified surface layer densities and optimizing the densities of all those including and below the first continuous layer across the model (with no alteration of layer depth values) within the narrow standard deviation of $\pm 0.017$ Mg m$^{-3}$, the fit between observed and calculated gravity values is significantly improved. However, the short-wavelength anomalies at 220 and 270 km remain. To reduce these, it was found necessary to impose the effect of such a change on the final seismic velocity model (see ‘Model resolution’ above). In this region Chadwick et al. (1983) analysed seismic and gravity anomaly data along deep seismic reflection profiles c. 30 km to the east of profile DELTA and in a similar orientation with respect to geological strike. They suggested that there may be an increase in the sub-Mesozoic density of the Lower Palaeozoic lithologies from north to south, those in the north being shales of density likely range of 2.5–2.7 Mg m$^{-3}$ resulting from the presence of tectonized limestones, sandstones (quartzites) and mudstones. The required increase in density of the lower part of the two modelled synformal structures to 2.65 and 2.67 Mg m$^{-3}$ respectively could be related to this lithological change.

The final model results in the calculated ‘Final’ Bouguer anomaly (Fig. 10) having a mean difference from the observed curve of 0.14 ± 4.78 mGal.

### Interpretation

The LISPB profile crosses three major geological units: the Palaeozoic Welsh Basin in the north, the western part of the Avalonian Midland Platform in the centre and the Rheno-Hercynian Cornubian zone in the south. The seismic and gravity models reflect these differences, the present interpretation being constrained by known structures and physical properties of the underlying crust.

Previous geophysical work relevant to the modelling of LISPB DELTA has mainly occurred along the southern part of the profile in south Wales and SW England. Seismic velocities for the late Precambrian and Phanerozoic cover have been derived primarily from refraction profiles, but also from reflection seismic studies and velocity logging. Brooks et al. (1983) identified a widespread pattern of seismic layering, which together with subsequent work (Mechie & Brooks 1984; Brooks et al. 1994) can be interpreted in terms of geological layering. Using the same classification beneath the Tertiary and Mesozoic cover as Brooks et al. (1983), the geological interpretation of these P-wave velocities is as shown in Table 3.

It can be seen that units with seismic velocities of less than c. 4 km s$^{-1}$ are almost certain to be Mesozoic or younger, whereas those with velocities higher than c. 5.4 km s$^{-1}$ are of Silurian, Ordovician or Cambrian age, or older. Velocities over 6.0 km s$^{-1}$ can safely be attributed to the Precambrian metamorphic–igneous basement.

### Near-surface structure

**The Palaeozoic Welsh Basin.** The shallow (seismically poorly constrained) layer at the north end of the model with an average velocity of 5.2 km s$^{-1}$ is coincident with outcropping Silurian marginal basin lithologies of the Welsh Basin. Although the profile crosses equivalent rocks to the south of SP04 and to the north of the Devonian outcrop in mid-Wales, the large separation between the shotpoints has resulted in there being no associated first arrival. The velocity of 5.7 km s$^{-1}$ derived for the top of the

### Table 3. Refractor velocities and their geological interpretation in south Wales and the Cornubian zone of SW England

| Seismic layer | Refractor velocity range (km s$^{-1}$) | Geological interpretation                                                                 |
|---------------|----------------------------------------|------------------------------------------------------------------------------------------|
| 1             | 1.7–2.5                                | Tertiary                                                                                 |
| 2             | 2.6–4.1                                | Mesozoic                                                                                 |
| 3 (1)*        | 4.3–4.7                                | Upper Carboniferous                                                                      |
| 4 (2)*        | 5.1–5.3                                | Carboniferous Limestone Series                                                           |
| 4a (2a)*      | 5.6–5.7                                | Dolomitized Carboniferous Limestone Series                                               |
| 5 (3)*        | 4.6–4.8                                | Old Red Sandstone                                                                        |
| 5a            | 5.1                                    | Devonian (marine)                                                                        |
| 6 (4)*        | 5.4–5.8                                | Various layers within pre-Upper Palaeozoic supra-basement (Silurian, Ordovician, Cambrian) |
| 7 (5)*        | 5.8–6.3                                | Precambrian igneous–metamorphic basement                                                 |

*Layer numbers from Brooks et al. (1983).*
thick upper crustal layer between c. 10 and 100 km beneath the Palaeozoic Welsh Basin is likely to represent that of Ordovician metasediments and volcanic rocks that crop out to the south of SP04 within the basin. There is a significant reflection from the base of this layer at a depth of c. 12 km. Also, the underlying layer is raised at the southern margin of the basin at c. 100 km at the point where the LISPB profile intersects the Welsh Borderland Fault System in the vicinity of the Church Stretton Fault. This lies along the southern margin of the Welsh Basin where it abuts the Midland Platform, which strongly suggests that this thick upper crustal layer in the northern part of the profile does represent the Palaeozoic Welsh Basin. The large depth of c. 12 km correlates well with the maximum depth of ‘about 10 km’ proposed for the Welsh Basin by Carruthers et al. (1992) from their modelling of an east–west gravity profile across central Wales. The small surface trough included at about 30 km coincides with the region of Ordovician volcanic rocks and sediments of the Berwyn Hills immediately to the south of the Bala Fault. Although the very poorly constrained low-seismic velocity of 4.1 km s\(^{-1}\) may not be significant, it is consistent with the need for a marginally lower density for this unit (2690 kg m\(^{-3}\)) in relation to that of the surrounding Lower Palaeozoic surface rocks (2720 kg m\(^{-3}\)) to explain the small Bouguer anomaly low over this feature. Sparse seismic data and the large source distances have resulted in the Bala Fault, a possible crustal-scale feature, not being resolved in the model.

The Welsh Borderlands Fault System. The anomalous diffraction arrivals discussed above occur at 104 and 116 km from SP04. Those centred on 104 km coincide almost exactly with the trace of the Church Stretton Fault (Fig. 1c). In the vicinity of LISPB DELTA this lies c. 10 km to the south of the Pontesford Lineament, which has been identified as a ‘basin-margin basement fault zone’ (Woodcock 1984a) between the southern boundary of the Palaeozoic Welsh Basin and the northwestern margin of the Midland Platform. The Swansea Valley Disturbance is sub-parallel to the Church Stretton Fault c. 10 km further to the SE. Smith (1987) proposed that the latter, in the vicinity of the Longmynd c. 25 km to the east of LISPB DELTA, marks the Caledonian thrust front to the Welsh Caledonides, the fold belt being thrust eastwards over the Midland microcraton. However, the Pontesford Lineament and Church Stretton Fault were considered by Woodcock (1984a, b) to be strike-slip, vertical faults on the basis of observed lineations, fault patterns and the deformation style in the cover rocks of mid-Wales. Modelling of LISPB DELTA diffractions resulting from these faults (see the section ‘Anomalous phases’) strongly suggests that they result from such near-vertical faults beneath the seismic profile rather than ones dipping at c. 30° to the north as proposed to the east by Smith (1987). The location of the diffractions identified from SPS1 and SPS2 would suggest that it is the Church Stretton Fault and the Swansea Valley Disturbance that are the major crustal lineaments affecting the boundary between the Midland Platform and the Welsh Basin. Although it is accepted that the Church Stretton Fault is likely to be a deep, possibly crustal-scale feature, Blenkinsop et al. (1986) reported Woodcock (pers. comm.) stating that the linear course of the Swansea Valley Disturbance together with its sub-vertical outcrop also implies a deep planar fault, consistent with the results here. They also reported a concentration of seismic activity around Swansea, ‘suggesting a possible link with the Swansea Valley Disturbance’, and suggested that four large events may lie on this feature, whereas ‘no such patterns exist in the seismicity throughout the rest of SE Wales’. Although the results here do not deny the importance of the Pontesford Lineament, they also do not enhance it. The seismic phase modelling of the diffrations from these faults would suggest that they are sub-vertical. Blenkinsop et al. (1986) suggested that the trace of the Swansea Valley Disturbance may dip locally to the SE. Although certainly not proven here, this is consistent with the steep dip at c. 120 km on the base of the low-velocity (5.1 km s\(^{-1}\)) basinal structure occurring between 105 and 205 km. The base of the near-surface northern margin of the same low-velocity basinal structure rises to the north, emerging at the Church Stretton Fault. However, there is no evidence for a deepening of this horizon to the north of the Pontesford Lineament, and it is not possible to confirm the presence of a ‘horst’ between this and the Church Stretton Fault beneath the LISPB DELTA profile as proposed by Nunn (1978) and reported elsewhere (Edwards & Blundell 1984; Griffiths & Westbrook 1992).

The Midland Platform Cover. To the SE of the Welsh Borderland Fault System lies the Devonian Anglo-Welsh basin, which includes the Silurian Usk inlier. To the south, Triassic cover overlies Carboniferous rocks to the north of the Variscan Front (Fig. 1c). Beneath SP05 the very poorly resolved low velocities (average 4.15 km s\(^{-1}\)) at the surface overlie a broader basinal structure of average velocity 5.1 km s\(^{-1}\). The reasonable correlation of this latter value (also not well resolved; see Fig. 4b) with Old Red Sandstone velocities of 4.6–5.1 km s\(^{-1}\) (Table 3), and its surface geology primarily including Old Red Sandstone cover (together with the Silurian Usk inlier), suggests that it represents the distribution of Devonian and underlying Lower Palaeozoic rocks, its southern extremity including Carboniferous and Triassic cover around the Bristol Channel. The thickness of this basin fill is large, reaching a maximum of c. 4 km close to the proposed Church Stretton Fault. However, this is not inconsistent with Williams et al. (1982) and Friend et al. (2000) suggesting that the Anglo-Welsh basin preserves a relatively uniform fill of up to 2.4 km of Old Red Sandstone rocks in its central–eastern part (crossed by LISPB DELTA), and also Butler et al. (1997) identifying extensional basins underlying the Devonian cover; for example, the Usk Basin (within the Usk inlier) containing over 2 km of Silurian sediment.

Although the geometry model appears problematic in requiring a high density for this basin fill (2.69 Mg m\(^{-3}\)), Old Red Sandstone densities elsewhere in SW England (e.g. Al-Sadi (1967) (2.67–2.73 Mg m\(^{-3}\)) and Brooks et al. 1977 (between 2.62 and 2.65 Mg m\(^{-3}\)) are also very high for rocks with associated low seismic velocities (4.5–4.9 km s\(^{-1}\)). The previous interpretation of LISPB DELTA (Nunn 1978) also included a thick (c. 2 km) basal layer of lower velocity (4.5 km s\(^{-1}\)) in the vicinity of the Devonian outcrop near the centre of the profile.

An unreversed seismic refraction profile (Mechie & Brooks 1984) (Line J) along a SW-NE line from the Bristol Channel ending in the Silurian Usk inlier (Fig. 1c) intersected LISPB DELTA. At the intersection point (see Fig. 4a) the present model includes a layer of 5.85 km s\(^{-1}\) at a depth of 4 km. This may be compared with the Mechie & Brooks (1984) model of Line J, which they interpreted in terms of lower Palaeozoic–Precambrian supra-basement rocks above Precambrian crystalline basement \((v_p = 6.0\ \text{km} \text{s}^{-1})\) at a depth of c. 6 km. This small difference in depth could arise from a plethora of causes, resulting from both the experiment geometries and the complex crustal structure.

Mesozoic. Mesozoic rocks including those of probable Permian age conceal the older rocks to the south of the Devonian Anglo-
Welsh basin to the English Channel. Whereas the Cretaceous rocks are fairly flat-lying, the Jurassic sequence is seen to change in thickness in the vicinity of major growth faults, and the Permo-Triassic sequence is very variable in thickness and facies (Chadwick et al. 1983). LISPB DELTA, affected by the poor resolution provided by wide-angle data in the top few kilometres of the velocity model, provides little or no control on the internal structure and distribution of these lithologies, the base Variscan sequence in the model being initially defined from a priori information (Kelly 2006). However, with relatively minor modification, this model surface and the overlying layer velocity of 3.9 km s\(^{-1}\), consistent with Mesozoic lithology velocities, provide a good fit between the relevant observed and calculated arrival times.

The Variscan Front, the Bristol Channel Fault Zone and sub-Variscan geology

LISPB DELTA intersects the Variscan Front to the south of the Bristol Channel in the vicinity of the Mendip Hills (c. 195 km in model coordinates). It is not, perhaps unsurprisingly, identified in the model, as (1) it is only a deformation front, and (2) it is once again a poorly resolved surface feature on the LISPB DELTA profile, lying approximately halfway between SP05 and SPS1.

The Sub-Mesozoic geology is extremely complex in this region, and not readily appropriate for examination by refraction seismology. Not only are the lithologies almost certainly highly folded and faulted, but also the Lower Palaeozoic rocks and the Precambrian supra-basement rocks have very similar velocities. Identifying the ‘boundary’ between the Precambrian supracrustal rocks and the underlying crystalline basement is subject to a high degree of uncertainty.

Chadwick et al. (1983) have derived a crustal model from a set of deep seismic reflection profiles sub-parallel and c. 40 km to the east of LISPB DELTA in the vicinity of the Mendips and including the Variscan Front. They interpreted their model in terms of Mesozoic over intensely deformed Palaeozoic lithologies, these thickening dramatically to the south and affected by major thrusts and folds ascribed to the Variscan orogeny. Beneath there is a thick (up to 11 km thick in the north, thinning to the south) sub-base Cambrian sequence, which they suggested may be identified with the late Proterozoic sediments and volcanic rocks of the Midland microcraton. Beneath this they identified reflectors originating from crystalline basement rocks of the upper crust. They indicated that the southern part of their section may overlie Armorican basement, but were unable to identify a clear boundary between this and the Avalonian basement to the north. However, beneath LISPB DELTA any such deep reflectors originate from within the Rheno-Hercynian Cornubian zone, whether in place or transferred 400 km from the east as proposed by Woodcock et al. (2007).

The LISPB DELTA model defined here identifies a refactor at a depth of between 9 and 14 km beneath the Mesozoic sequence at the southern end of the model. The depth of Chadwick et al. (1983) proposed crystalline basement horizon is consistent with this. Above this, LISPB DELTA defines a layer of velocity between c. 5.7 and 6 km s\(^{-1}\), which is interpreted to comprise the thick folded and faulted sequence of Palaeozoic and supra-crustal Precambrian lithologies, but which, once again owing to the resolution of the method, cannot be separated. The thickening of this sequence south of the Variscan Front at c. 195 km is once again consistent with that identified by Chadwick et al. (1983), which they suggested results primarily from thickening of the Palaeozoic rocks caught up in the folding and thrusting of the Variscan orogeny. The model does not resolve an eastward extension of the Bristol Channel Fault Zone, although the synformal nature of the upper crustal layer to the south may be related to the presence of this proposed major crustal feature.

LISPB DELTA intersects a second profile (Line I) of the Mooney & Prodehl (1978) survey (Fig. 4a). At the intersection point the present model has a layer of continuous increasing velocity to 6 km s\(^{-1}\) at 11 km; that is, it apparently does not identify a boundary at the Lower Palaeozoic–supra-basement Precambrian divide as proposed by Mooney & Prodehl (1984). This would suggest that this boundary lies somewhere within the increasing velocity layer, with the deeper horizon being the crystalline basement layer proposed by Chadwick et al. (1983). However, once again the complexity of structure within the Variscan deformation front, the presence of the Avalonian–Rheno-Hercynian Cornubian zone boundary, and the different geometry of the two profiles, one of which is unreversed, could all combine to cause the apparent discrepancy between the two models.

Upper crust

The layer above the continuous refactor across the model at a depth between c.8-14 km beneath both the Welsh Basin and the Variscides is therefore likely to represent lithologies of different age along its length. From north to south, it is interpreted as follows.

1. To the north of the Welsh Borderland Fault System is the Palaeozoic Welsh Basin, possibly including at its base late Proterozoic supra-crustal sediments and volcanic rocks of the Avalonian terrane. Between the Welsh Borderland Fault System and the Variscan Front are Lower Palaeozoic and late Proterozoic supra-crustal Avalonian sediments and volcanic rocks beneath Devonian and other Palaeozoic and Mesozoic rocks of south Wales and south of the Bristol Channel.

2. To the south of the Variscan Front is Pre-Variscan basement inclusive of strongly folded and faulted Palaeozoic and Late Proterozoic Avalonian and Rheno-Hercynian Cornubian zone supra-crustal lithologies beneath Mesozoic rocks.

Mid- and lower crust

With this presumed variation in layer lithology across the model, care has to be taken in assuming that the layer beneath the upper crust is in fact continuous. It is more likely that at the southern end it is correlated with Rheno-Hercynian Cornubian zone basement either in place or transferred some 400 km from the east (Woodcock et al. 2007), whereas the northern part is crystalline basement of the Avalonian terrane. The velocity of this layer decreases from c. 6.5 km s\(^{-1}\) in the north to c. 6.2 km s\(^{-1}\) beneath southern England. The former value is slightly higher than that for similar depths beneath the southern end of LISPB GAMMA (c. 6.3 km s\(^{-1}\) derived from Barton (1992)), an explanation for this being as discussed below for the lower crust. One might expect that results from Rheno-Hercynian crust in West Germany should be comparable with those from the Rheno-Hercynian Cornubian zone, and certainly so if Woodcock et al.’s (2007) hypothesis concerning the c. 400 km dextral offset on the Bristol Channel Fault Zone is correct.

Mooney & Prodehl (1978) derived an average crustal velocity (excluding sediments) of c. 6.2–6.3 km s\(^{-1}\) and an average crustal thickness of 28–29 km for a ‘Rheno-Hercynian crustal model’ to the east of the Rhenish Massif in West Germany.
Mechie \textit{et al.} (1983), from a subsequent experiment in the same region, derived ‘similar’ values. This average crustal thickness is equivalent to that of the Cornubian zone beneath LISPB DELTA (29–30 km). The average crustal velocity (excluding sediments) beneath southern England is c. 6.2 km s$^{-1}$; that is, also consistent with that derived for the Rhenoo-Hercynian in West Germany. The north–south crustal thinning and decrease in crustal velocity beneath LISPB DELTA are also consistent with results from fundamental mode Rayleigh wave dispersion studies (Meredith & Pearce 1991).

The present model cannot confirm the terrane boundary between Avalonian and Cornubian crust as proposed by Woodcock \textit{et al.} (2007) to occur at the Bristol Channel Fault Zone. This is unsurprising, owing to this feature being subject to Variscan compression before Mesozoic reactivation, which has almost certainly resulted in an extremely complex depth profile should such a boundary exist. Also, as commented on by Kelly \textit{et al.} (2007), terrane boundaries in the UK are not resolved by wide-angle profiling.

There is a decrease in the modelled velocity and density of the lower crustal layer from north to south, as there is for the mid-crustal layer. The density model (Fig. 10), in fact, also shows a small decrease in density at the north end of the lower crustal layer. However, this is not supported by the velocity model owing to a lack of ray coverage. The cause of the north to south decrease in velocity is not clear. It appears to be primarily a variation within the Avalonian terrane. It may be related to the presence of intrusive material beneath north and mid-Wales originating during the period of calc-alkaline and subsequent marginal basin volcanism during the Ordovician. However, regional studies of extinction patterns, the CSSP wide-angle profile and gravity data (Jones \textit{et al.} 2002; Al-Kindi \textit{et al.} 2003) and modelling of receiver functions (Tomlinson \textit{et al.} 2006) show that a region of elevated seismic velocities and densities beneath the Welsh Basin and extending into the East Irish Sea mirrors the general pattern of Palaeogene extinction. Hillis \textit{et al.} (2008) showed that part of the extinction must be the result of regional shortening. However, the pattern is consistent with early Tertiary lower crustal igneous intrusion and magmatic underplating resulting from an underlying mantle plume, which would produce a broad pattern of uplift (Tiley \textit{et al.} 2004). Igneous material (most probably in the form of sills in the lower crust; White \textit{et al.} 2004) would produce the region of elevated lower crustal velocities and densities observed in the modelled profiles presented here. The higher mid-crustal velocities in the same region and discussed above may also result from such intrusion.

The crustal thickening beneath the Palaeozoic Welsh Basin is at odds with the decrease in Bouguer anomaly from north to south. However, the high velocities in the mid- and lower crust beneath this region help to explain the gravity data. The maximum crustal thickening occurs beneath the southern part of the Welsh Basin where it abuts the Avalonian Midland Platform. It is possible that this thickening may be associated with the presence of the major strike-slip, possibly crustal-scale, sub-vertical faults of the Welsh Borderland Fault System at this boundary. However, alternatively it may be due to igneous intrusion within the lower crust. The thickening would be consistent with the Welsh basin region appearing to have been largely unaffected by the Mesozoic extension, which resulted in the Cardigan Bay and East Irish Sea basins to the north and west, and the Western Approaches, Bristol and English Channel Basins to the south, as observed on previous compilations of Moho depths around the UK, Ireland and surrounding seas (Dézes & Ziegler 2001; Kelly \textit{et al.} 2007).

### Mantle

The mantle beneath the crustal model provides both a Moho arrival and one from a deeper intramantle layer. The velocity of the sub-Moho material is near constant beneath the profile at 8.0 km s$^{-1}$. At a depth of c. 10 km beneath the Moho there appears to be a deeper layer. Arrivals from it (sampled by shots SPS1 and SPS2) are unresolved, its velocity is poorly resolved and its upper boundary may be merely a velocity gradient. However, a reflection from this depth has been modelled from SP05 to the south. Faber & Bamford (1979) noted that differences in travel times and amplitude characteristics of mantle phases identified on the northern LISPB profiles qualitatively suggested lateral variations in the stratification of the lithosphere at similar depths to those here. They suggested that a group of phases both to the north and south of the Iapetus Suture zone may be reflected from horizons at depths between c. 55 and 45 km respectively, overlying material of velocity 8.2 km s$^{-1}$, equivalent to that observed beneath LISPB DELTA. Subsequent tomographic studies (Goes \textit{et al.} 2000; Arrowsmith \textit{et al.} 2005) are difficult to reconcile with such models, but this could result primarily from the different methods used. Although the continuity of this sub-crustal layer cannot be proven with the present data, its presence is consistent with other UK results for which similar methods were used.

The schematic illustration of Figure 12 obtained for the major part from LISPB DELTA, but including structures (e.g. the Bala Fault and the Bristol Channel Fault Zone) discussed above and derived from elsewhere, shows our preferred crustal model beneath the profile.

### Crust–mantle loading

The Bouguer anomaly profile, as already shown, is anti-correlated with crustal thickness along LISPB DELTA. To examine the effect of loading in the crust and mantle beneath the profile, the load was calculated across the model at 5 km depths in the crust to a depth of 40 km and also at 70 km (the latter at the base of the model). The loads were calculated at 5 km intervals along the profile. As the model incorporates the surface topography with the best-fit values of density for all layers, isostatic compensation of the model occurs at the level at which the load is constant across the model.

At 70 km the lithostatic load averages 2.16 ± 0.007 GPa. A null standard deviation would indicate that the model is in perfect isostatic equilibrium. Barton (1992), in an equivalent discussion, suggested that uncertainties in the LISPB model for northern Britain are large enough to accommodate the difference between observed and predicted values for a chosen parameter (topography) used to examine isostatic compensation. Barton stated that ‘The probable uncertainties in the calculations are fairly considerable’.

### Lithostatic load variations throughout the crustal model

Lithostatic load variations throughout the crustal model are shown in Figure 13. The average lithostatic load in the crust has been calculated at each 5 km depth across the model and subtracted from each value at that depth. The resulting plot (Fig. 13a) is contoured at 5 MPa intervals and shows the pattern of
load above and below the mean value with depth. Also calculated is the absolute value of the horizontal gradient of the load at each 5 km depth to the base of the crust throughout the model (Fig. 13b).

It can be seen (Fig. 13a) that there is a long-wavelength change from high values beneath the Precambrian and Palaeozoic geology in the north to low values beneath the Mesozoic and younger geology of SW England. Superimposed on this variation are major shorter wavelength increases beneath the topographic loads of the Berwyn Hills in north Wales, the Black Mountains in mid-Wales and the Mendips to the south of the Bristol Channel. The maximum values of \(c. 20\) MPa above the mean value at depth appear to concentrate in the lower crust. There is a significant lateral variation resulting in lower values throughout the crust beneath the central part of the Palaeozoic Welsh Basin. The largest changes in the horizontal gradient of the load occur beneath the region including the Welsh Borderland Fault System, and also beneath the Mendip Hills to the south of the Bristol Channel.

The likelihood of this distribution of load causing rock failure and earthquakes is dependent on the regional stress in the crust and its proximity to the critical state. However, given that the vertical loading in the crust is greatest in the northern part of the profile, if the crust is in a critical state, it is here that earthquakes with a predominantly tensional mechanism are likely to occur. The location of the maxima in the horizontal gradient of the load are likely to define those regions where predominantly strike-slip earthquakes occur.

Heidbach et al. (2007) have produced a stress map for Europe showing the maximum horizontal stress derived from earthquake focal mechanisms, well bore breakouts and drilling-induced fractures, \textit{in situ} stress measurements and young geological data. There is sparse coverage over Wales and SW England; however, the orientation in this region is approximately NW–SE. Hardwick (2008) in a study of over 1000 earthquakes occurring in the southern British Isles in the past 25 years, has relocated and examined the focal mechanism of 185 events in Wales and central and northern England and refined this orientation estimate, but it still remains approximately NW–SE along LISPB DELTA. All events occurring within \(15\) km of LISPB DELTA (from Hardwick 2008) have been overplotted on the sections of Figure 13. Although accepting that earthquakes are likely to occur where the crust is already cut by pre-existing zones of weakness and faults, the correspondence between the load distribution and the hypocentral distribution is striking.

The earthquakes occur mainly within the northern part of the profile, where the load values are predominantly above the mean value. They also cluster beneath the principal topographic loads. Hardwick (2008) identified a predominant transtensional mechanism for events occurring in the vicinity of LISPB DELTA consistent with the load distribution determined here, and in particular near the Welsh Borderland Fault System, where the horizontal gradient is large.

The lack of events beneath the southern part of the profile other than in the vicinity of the Mendip Hills, in correlating with the region of load variation significantly below the mean value would suggest that this contributes to the paucity of earthquakes here. The events occurring beneath the Mendips are consistent with this being a region where once again the horizontal gradient of the load is large.

We suggest that there is strong evidence that as well as possible zones of weakness, the distribution of topographic and subsurface load in the crust beneath LISPB DELTA controls the distribution of earthquakes beneath the profile.
Heat flow

A simple heat-flow profile for the onshore section of LISPB DELTA between 0 and 300 km has been derived from Lee et al. (1987) (Fig. 14). The values are low (c. 50 mW m\(^{-2}\)) and at the extremities can be seen to be about 10–15 mW m\(^{-2}\) greater than in the centre. The base of the seismogenic zone (the cut-out depth), below which only a small percentage of crustal earthquakes occur, is at or near the brittle–ductile transition. This transition should be temperature, strain rate, lithology and stress-state dependent. Although the heat-flow variation as well as the number of earthquakes beneath LISPB DELTA is small, it is of interest to note that there appears to be a simple inverse correlation between heat flow and focal depth along the profile. A number of studies (e.g. Chen & Molnar 1983; Bonner et al. 2003; Tanaka 2004) have examined heat flow and its relation to the base of the seismogenic zone. Bonner et al. (2003), from a study in California, showed that in regions of low heat flow, the cut-out depth temperature is at its lowest. Where the heat flow is near or below 50 mW m\(^{-2}\) (as along the LISPB DELTA profile) the cut-out temperature is 260 ± 40° C.

A simple crustal temperature distribution, sampled at 5 km spacing, was calculated between 0 and 300 km solving the 1D heat conduction equation for a three-layered model of the crust. The topmost layer included those units with seismic velocity less than 5.2 km s\(^{-1}\); that is, the Mesozoic basin at the southern end of the profile, the broad Old Red Sandstone and possibly earlier Palaeozoic basin in the central section of LISPB DELTA, and the two shallower units at the north end of the profile. The second layer is the upper crustal layer continuous across the model, and the third includes the two layers above the Moho.

The heat production values for these three layers were initially estimated using the empirical relationship between radiogenic heat production and P-wave seismic velocity proposed by Rybach & Buntebarth (1982, 1984)

\[
\ln A = 12.6 - 2.17V_p
\]

where \(A\) is the heat production in \(\mu W \text{ m}^{-3}\) and \(V_p\) is the seismic velocity in km s\(^{-1}\). Average P-wave velocities were derived for these three layers every 5 km across the model and converted into heat production values. Those for the Mesozoic basin at the southern end of the model were substantially greater than realistic and were replaced by a value of 2 \(\mu W \text{ m}^{-3}\), which was estimated from a study of the thermal blanketing effect of sediments (Zhang 1993) and of radiogenic heat production in sedimentary rocks (McKenna & Sharp 1998).

Using a basal heat-flow value \(Q_m\) of 30 mW m\(^{-2}\) and an average thermal conductivity throughout the model of 3.35 W
m\(^{-1}\) K\(^{-1}\) (e.g. Turcotte & Schubert 2002) we obtained the surface heat-flow profile across the model (Fig. 14a).

It can be seen that there is a major excursion in the central part of the profile corresponding approximately to the Old Red Sandstone and earlier Palaeozoic basin. A smaller excursion occurs over the small unit close to the Bala Fault. The simplicity of this modelling does not warrant a full optimization, rather we have included a value of \(1 \times 10^{-3} \text{ W m}^{-2}\) for the heat production values for these units, which is approximately that derived for sandstones (McKenna & Sharp 1998). This simple substitution immediately improves the fit between observed and calculated heat-flow values in these regions.

There remain two discrepancies, as follows.

1. At the southern end of the profile there is an excursion that occurs approximately over the Jurassic and Cretaceous outcrop in southern England. The short-wavelength nature suggests that it is likely to originate in the upper crust. Lee et al. (1987) suggested that secondary anomalies superimposed on the major anomalies derived from varying crustal heat-flow production values and basal heat flow beneath the UK may be related to groundwater movement. They stated that this may be an important factor in the onshore basins, and it is therefore a possible cause of the present excursion occurring over this section of the profile, which lies over the Wessex Basin in southern England.

2. The major excursion of the modelled from the observed profile occurs in the north. Here the long wavelength of the discrepancy indicates that its source is deeper. We suggest that this source lies in the heat flow into the base of the crust. If this value is raised linearly to \(40 \text{ mW m}^{-2}\) between 90 km and the north end of the model, the resulting fit is much improved (Fig. 14a). The temperature at the base of the crust is in this case raised by \(c. 100^\circ \text{C}\) (see Fig. 14b).

There are two major corollaries.

1. It can be seen from Figure 14b that the temperature and earthquake distribution in the crust are consistent with results from elsewhere, even in very different tectonic environments (Bonner et al. 2003); namely, the focal depths of events in this region of low heat flow inversely correlate with that heat flow and occur above a low cut-out temperature of \(c. 250^\circ \text{C}\). Although not proven, they would also suggest agreement with the statement that ‘temperature is the dominant factor governing focal depth in the crust’ (Tanaka 2004).

2. Arrowsmith et al. (2005) provided a 3D model of P-wave velocity beneath the British Isles. They identified \(\pm 1.5\%\)
velocity anomalies at depths of 50–250 km, with a particularly significant low-velocity anomaly beneath the Irish Sea. They stated that for a typical pyrolite mantle composition, the observed velocity variations can be produced by a thermal anomaly of c. 200°C (Carmichael 1982). They suggested that the most plausible cause of these anomalies is thermal upwellings in the asthenosphere. Those at shallower depths (c. 50 km) were suggested as possibly being lithosphere that has been heated by the underlying asthenosphere. Arrowsmith et al. showed a north–south section immediately to the west of LISPB DELTA, which includes a major low-velocity zone at shallow depth that correlates with the region of increased mantle heat flow suggested by the present model. Examination of the LISPB DELTA profile on the horizontal slice through Arrowsmith et al.’s (2005) P-wave velocity model at a depth of 100 km (Fig. 14c) shows that the northern part of the profile traverses a region where the velocity anomalies are of the order of c. 0.8% below the mean. This is broadly consistent with the increase in temperature at the base of the crust resulting from the increased basal heat flow in this region suggested by the present model. That part to the south is characterized by higher mantle velocities (and therefore lower temperatures) and is again consistent with the present model. Arrowsmith et al. (2005) were not able to distinguish the hot upper mantle zones in their model as resulting from (1) part of a present convective system, or (2) hot material that became ponded beneath thin lithosphere, in this case representing the thermal remanence of a convective system that occurred at c. 60 Ma. Our thermal model focusing on the crust is consistent with theirs. However, our seismic P-wave model has been interpreted in terms of igneous material intruded into the lower crust beneath the northern part of LISPB DELTA. The normal mantle velocity of c. 8 km s⁻¹ immediately beneath the Moho suggests that in this locale there is now little evidence of abnormally hot mantle at this depth. This would strongly suggest that the second of their two hypotheses is the more likely.

Conclusions

(1) The LISPB DELTA final model, derived from wide-angle seismic data, is of a sub-horizontally layered structure that represents an extremely complex crust. From examination of the surface geology this includes two major orogenic zones, one being the northern margin of the Avalonian microcontinent involved in the Caledonian orogeny and the other being the deformation front of the Variscan orogeny overlying the southern margin of Eastern Avalonia, the Rheno-Hercynian Cornubian zone. Although the profile identifies lateral variations in seismic velocity within the various crustal layers, which we suggest may result from the different crustal constituents beneath these zones, we cannot identify precisely the deep boundaries between them should they exist, owing to the lateral resolution of the method and the observation scheme used.

(2) The Palaeozoic Welsh Basin can be identified in the model. Its base would appear to lie at a depth of c. 12 km beneath north Wales. To the south the basin abuts the Welsh Borderland Fault System at the Church Stretton Fault. This and the Swansea Valley Disturbance some 10 km to the south appear as well-defined diffractions on the LISPB DELTA seismic records, and we suggest that they are probably sub-vertical crustal-scale faults.

(3) There is a wide, low-velocity, c. 4 km thick, synformal unit (between 105 and 205 km in model coordinates) which includes primarily Devonian Old Red Sandstone lithologies at the surface. The density of this layer needs to be significantly higher than that derived from the seismic velocity using the familiar Nafe–Drake relationship (Ludwig et al. 1970) to explain the gravity profile along LISPB DELTA. However, equivalent Old Red Sandstone rocks in SW England have also previously been shown to have a similar velocity–density relationship.

(4) The Variscan Front lies just to the south of the Bristol Channel on the LISPB DELTA profile. It is to the north of the possible eastward extension of the Bristol Channel Fault Zone. It is unlikely that the simple sub-horizontal layering of the present model adequately represents the significant lateral change in crustal structure across this region. We suggest that the upper crustal layer to the south of the Palaeozoic Welsh Basin comprises Palaeozoic and late Proterozoic supracrustal sediments and volcanic rocks. These are likely to be Avalonian to the north of the Variscan Front and possibly Bristol Channel Fault Zone, but Rheno-Hercynian at some point to the south. Beneath this a continuous layer across the model represents Proterozoic crystalline basement. The average crustal velocity (c. 6.2 km s⁻¹) and thickness (c. 29–30 km) at the southern end of the model in the Rheno-Hercynian Cornubian zone are consistent with those of equivalent Rheno-Hercynian crust to the east of the Rhenish Massif in West Germany, and possibly from where this region is suggested to have been displaced by some 400 km by late Acadian or Variscan dextral shear (Woodcock et al. 2007).

(5) The lowest crustal layer is again marked by a significant decrease in seismic velocity from north to south. However, in this case the velocity beneath the northern end of the profile reaches >7.3 km s⁻¹. It is supposed that this identifies igneous intrusion into the lower crust occurring possibly partly in the Ordovician, but also, and most probably in the form of sills, derived from an upwelling plume associated with the early opening of the North Atlantic Ocean.

(6) Crustal thickness beneath the profile is greatest beneath north Wales and the Welsh Borderland Fault System. This thick crust results from both the presence of the intrusive material at the base of the crust beneath the Palaeozoic Welsh Basin and also, possibly, from the presence of the Welsh Borderland Fault System. This thickening is consistent with the lack of Mesozoic extension in the Welsh Basin. The crust thins to the south, this probably being associated with both the transition from Avalonian crust into the Rheno-Hercynian Cornubian zone and also Mesozoic extension across the Western Approaches to the English Channel.

(7) Study of the lithostatic stress derived from the LISPB DELTA density model suggests that the lithosphere beneath the profile is in near-isostatic equilibrium as elsewhere in the British Isles.

(8) Examination of deviations in lithostatic load in the crust from the horizontally averaged mean value at sampled depths demonstrates a long-wavelength decrease from north to south, from Avalonian crust into the Rheno-Hercynian Cornubian zone.

(9) Short-wavelength load excursions from the horizontal mean value at sampled depths through the crust are dominated by the topographic load. The absolute values of the horizontal gradient of the vertical load show large values beneath the region including the Welsh Borderland Fault System and also the Mendip Hills to the south of the Bristol Channel.

(10) These load variations are strongly correlated with the 2D distribution of earthquakes along LISPB DELTA and are consistent with their predominantly transtensional focal mechanisms. This suggests that the earthquake activity in Wales and southern England is controlled by the crustal structure and density distribution. This results from both Caledonian and Variscan signatures in the crust and also the presence of Cenozoic lower
crustral igneous intrusive rocks associated with the opening of the North Atlantic Ocean.

(11) Examination of heat flow along LISPB DELTA shows that although low (c. 50 mW m⁻²) and of small variation, it is correlated with earthquake focal depth. Crustal temperatures have been obtained from 1D conductive heat-flow modelling, heat production values being derived from the seismic velocity distribution. The focal depth variation along the profile correlates with temperature, the base of the seismic zone or cut-out depth occurring at c. 250°C, consistent with equivalent results from elsewhere with low heat flow.

(12) The model requires increased mantle heat flow beneath the northern end of the profile. This is consistent with previous studies of the mantle seismic velocity distribution beneath the British Isles (Arrowsmith et al. 2005) interpreted in terms of mantle temperature distribution. Our results, interpreted in terms of igneous intrusion into the lower crust in the region of mantle temperature distribution. The focal depth variation along the profile correlates have been obtained from 1D conductive heat-flow modelling, correlated with earthquake focal depth. Crustal temperatures from elsewhere with low heat flow.

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