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Nature of the Moho beneath the Scottish Highlands from a receiver function perspective

Jeanette Di Leo [a], Ian D. Bastow [b,*], George Helffrich [b]

[a] Institut für Geologie, Universität Würzburg, Pleicherwall 1, D-97070 Würzburg, Germany
[b] Department of Earth Sciences, University of Bristol, Bristol, BS8 1RJ, UK

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ABSTRACT

The geology of Scotland documents a protracted geological history from Precambrian basement formation through the Caledonian Orogeny and to the Tertiary opening of the Atlantic. A temporary deployment of 22 closely spaced broadband seismic stations in Scotland traverses many of the major terrane boundaries in the region and provides a unique opportunity to place spatial and temporal constraints on variations in crustal structure. With teleseismic *P*-wave receiver functions, we use the $H-\kappa$ method to determine variations in bulk crustal parameters: crustal thickness (H) and V_P/V_S ratio. Mean crustal thickness is ~28 km, varying from ~23 km in the NE highlands and increasing to >30 km near the Highland Boundary Faultin the southern part of the study area. Mean V_P/V_S values of ~1.76 show no significant variation across the study area. An abrupt increase in crustal thickness of ~4.5 km NW across the Moine Thrust is not easily linked to Tertiary–Recent tectonic activity. Instead it appears that Scotland's crust has retained features for hundreds of millions of years since it was first formed, despite abundant volcanic activity in Tertiary times. Our work favours the view that the Moho is a compositional boundary rather than a mineralogical one defined by a reaction in *P*–*T* space.

1. Introduction

The prominent Mohorovičić (Moho) discontinuity separates the crust and mantle, and is readily investigated seismically. Thrusting and strike-slip faulting potentially reorganises the Moho by thickening the crust or by laterally transporting Moho architecture from different deformation regimes. Changes in Moho depths are observed in extensional settings such as western United States metamorphic core complexes (Myers and Beck, 1994) as well as in strike-slip environments (Faber and Bamford, 1981; McGeary, 1987; Lemiszki and Brown, 1988; McGeary, 1989; Blundell, 1990; Klemperer and Hurich, 1990; Klemperer et al., 1991; Jones et al., 1992). Cook (2002) suggests that the Moho will remain frozen and retain its primary features until affected by significant subsequent re-heating. Yet in some regions the Moho seems to be mutable, adopting a surprisingly flat morphology following continental collision (Meissner et al., 1991), which would suggest that rather than being a compositional boundary, the Moho is a type of facies boundary, which, like its metamorphic analogue, can be affected by changes in pressure, temperature or chemical environment. Another mechanism by which the Moho can be flattened is by gravitational collapse: excess gravitational potential energy drives crustal material flow away from the thickened orogenic syntaxis (Costa and Rey, 1995; Rey et al., 2001). Alternatively, some authors support the notion of a recycling of parts of the lower crust into the mantle through foundering of eclogites at the base of the crust (Arndt, 1989; Jarchow and Thompson, 1989; Laubscher, 1990; Austrheim, 1991; Nelson, 1991; Zandt et al., 2004).

The Moho is a first order velocity discontinuity where *P* wavespeeds increase abruptly to values ≥7.6 km/s (Giese, 2005). While easy to define, the discontinuity's origin is harder to explain and is a widely debated topic. A key observation that leads to more than one possible Moho origin theory is repetition of Moho-like discontinuities in the shallow mantle (McGeary and Warner, 1985; Snyder, 1991). Thus, some authors recognise a seismic Moho, wherein a transformation in mineral assemblage (gabbroic to eclogitic) due to a change in pressure and temperature, but at constant bulk composition, causes the discontinuity (Mengel and Kern, 1992). Other authors regard the Moho as a compositional boundary between mafic lower crust and peridotitic mantle and call it the petrological Moho (Mengel and Kern, 1992). Multiple origins provide a plausible explanation for the so-called W reflector in northern Scotland (McGeary and Warner, 1985) that lies below a shallower Moho. Snyder (1991) attributed the W reflector to the petrological Moho and the shallower discontinuity to the seismic Moho. Hynes and Snyder (1995) go on to posit the mutability of the petrologic Moho by invoking the possibility of granitoid melt extraction changing the bulk composition of the crust and the Moho's depth. If mutable, the process obviates the need to invoke foundering or crustal thinning by gravitational collapse to

* Corresponding author.
E-mail address: ian.bastow@bristol.ac.uk (I.D. Bastow).

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adjust Moho depth. If more than one process is required, however, why not many? In a recent summary of results from the Canadian LITHOPROBE programme, Eaton (2006) addresses numerous hypotheses for Moho formation and concludes that a multi-genetic origin for the Moho is likely.

The purpose of this study is to investigate the nature of the Moho in Scotland where within a compact, few hundred kilometre span, one encounters a geological record of both orogenesis and hotspot tectonism ranging in age from >2 Ga to ~60 Ma. We use receiver function analysis of teleseismic data from a network of closely spaced broadband stations to determine bulk crustal properties: thickness (H) and V_P/V_S ratio. We find that Moho depths do not vary greatly under northern Scotland, but feature distinct offsets at the Moine Thrust (MT) and along the Highland Boundary Fault (HBF), in accord with suggestions from active source studies (Barton, 1992; Chadwick and Pharaoh, 1998). These results show that the Moho is capable of

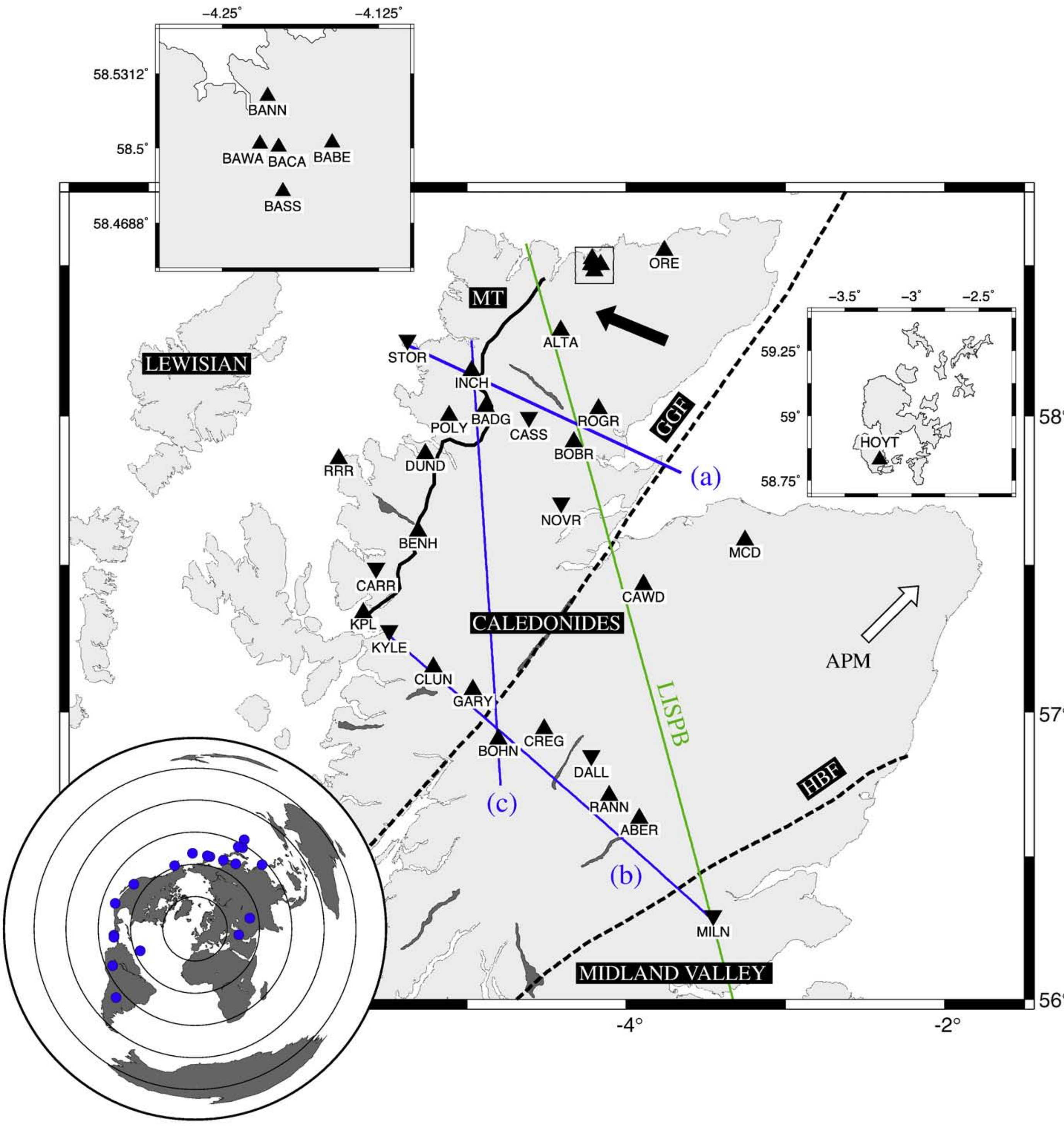

Fig. 1. Seismic recording stations and regional tectonic domains. Triangles are RUSH stations. Lines marked (a), (b) and (c) are the orientations of the cross-sections shown in Fig. 4a, b and c. MT, Moine Thrust; GGF, Great Glen Fault; HBF, Highland Boundary Fault. APM, Absolute Plate Motion (open arrow). The black arrow shows the WNW motion of material by thrust tectonics during the Silurian. The line marked LISPB shows the location of the LISPB wide-angle seismic profile (Bamford et al., 1976). Top left inset: stations BANN, BAWA, BACA, BABE and BASS, which have station spacing ~1 km. Right inset: station HOYT on the Orkney Islands, north of mainland Scotland. Bottom left inset: the distribution of earthquakes used in this study. Concentric circles, at 30° intervals from the centre of the RUSH are in azimuthal equidistant projection.

retaining abrupt level changes throughout Phanerozoic times, as well as through magmatic disturbances during the breakup of Pangea. Given that widespread eclogitisation of the lower crust and incorporation into the mantle would have eliminated these features, we conclude that in the Scottish Highlands the Moho is a compositional boundary.

2. Geology of Scotland

The Scottish Highlands record a long geological history ranging in time from Precambrian basement formation to Tertiary magmatism and rifting that marked the breakup of Pangea. The Highlands are composed of several geological terranes, juxtaposed by a series of tectonic events, most importantly the Caledonian Orogeny in Early Paleozoic time (Craig, 1991). In the far NW part of the study area (Fig. 1) Precambrian Lewisian (2.6–1.8 Ga) basement is cross-cut by the ~2.4 Ga E–W to NW–SE trending dolerite Scourie dyke complex. To the east of this region, the MT (Fig. 1) marks the northwesterly margin of the Caledonian (430–400 Ma) orogenic belt on mainland Scotland (Coward, 1983; Butler and Coward, 1984; Butler, 2004). The foreland comprises Lewisian gneissose basement, which is unconformably overlain by the Proterozoic sedimentary rocks of the Torridon Group (~0.9 Ga) and a cover of Cambro–Ordovician sediments that were thrust ~100 km in a WNW direction 437–408 Ma during the Silurian (Freeman et al., 1998; Butler, 2004), with little deformation occurring after ~430 Ma (Freeman et al., 1998). There remains a debate as to the nature of the MT, however. Smythe (1987) suggests that the Lewisian foreland of the NW Scottish mainland does not extend further eastwards below the Caledonian over-thrust belt for more than 20–30 km east of the MT. Soper and England (1995), on the other hand, suggest that the entirety of Scotland north of the Highland Boundary Fault (HBF, Fig. 1) formed on the continental margin of Laurentia. In this scenario, the MT would not be considered a terrane boundary. Unlike previous authors (Watson and Dunning, 1979; Soper and Barber, 1982), Butler and Coward (1984) consider the MT to be a relatively flat-lying, shallow structure, so whether it is a crustal-scale feature or not remains controversial.

In the centre of the study area, the Great Glen Fault (GGF, Fig. 1) separates the NW highlands from the Grampian mountains (Canning et al., 1998). The GGF was active during Caledonian times with initial sinistral shear movement dated at ca. 425–427 Ma (Rogers and Dunning, 1991; Stewart et al., 2001), lasting until ~390 Ma. The Walls Boundary Fault (WBF) is a crustal-scale strike-slip fault that cuts Precambrian–Caledonian basement terranes in Shetland and has been interpreted as the northern continuation of the GGF in Scotland (Watts et al., 2007). There are some similarities to the kinematic history of the faults, but the post-Triassic reactivation histories appear to differ significantly. In particular, the later strike-slip movements have opposite senses of shear:dextral along the GGF and sinistral along the WBF (Watts et al., 2007). Disparate structures on either side of the WBF support the idea of large scale lateral displacements, but a major vertical step in the Moho discontinuity cannot be substantiated from the seismic data (McBride, 1994a,b, 1995).

Across the GGF, Dalradian (Late Precambrian to Ordovician aged) sedimentary rocks extend to the HBF (Harris and Johnson, 1991) that was active during the Devonian (~350 Ma). The HBF is thought to be a terrane suture that lacks a major thrust component but does feature lateral displacement (Harris and Johnson, 1991). While the MT marks the northwesterly margin of the Caledonian orogenic belt on mainland Scotland (Butler and Coward, 1984; Coward, 1985; Butler, 2004), the HBF, separating the Grampian Mountains from the Midland Valley, marks the southerly margin of the metamorphic Caledonides. The last major event to affect Scotland was Tertiary (Paleocene–Eocene) magmatism from 61–55 Ma associated with the breakup of Pangea. Since then, Scotland has experienced no significant tectonic activity. For a more detailed overview of the geology of Scotland we refer to Craig (1991) and references therein.

3. Data and methods

Our data come from two temporary broadband seismological networks in Scotland (Fig. 1). RUSH-I (Reflections Under the Scottish Highlands) consisted of 9 broadband stations in northern Scotland that recorded continuously at 20 Hz for ~14 months in1999–2000 (Asencio et al., 2003). The 24-station RUSH-II experiment (Bastow et al., 2007) recorded between 2001 and 2003, completing a closely spaced (~20 km) network across major terrane boundaries in Scotland (Fig. 1). Thus, it is possible to resolve shorter length scale variations in bulk crustal structure using broadband seismological data than has been possible to date (Tomlinson et al., 2006) and allows tighter temporal constraints on variations in crustal structure to be developed.

In order to characterise crustal structure in Scotland, we use the Extended-Time Multitaper Frequency Domain Cross-Correlation Receiver-Function Estimation method (ETMTRF) (Helffrich, 2006) to compute 432 P-wave receiver functions (RFs) for 32 teleseismic earthquakes of magnitude mb $\geq$5.5 recorded by RUSH stations (Fig. 1). Data were filtered using a zero-phase Butterworth bandpass filter with corner frequencies of 0.01–0.4 Hz. The filters were designed to eliminate both high frequency cultural and long period microseismic noise. We then used the $H-\kappa$ stacking method (Zhu and Kanamori, 2000) to deduce bulk crustal properties (crustal thickness, H; V_P/V_S ratio, and κ).

Using the direct P-to-S conversion only (Ps at t_1), there is a trade-off between the determination of H and V_P/V_S (Ammon et al., 1990; Zandt et al., 1995): a change of 0.1 in κ can lead to a change of ~4 km in H (Zhu and Kanamori, 2000). To reduce this ambiguity, (Zhu and Kanamori, 2000) incorporate the later arriving multiple converted phases $PpPs$ (t_2) and $PpSs+PsPs$ (t_3). The arrival-times of the three arrivals (t_1,t_2,t_3) relative to the direct P-wave are defined as:

$$t_1 = H\left[\sqrt{\frac{1}{V_S^2}-p^2}-\sqrt{\frac{1}{V_P^2}-p^2}\right], \tag{1}$$

$$t_2 = H\left[\sqrt{\frac{1}{V_S^2}-p^2}+\sqrt{\frac{1}{V_P^2}-p^2}\right], \tag{2}$$

$$t_3 = 2H\sqrt{\frac{1}{V_S^2}-p^2}, \tag{3}$$

where p is the ray parameter of the incident wave. A flat Moho is assumed. The travel-times can be used to estimate crustal thickness

Table 1
Hypocentral information for earthquakes used in the $H-\kappa$ analysis. Z, earthquake depth.

Date	HH:MM:SS.sss	Lat. (°)	Lon. (°)	Z (km)
1999/10/16	09:57:00.009	34.71	−116.27	15
1999/11/08	16:53:00.040	36.48	70.81	237
1999/12/06	23:21:00.047	57.35	−154.35	54
2000/03/28	11:12:00.039	22.32	143.76	100
2000/04/23	09:39:00.021	−28.41	−63.04	608
2000/08/04	21:23:00.015	48.77	142.03	15
2000/08/06	07:38:00.010	28.89	139.68	411
2001/12/18	04:02:58.280	23.95	122.73	14
2002/03/03	12:08:19.740	36.50	70.48	226
2002/06/16	02:46:14.030	8.78	−83.99	35
2002/06/22	02:58:21.300	35.63	49.05	10
2002/06/28	17:19:30.270	43.75	130.67	566
2002/07/31	00:16:44.610	7.93	−82.79	10
2002/08/20	10:59:32.020	30.99	141.97	9
2002/10/12	20:09:11.460	−8.30	−71.74	534
2002/10/16	10:12:21.430	51.95	157.32	102
2003/01/22	02:06:34.610	18.77	−104.10	24
2003/03/15	19:41:28.700	52.25	160.39	30
2003/03/17	16:36:17.310	51.27	177.98	33
2003/05/14	06:03:35.860	18.27	−58.63	42

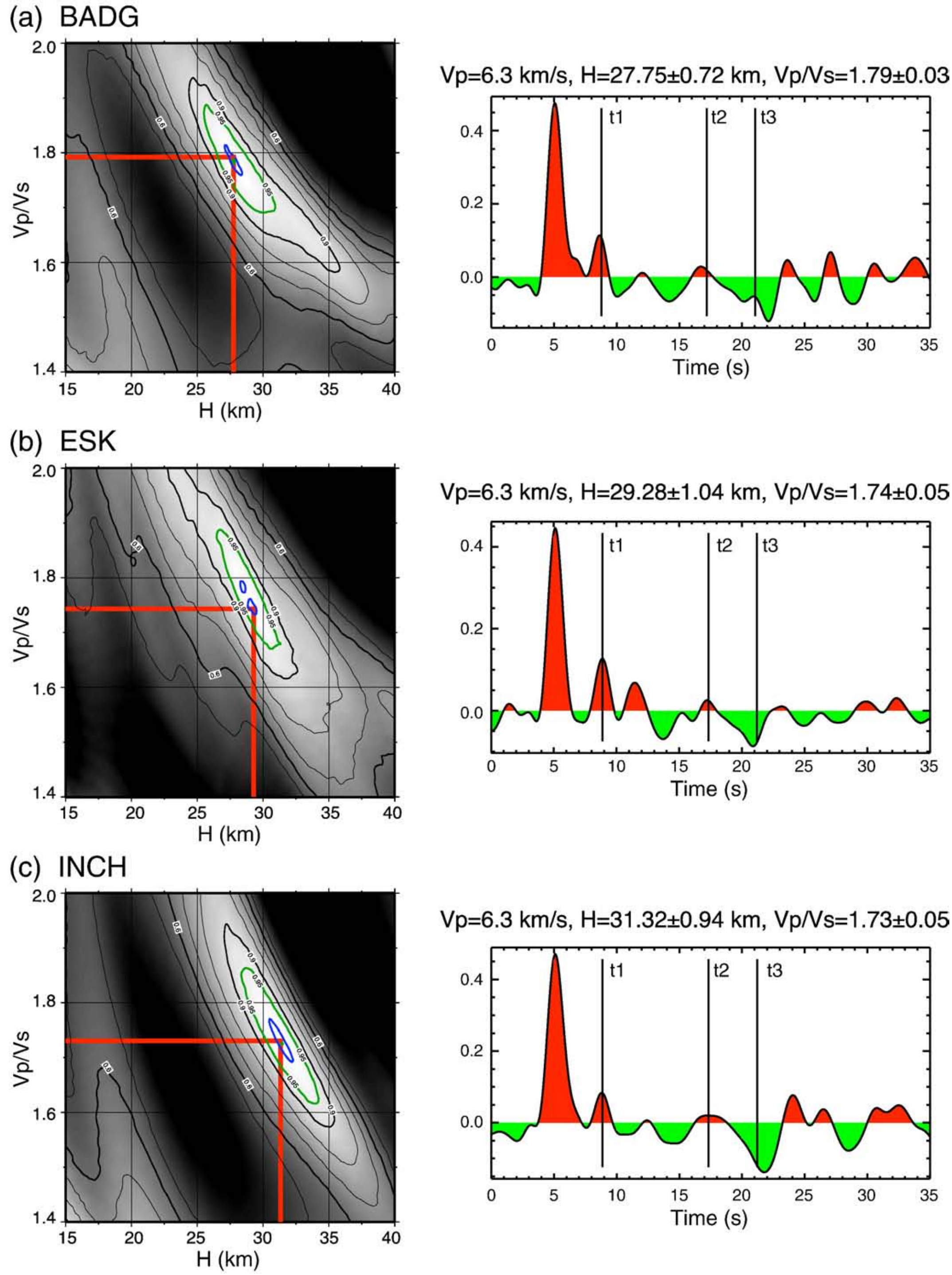

Fig. 2. Examples of crustal thickness (H) versus V_P/V_S plots from the method of Zhu and Kanamori (2000) and receiver functions for: (a) BADG, ESK, and INCH. On each plot the arrival-times of the Moho phase Ps (t_1) and subsequent reverberant phases $PpPs$ (t_2) and $PsPs+PpSs$ (t_3) (determined by Eqs. (1), (2) and (3) respectively) are marked based on the crustal thickness and V_P/V_S values shown in Table 2. The horizontal and vertical lines on the contour plots mark the maximum of the stack defined by Eq. (4).

(H). The stacking itself transforms the RFs directly into the $H-\kappa$ domain. This is achieved by summing the amplitudes at the predicted arrival-times for a range of $H-\kappa$ combinations with:

$$s(H,\kappa) = \sum_{j=1}^{N} w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3), \tag{4}$$

where $r_j(t)$ are the receiver function amplitude values at the predicted times. The negative sign on the $PpSs+PsPs$ arrival is due to the reversed polarity of this phase. w_1, w_2, and w_3 are weighting factors with $\sum w_i = 1$. Zhu and Kanamori (2000) recommend $w_1 > w_2 + w_3$, since Ps has the highest signal-to-noise ratio and should therefore be given a higher weight. They choose $w_1 = 0.7$, $w_2 = 0.2$, and $w_3 = 0.1$, but exact values vary from study to study; Dugda et al. (2005) use $w_1 = 0.6$, $w_2 = 0.3$, and $w_3 = 0.1$, for example. In this study, we use weights $w_1 = 0.4$, $w_2 = 0.3$, $w_3 = 0.3$, or $w_1 = 0.6$, $w_2 = 0.3$, $w_3 = 0.1$ in instances where $PpSs+PsPs$ is less clearly identifiable. $s(H,\kappa)$ then reaches its maximum for the correct H and V_P/V_S ratio.

We perform a grid-search of 2500 combinations of H and V_P/V_S (H = 15–40 km; V_P/V_S = 1.4–2.0) and assume a mean crustal P-wave velocity (V_P) of 6.3 km/s^{-1}. This mean crustal P-wave velocity comes from wide-angle seismic surveys (Edwards and Blundell, 1984; Barton, 1992; Jones et al., 1996; Grandjean et al., 2001; Al-Kindi et al., 2003) and has been used in earlier receiver function studies in the British Isles (Tomlinson et al., 2006).

We estimate the uncertainties in H and κ by following the method set out in Eaton (2006). We construct contours in $s(H-\kappa)$ that are one standard error below the maximum value, where the standard error is given by $(\sigma^2/N)^{1/2}$. Table 2 shows the results for H and κ with their associated errors.

Of the 432 computed RFs, our final $H-\kappa$ analysis uses 49 RFs from 20 of the 32 earthquakes (Fig. 3; Table 1), from which one or more reverberant phases (*PpPs* or *PpSs* + *PsPs*) were clearly identifiable. Unfortunately, many RFs were not of use during the analyses principally due to ocean microseisms that are inescapable in Scotland and which bandpass filtering cannot successfully remove (Restivo and Helffrich, 1999; Helffrich et al., 2003; Bastow et al., 2007). Examples of receiver functions and $H-\kappa$ stacks are shown in Fig. 2.

4. Results

Table 2 and Fig. 3 show the results of our $H-\kappa$ stacking analyses. Spatially overlapping results from the study of Tomlinson et al. (2006) are also presented. We find an average crustal thickness (H) of 27.89 ± 0.52 km, with individual station measurements ranging from 23.67 to 31.83 km. Average V_P/V_S ratios are ~1.76 ± 0.02, with range 1.68–2.02 across the RUSH network. To first order we do not observe discernible trends in V_P/V_S ratio, but robust variations in H exist.

Fig. 3 shows variations in H across major structural trends in Scotland: the MT, the GGF, and the HBF. We detect little variation in structure across the GGF (Fig. 4b and c), but H varies significantly along strike: from ~29 km at BOHN/GARY in the west to ~24 km at CAWD in the east (Fig. 3). A ~4 km thickening of the crust occurs E–W across the MT zone in a remarkably short lateral distance (~20 km; Fig. 4a and c). In the southern part of the study area, towards the HBF, the crust thickens gradually to ~30 km.

Table 2
Bulk crustal properties across the RUSH broadband network.

Station	H (km)	V_P/V_S	σ	t_1	n
ABER	31.32 ± 1.23	1.71 ± 0.07	0.24	3.61	1
ALTA	24.18 ± 0.64	1.69 ± 0.06	0.23	2.71	1
BABE	26.73 ± 0.42	1.76 ± 0.03	0.26	3.37	3
BACA	26.73 ± 0.49	1.76 ± 0.02	0.26	3.33	3
BADG	27.75 ± 0.72	1.79 ± 0.03	0.27	3.66	1
BANN	27.24 ± 0.32	1.75 ± 0.02	0.26	3.35	3
BASS	27.24 ± 0.40	1.73 ± 0.02	0.25	3.24	3
BAWA	27.24 ± 0.50	1.74 ± 0.03	0.25	3.30	5
BENH	27.75 ± 0.55	1.77 ± 0.03	0.27	3.57	1
BOBR	23.67 ± 1.65	1.73 ± 0.09	0.25	2.85	4
BOHN	29.28 ± 0.54	1.77 ± 0.04	0.27	3.69	1
CAWD	23.67 ± 0.60	1.77 ± 0.04	0.27	2.98	1
CLUN	30.30 ± 0.52	1.68 ± 0.02	0.23	3.39	4
CREG	29.28 ± 0.68	1.70 ± 0.04	0.24	3.39	3
DUND	26.73 ± 0.31	2.02 ± 0.03	0.34	4.44	3
ESK	29.28 ± 1.04	1.74 ± 0.05	0.25	3.60	2
GARY	28.77 ± 0.70	1.74 ± 0.05	0.25	3.52	2
HOYT	31.83 ± 0.32	1.68 ± 0.02	0.23	3.56	1
INCH	31.32 ± 0.94	1.73 ± 0.05	0.25	3.83	1
POLY	29.28 ± 0.52	1.79 ± 0.03	0.27	3.83	3
RANN	29.79 ± 0.76	1.86 ± 0.05	0.30	4.26	2
ROGR	24.18 ± 0.33	1.70 ± 0.03	0.24	2.75	1
Tomlinson et al. (2006)					
KPL	28.00 ± 0.96	1.75 ± 0.06	0.26	–	67
MCD	31.20 ± 1.45	1.76 ± 0.07	0.26	–	59
ORE	26.10 ± 1.20	1.74 ± 0.07	0.25	–	41
RRR	24.30 ± 1.08	1.75 ± 0.07	0.26	–	48

H, crustal thickness (km); t_1, *Ps* travel-time; n, number of earthquakes. Poisson's ratio, $\sigma = 0.5[(V_P/V_S)^2-1]^{-1}$. Results from spatially overlapping stations are also presented from Tomlinson et al. (2006), which used ~11 years of data from permanent stations in Scotland.

5. Discussion

Average V_P/V_S or Poisson's ratio, ($\sigma = 0.5[(V_P/V_S)^2-1]^{-1}$) can be used to complement petrological studies of crustal composition (Chevrot and van der Hilst, 2000). σ does not vary significantly with temperature, and in the absence of melt, mineralogy is the most important factor controlling it. The relative abundance of quartz (σ = 0.09) and plagioclaise feldspar (σ = 0.30) has a dominant effect (Christensen, 1996): for felsic quartz-rich rocks such as granite σ = 0.24; for intermediate rocks, σ = 0.27; and for mafic plagioclaise rich rocks such as gabbro σ = 0.30. The lack of systematic variations in V_P/V_S ratio across the major terrane boundaries in Scotland is consistent with the broader study of Tomlinson et al. (2006) and means that mapping compositional variations in the Scottish subsurface is somewhat difficult. The observed values are generally low however (mean = 1.75), precluding the possibility of partial melt beneath the region as suggested by Arrowsmith et al. (2005), for example.

Despite the lack of systematic variation in V_P/V_S across Scotland, variations in (H) are intriguing. The notion that sharp Moho topography is a transient phenomenon and cannot be observed in western Europe for tectonothermal provinces older than 300 Ma is held by some authors (Meissner et al., 1987). Meissner et al. (1987) believe that this may be due to extension related creep processes in the Earth's crust and upper mantle. Other studies, however, show results similar to ours in other parts of Europe. Thybo (1997) reports variations in Moho depth beneath the Tornquist Fan, a splay of late Carboniferous–early Permian fault zones in and around Denmark. The undulating Moho topography there shows a clear correlation with the major late Palaeozoic tectonic features of the region. Clowes et al. (1987) found a sharp, ~12 km Moho offset over a very short lateral distance (few tens of kilometres or less) beneath the Fennoscandian Shield in southern Sweden at the lithospheric boundary between the Småland-Värmland Granite Belt (~1.75–1.62 Ga) and the Svecofennides (~1.90–1.75 Ga). This shows that such pronounced Moho features can occur in regions even older than the Scottish Highlands and that these regions therefore do not seem to have undergone complete Moho depth re-equilibration by processes such as gravitational extensional collapse (Costa and Rey, 1995; Rey et al., 2001) and foundering (Meissner et al., 1991; Zandt et al., 2004).

Of the hypotheses described by Eaton (2006), the one of the relict Moho matches our results best: the Moho can freeze (Cook, 2002) at the base of pre-existing crustal blocks during continental assembly, and regardless of post-collisional processes retain its primary features unless significantly higher temperatures, such as those in extreme environments like volcanic regions or rifts occur.

Crustal structure in the British Isles has been studied extensively off-shore by numerous wide-angle seismic profiles (Chadwick and Pharaoh, 1998; Morgan et al., 2000; Price and Morgan, 2000; Kelly et al., 2007) that carry evidence for short wavelength changes in crustal thickness (Klemperer and Hurich, 1990). Detailed on-shore crustal information comes from the Lithospheric Seismic Profile in Britain (Bamford et al., 1976; Barton, 1992): a 700 km-long wide-angle seismic profile that ran approximately N–S through northern Britain (Fig. 1). Several features of the resulting crustal velocity model are reproduced in our results, which match the controlled source derived crustal thicknesses to within ± 3 km. In particular, there is an increase in crustal thickness of ~4.5 km across the MT (Figs. 3, 4a, and c). From N–S, there is a thickening of the crust to ~34 km towards the HBF (Figs. 3 and 4a). There is no variation in crustal thickness ($H \approx 29$ km) across the GGF (Figs. 3, 4b, and c). The receiver function studies of Tomlinson et al. (2003, 2006) also yielded structures closely

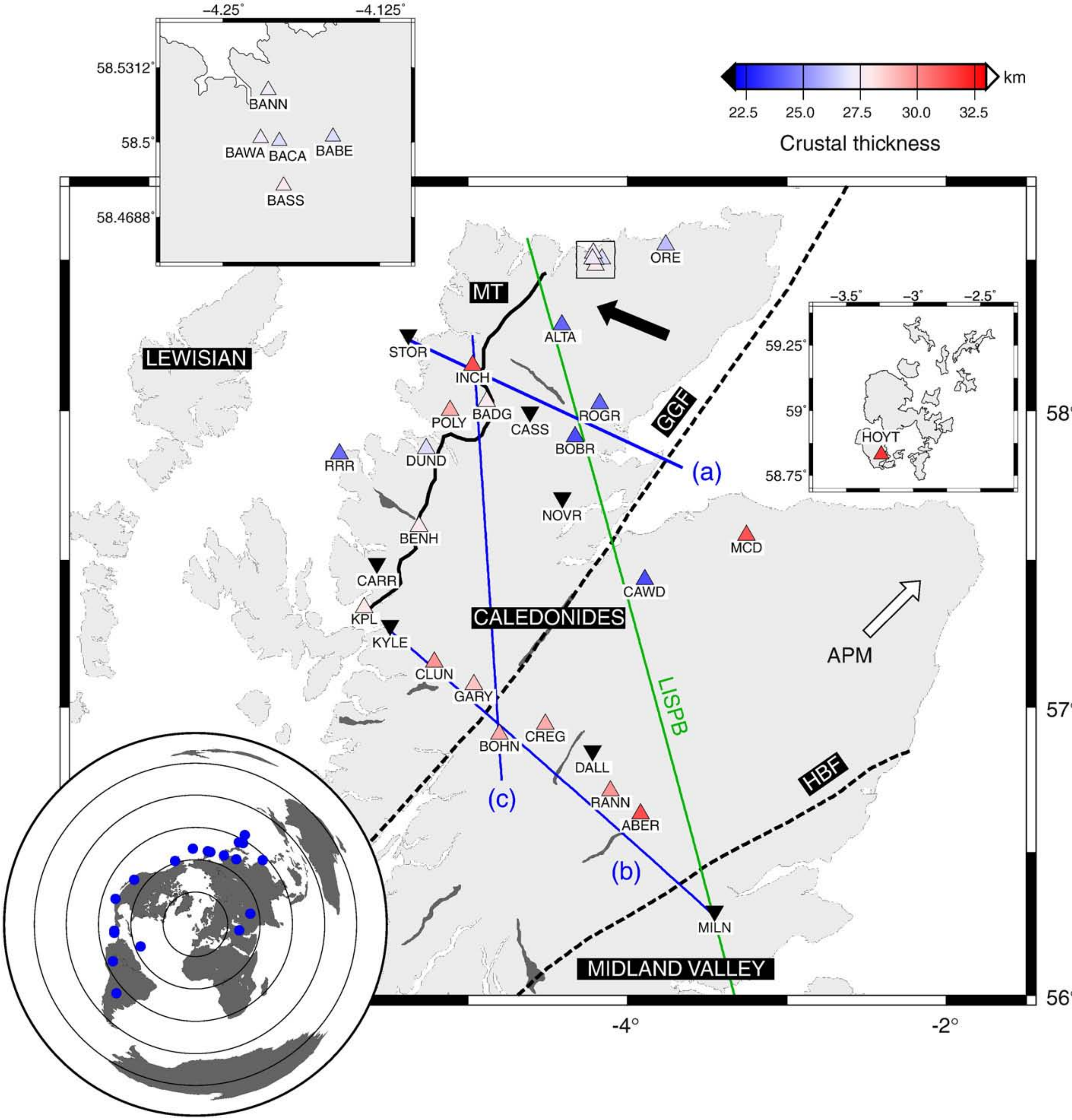

Fig. 3. Variations in crustal thickness across the Scottish Highlands determined from receiver function analysis. Inverted triangles are stations from which high quality receiver function analyses were not obtained. Bottom left inset: the distribution of earthquakes used in the $H-\kappa$ analysis. Concentric circles, at 30° intervals from the centre of the RUSH network are in azimuthal equidistant projection.

corresponding to the LISPB line, but our closely spaced stations make it possible to resolve lateral variations in crustal structure to wavelengths of ~20 km.

A simple equilibrium calculation shows that the observed Moho offset cannot simply be due to a change in pressure or temperature across the MT. Pressure cannot be constant across a 4.5 km depth change, which corresponds to about a 1 kb change in lithostatic pressure. This would constitute a driving force to transform a low- or high-pressure mineral assemblage to its opposite counterpart. A temperature contrast might also arise across two juxtaposed crustal blocks. If, however, a thermal contrast existed, it would decay significantly according to one dimensional conduction. The station spacing, 20 km, sets the horizontal length scale for diffusion. Using $l=\sqrt{Dt}$, with $l=20$ km and thermal diffusivity $D=1\ \mathrm{mm}^2/s$ (Turcotte and Schubert, 1982), then the time t for significant change in the thermal anomaly would be roughly 10 Ma. Since the last thermal event was over 55 Ma, the temperature contrast expected is negligible. Thus, it is unlikely that a pressure- or temperature-mediated mineralogical change occurs across structure below the MT. This suggests that the Moho offset demarks a chemical change.

Many studies of wide-angle data from the Highlands and off-shore regions infer the presence of magmatic underplate (Al-Kindi et al.,

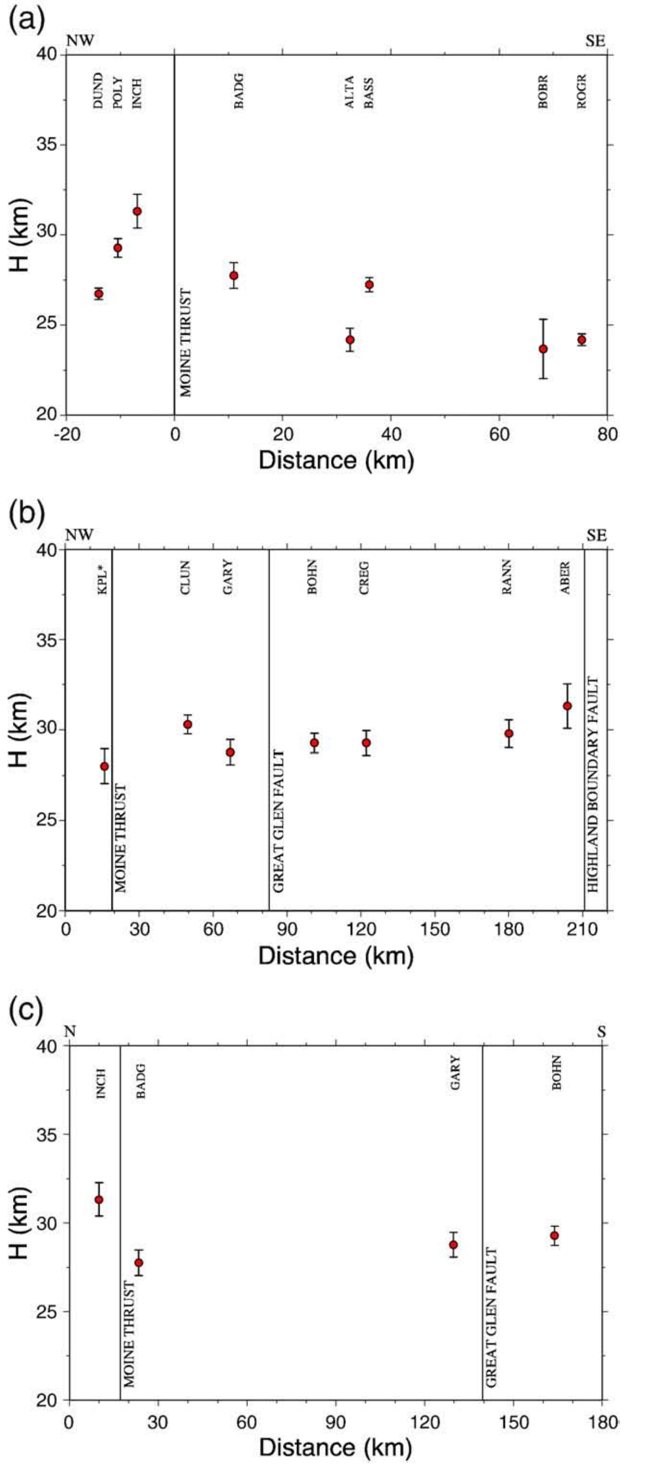

Fig. 4. Variations in crustal thickness along profiles (a), (b) and (c) on Figs. 1 and 3. * denotes a result from the study of Tomlinson et al. (2006).

2003). The underplate, which is up to 12 km thick off-shore Scotland, is believed to have formed as the result of lateral flow and ponding of Iceland plume material at the base of the lithosphere since ~65 Ma. By inference, a thicker crust would be predicted in NW Scotland compared to the south, and our observation of a ~4.5 km crustal thickening in the NW part of the study area (Figs. 3, 4a, and c) near the MT could therefore have Tertiary–Recent origins. Consistent with this hypothesis, the *P*-wave tomographic study of Arrowsmith et al. (2005) cites the presence of low velocities beneath the British Isles at ~100 km depth as evidence for the dominance of Tertiary volcanics on upper-mantle structure. The coincidence of the conjectured underplate's extent with the MT is peculiar, however, and in any case, recent work by Price and Morgan (2000) and Kelly et al. (2007), for example, does not provide clear evidence for high-velocity lower crust in the region that would be supportive of the underplate hypothesis.

Bastow et al. (2007) presented SKS shear-wave splitting results at RUSH stations. They found that asthenospheric fabrics due to Tertiary rifting and magmatism are not strongly expressed in the splitting observations. Instead, the results correlate with lithospheric scale trends inferred from surface geology such as the MT, GGF, and HBF. The shallow lithosphere beneath Scotland has apparently preserved a fossil anisotropic signature up to ~300 Ma after it was formed (Helffrich, 1995; Bastow et al., 2007).

Fresnel zone arguments (Alsina and Snieder, 1995) suggest that ~20 km wavelength lateral variations in seismic properties obtained from broadband data can be resolved. The ~3.5 km crustal thickening in ~20 km that we observe in crustal structure near the MT (stations INCH and BADG: Fig. 4a and c) is therefore real. Such abrupt variations in *H* cannot easily be explained by the underplate hypothesis and are more likely associated with shallower, more localised crustal deformation processes during the formation of the lower-crustal structure of the footwall of the MT. Our data do not favour the view that the entirety of Scotland north of the HBF (Fig. 1) formed on the continental margin of Laurentia (Soper and England, 1995) since our observations indicate a basement offset, at least beneath the MT. The nature of the basement of the Scottish Highlands, however, is complex because it was formed when Avalonia, Laurentia and Baltica collided during the Caledonian Orogeny (Lyngsie et al., 2006). Whatever the nature of the basement of the Scottish Highlands is, our results show that the Moho is capable of retaining level changes throughout Phanerozoic times, as well as through magmatic disturbances during the Tertiary breakup of Pangea. They favour the view that the Moho is

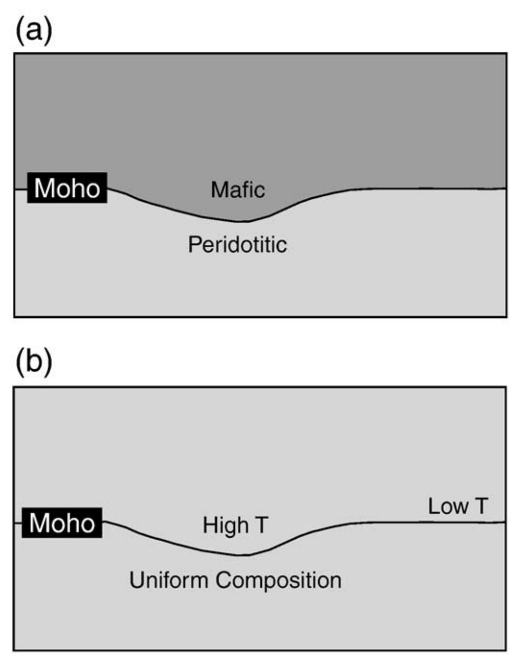

Fig. 5. Two cartoons showing possible explanations for the Moho offset across the Moine Thrust. (a) shows that the Moho might be a material boundary between two bulk compositions. Topography is due to undulations in the material boundary. (b) shows that the Moho might be a mineralogical reaction boundary in pressure–temperature space. At low temperature (*T*) the reaction progresses at shallower levels than at high temperature. Topography is due to lateral variations in lithospheric temperature. Our observations favour the compositional boundary model because significant lateral temperature variations cannot persist over the ~60 Ma since the last tectonic event.

a compositional boundary rather than a simple trajectory through *P*–*T* space (Fig. 5).

6. Conclusions

Our results, where spatially coincident, match the results of onshore wide-angle profiles to within ± 3 km. A thickening of the crust of ~4.5 km NW across the MT occurs over too short a length scale (~20 km) to be explained by laterally extensive magmatic underplate material inferred to lie beneath much of northern England, and the change in *H* is more likely linked to shallower lithospheric processes such as tectonic thickening by thrusting. Correlations with Caledonian age (~430 Ma) structural trends seem more appropriate in Scotland, consistent with studies of SKS shear wave splitting (Helffrich, 1995; Bastow et al., 2007). Thus, the Scottish crust appears to have retained structural features up to hundreds of millions of years after they were first formed, despite the abundance of magmatic activity during Tertiary times. This, in turn, argues for the Moho being a compositional boundary rather than a thermobarometric one (Fig. 5).

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