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Proto-Iceland plume caused thinning of Irish lithosphere

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Abstract

The lithosphere–asthenosphere boundary (LAB) separates the rigid lithospheric plates from the hot convective mantle. In the region of the North Atlantic this boundary is widely influenced by the spreading head of the proto-Iceland plume [P. Kumar et al., The lithosphere–asthenosphere boundary in the North-West Atlantic region, Earth Planet. Sci. Lett. 236 (2005) 249–257.] that also caused the evolution of the British Tertiary volcanic province (BTVP) with the Giant's Causeway basalts as famous landmark. Using teleseismic shear-wave receiver functions we map the LAB underneath Ireland and find a distinctive lithospheric thinning of about 30 km towards the northern part of the island. As this region is part of the BTVP, we explain the lithospheric thinning, in agreement with geochemical results, by thermal erosion originating from the head of the proto-Iceland plume.

Keywords: receiver functions; LAB; Ireland

1. Introduction

The concept of plate tectonics defines the lithosphere as the rigid outer layer of the Earth, which is divided into the crust and uppermost mantle. Its base, the lithosphere–asthenosphere boundary (LAB), is defined either mechanically in that the lithosphere floats on the mechanically weaker asthenosphere, thermally as the level above which heat is transferred mainly by conduction, geochemically by changes in the composition of the mantle-forming rock peridotite (for definitions see [2]), or seismically by the onset of a deep-seated low-velocity zone. Below the lithosphere, temperatures exceed the mantle solidus and visco-elastic creep, viscous flow or even partial melting occurs thus allowing convective heat transfer.

In the Palaeocene, during make-up of the North Atlantic region, the lithosphere–asthenosphere system there was widely influenced by the spreading head of the Iceland plume [3,4,1]. Thermomechanical processes during plume–lithosphere interactions are, for instance, demonstrated in that the dynamic pressure and support of the plume material, which floats within the uppermost asthenosphere, may uplift the overriding lithospheric plate [5]. In this case the lithospheric thickness remains relatively constant. Alternatively, when the plume material heats and erodes the lower lithosphere through melting [6], the lithosphere becomes thinner thus causing isostatic uplift [7].

Furthermore, basaltic melt extracted from the plume material may underplate the base of the lithosphere [8,9], increasing its overall thickness [10]. Although the

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influential role of the proto-Iceland plume on the evolution of the British Tertiary volcanic province (BTVP [11,12]) is confirmed by numerous geochemical observations [13,14,12], no direct seismic evidence for these plume–lithosphere interactions has yet been found beneath Ireland. In this study we determine the thickness of the Irish lithosphere (from teleseismic data) and interpret its anomalous structure through interaction with the proto-Iceland plume.

2. Experiment and motivation

From October 2002 until July 2003, a passive (teleseismic) network with 23 stations was deployed in southern Ireland (Fig. 1; [15]) — called Irish Seismological Lithospheric Experiment (ISLE). The aperture of the network was approximately 135 km in NE–SW direction and 105 km in NW–SE direction with an average station spacing of 30–50 km. Fifteen stations were equipped with broadband seismometers (filled circles in Fig. 1), the remaining eight with short-period seismometers (open circles). Additionally, data from two permanent broadband stations (DSB, VAL) were included. ISLE was deployed across the proposed surface trace of the late-Caledonian Iapetus suture zone in southern Ireland (ISZ; [16]). This suture is a result of the closure of the Iapetus ocean by End-Silurian times and subsequent collision of the Laurentian and Eastern Avalonian continents during the Caledonian orogeny. Large teleseismic P-wave traveltime residuals (0.5–0.7 s) were observed across the western ISZ, and Masson et al. [17] modelled a steeply south-dipping (compositional) boundary within the lithosphere (30–106 km depth) separating Laurentia and Eastern Avalonia.

Their model implies that considerable structural differences across the ISZ are still present at subcrustal levels. Since the number of teleseismic events used by Masson et al. [17] was very limited, ISLE has been designed to more clearly define the extent of the P-wave residual anomaly and to gain new insights into the nature of Caledonian deformation in Ireland by

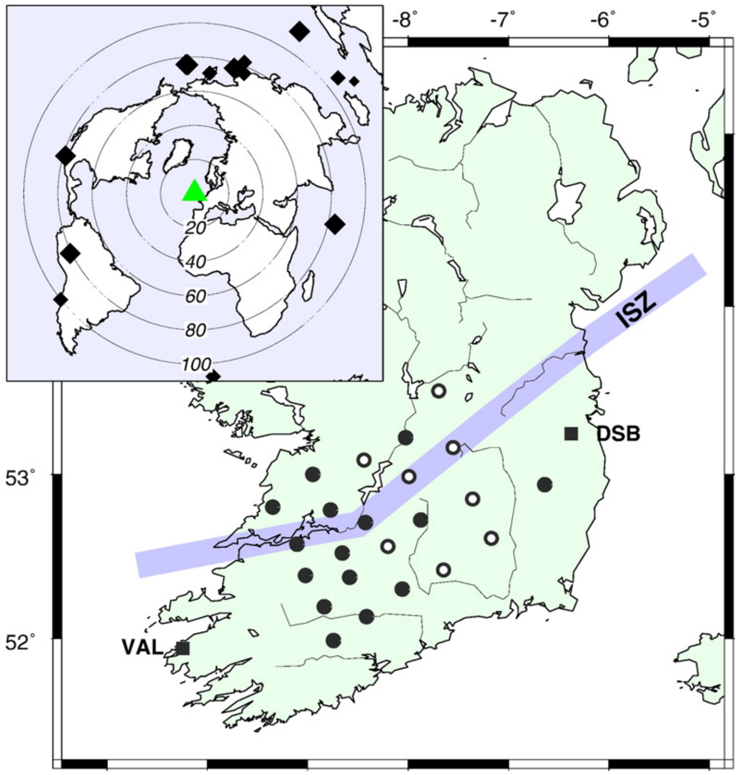

Fig. 1. Map of Ireland with temporary and permanent seismic networks. Open circles: short-period instruments, filled circles: broadband instruments, squares: permanent stations (DSB station is part of GEOFON network, GFZ Potsdam; VAL is operated by MET EIREANN). Also shown is the proposed surface trace of the Iapetus suture zone (ISZ, grey band [16]). Superimposed is an azimuthal projection, centred on southern Ireland, which comprises teleseismic event locations (diamonds) used in this study and recorded between November 2002 and July 2003. The moment magnitude of the 16 events varies from 5.5 to 7.6, while depths extend to 580 km. The size of the diamond is proportional to the magnitude of the teleseismic event. Circles mark epicentral distances of 20°, 40°, 60°, 80° and 100° from southern Ireland.

investigating its deep lithospheric and asthenospheric structure in more detail.

3. Measuring the thickness of Irish lithosphere

3.1. S receiver function technique

We examine the structure of the lithosphere beneath the ISLE network by the S-to-P receiver function analysis (Sp RF [1,19,20]). Sp Rfs are seismic waveforms that enhance compressional (P) wave signals which were converted from shear (S) waves at seismic discontinuities underneath the receivers. The Sp Rfs are processed in such a way that they are directly comparable with P-to-S (Ps) Rfs. In P Rfs [21], the time window of the Ps conversion from the lithosphere–asthenosphere boundary (LAB) is disturbed by crustal reverberations (multiples). These reverberations are not present in the S Rfs, because the Sp converted phases are S precursors and therefore do not interfere with the main incoming S phase or its multiples which arrive later. Therefore, Sp data have the great advantage in that they are free of disturbing multiples. Synthetic S Rfs are given by Yuan et al. [20]. In general, S Rfs are much noisier than P Rfs, because S waves arrive in the P-coda.

ISLE yielded high-quality earthquake waveform recordings. For this study, mantle (S) and core (SKS) shear waves with a sufficient signal-to-noise ratio could be selected from 16 teleseismic events of the ISLE dataset (inset in Fig. 1) at distances of 55–85° for S and 85–125° for SKS phases. All recordings were initially converted to a uniform data format [22], and thereafter restituted to true ground displacement in order to correct for the response function of the seismometer. Since the 8 short-period seismometers used have a natural frequency of about 1 Hz, the low frequency signals must be restored from the recordings (prior to calculation of the receiver functions) by simulation with the frequency response function of a 30 s broadband seismometer. The efficiency of this method was demonstrated by Niu et al. [23].

In order to separate the Sp conversions from S phases, we rotated the original ZNE (vertical, N–S, E–W) components into the ZRT (vertical, radial, transversal) co-ordinate system using the theoretical back-azimuth. Z, R and T are then rotated into the LQT

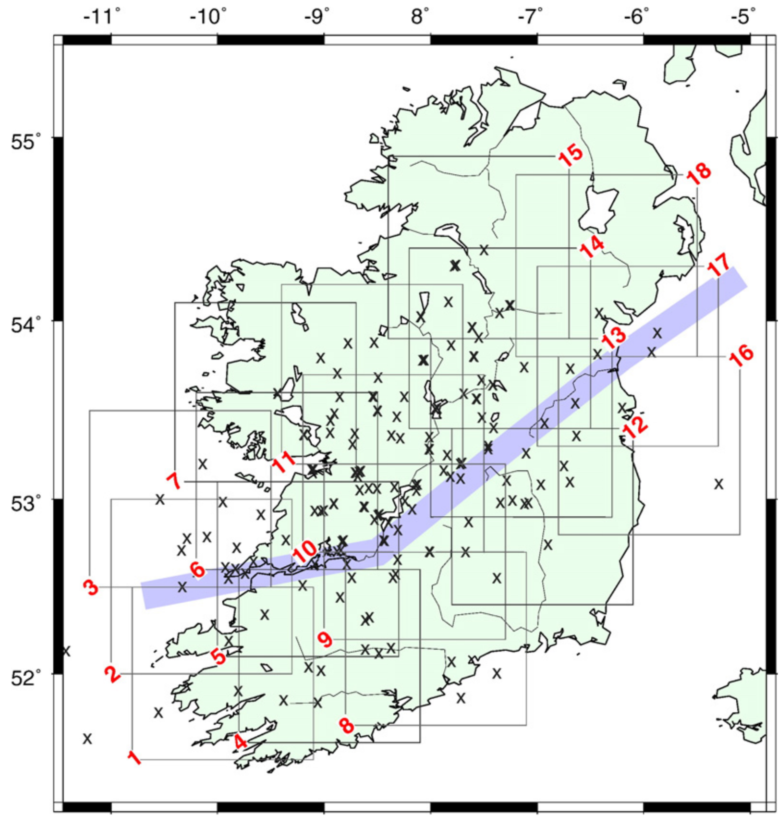

Fig. 2. Map of Ireland with black crosses indicating conversion points at 80 km depth used for stacking of S RFs within 18 rectangular sectors of size 1.7°×1°.

(longitudinal, vertically perpendicular, transversal) co-ordinate system, whereas the (true) angle of incidence is defined by the minimum energy in the L component at the arrival time of the S phase. Relatively large amplitudes of S Rfs at time 0 s are due to remaining S wave energy and usually result from imperfect co-ordinate rotation or scattered waves. A bandpass filter from 4 to 20 s was applied in order to suppress noise. The S source function was removed from the L component by deconvolution using a source time window of 100 s (10 s before and 90 s after the S wave). All seismic traces were moveout-corrected with a reference slowness of 6.4 s/° to enable the summation of recordings from events with different epicentral distances. This reference slowness is identical to that in (many) P RF studies. Crustal v_p and v_s velocities in the iasp91 model [24] were replaced by the ”true” values based on the local Irish 3-D velocity model [25].

Sp and SKSp conversions have different polarities. For easier comparison, we reversed the polarity of SKS RF amplitudes. Additionally, the time axis is reversed so that the Sp precursors then appear after S (and have positive delay times). Zero time is the S arrival time, where positive delay times indicate the period prior to the S arrival. As in P Rfs, positive (negative) amplitudes now indicate velocity contrasts with velocities increasing (decreasing) downwards [20].

3.2. Observed S receiver functions

The ray paths of P and S Rfs are very different, with the result that conversion points of S Rfs are situated

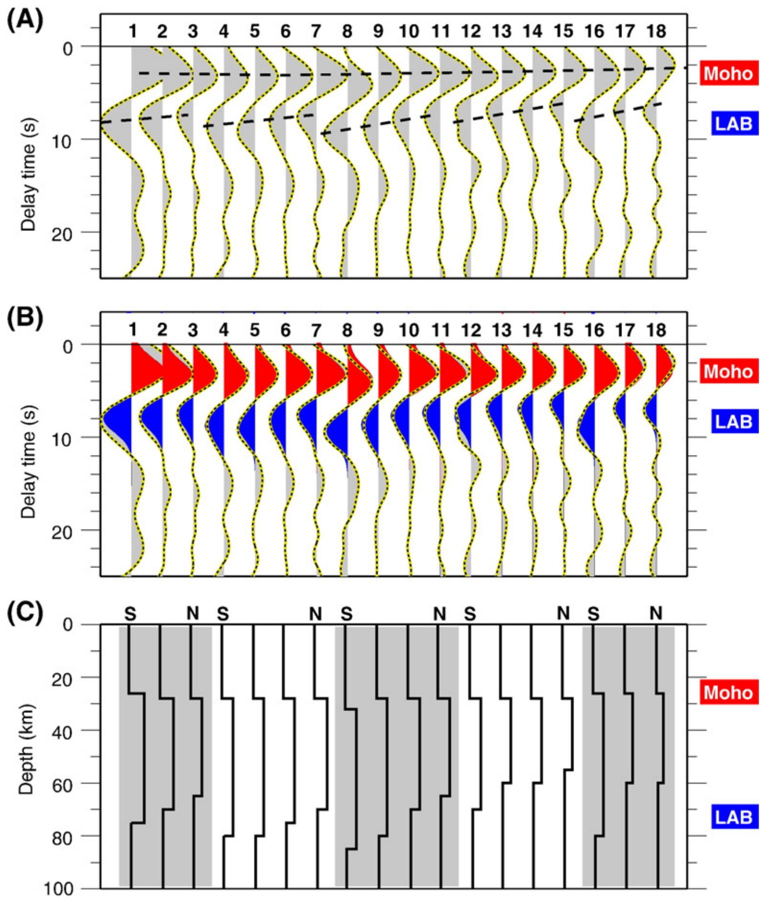

Fig. 3. (A) Observed (stacked) S and SKS receiver function time series for the 18 sectors in Fig. 2 with sector numbers atop. Zero is the arrival time of the S phase. Marked phases (dashed lines) are Sp conversions from the Moho (positive to the right) and the lithosphere–asthenosphere boundary (LAB; negative to the left). The signs of the Moho and the LAB phases are opposite, indicating downward increasing and decreasing velocities, respectively. (B) Synthetic S RFs based on the S velocity models in (C). For comparison, observed receiver functions (yellow line with black dots) are superimposed. (C) Shear-wave velocity models, consisting of Moho and LAB, are grouped from west to east (different shading for sectors 1–3, 4–7, 8–11, 12–15 and 16–18) to indicate the lithospheric thinning from south to north.

much further away from the actual receiver site. In order to enhance the spatial coherence of the S Rfs, their individual (moveout-corrected) traces were stacked within 18 overlapping sectors of size 1.7° (E–W) by 1° (N–S) each, evenly distributed across Ireland (Fig. 2). This stacked trace represents the mean S RF for conversion points at a reference depth of 80 km (crosses in Fig. 2) within a specific geographical sector. The reference depth was estimated from the average Sp delay time of all LAB phases in S and SKS Rfs (in seconds), which can roughly be converted into depth (in kilometres) by multiplication with a factor of 8–10 (based on the iasp91 model [24]).

The spatially stacked receiver functions in Fig. 3A show a prominent (positive) Sp conversion from the crust–mantle boundary (Moho), and a clear negative Sp phase (at 7–9 s) which is due to a velocity reduction below the related seismic discontinuity. This second phase is commonly associated with the conversion from the LAB [1,19]. As a result of the deconvolution, the arrival times of the Moho and the LAB conversions must be measured at the peak of the respective signal.

The delay times of the Moho phase in the 18 sectors are relatively constant at approximately 2.5–3 s, which relates to a 28.5–32 km thick crust found along seismic refraction profiles onshore Ireland [25]. The delay times of the LAB phase clearly decrease from the southern to the northern sectors, implying that the lithospheric thickness also decreases from south to north (e.g. compare sectors 8–11).

3.3. Synthetic S receiver functions

To determine the overall seismic S-velocity structure below Ireland we model synthetic S Rfs for the different sectors. These synthetics RF responses are determined by plane-wave propagation in a stack of homogeneous, isotropic layers above a half-space, based on the reflectivity algorithm [26]. The resulting coloured waveforms in Fig. 3B show an excellent agreement in delay time and amplitude with the measured S Rfs (related velocity models are displayed in Fig. 3C). The strong negative phase at 7–9 s is well matched by Sp conversions from the top of a deep-seated low-velocity zone representing

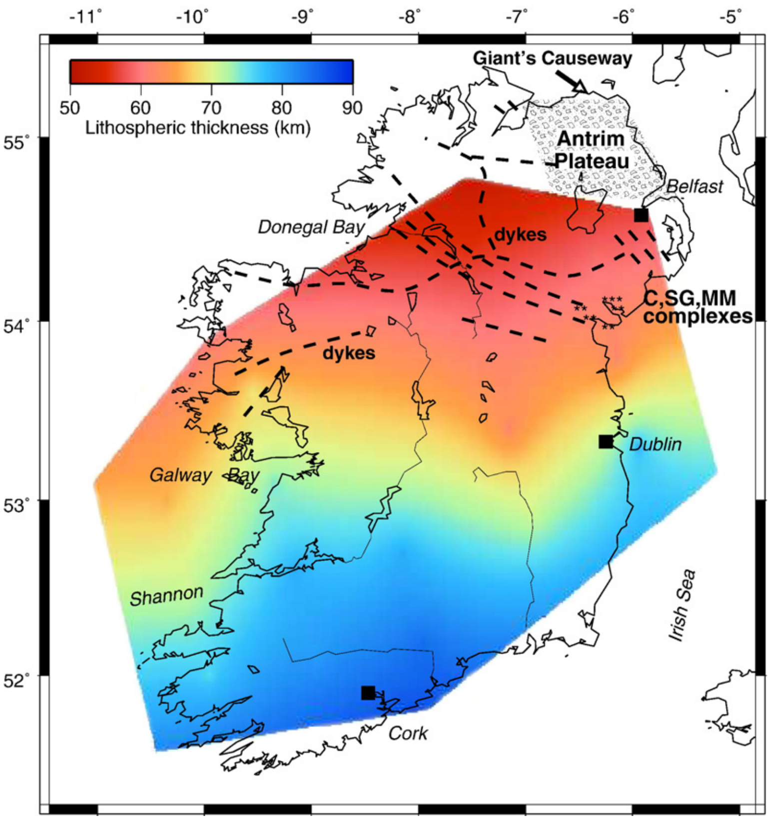

Fig. 4. Map of Ireland with smoothed LAB topography. An overall uncertainty in depth of ±5 km is estimated. Superimposed are locations of known BTVP intrusions found in northern Ireland [14,30]. Dashed lines reflect distribution of Tertiary dykes. Tertiary central complexes (∗) C: Carlingford, SG: Slieve Gullion, MM: Mourne Mountains complexes.

the asthenosphere. An origin from intra-crustal multiples can be excluded, because in S Rfs multiples do not interfere with the direct Sp phase from deeper discontinuities. In the overlying lithosphere, average crustal S-velocities range from 3.70 to 3.85 km/s and increase to 4.30–4.65 km/s in the uppermost mantle. The Moho is relatively flat between 26 and 32 km depth over the entire region. In the underlying asthenosphere velocities decrease to 3.80–3.95 km/s, which supports the concept that the mantle lithosphere is a high-velocity lid on top of the asthenosphere.

In a map view of the best-fit LAB models in Fig. 3C, the LAB clearly rises towards the north from approximately 85±5 km to 55±5 km depth (Fig. 4). The shallow

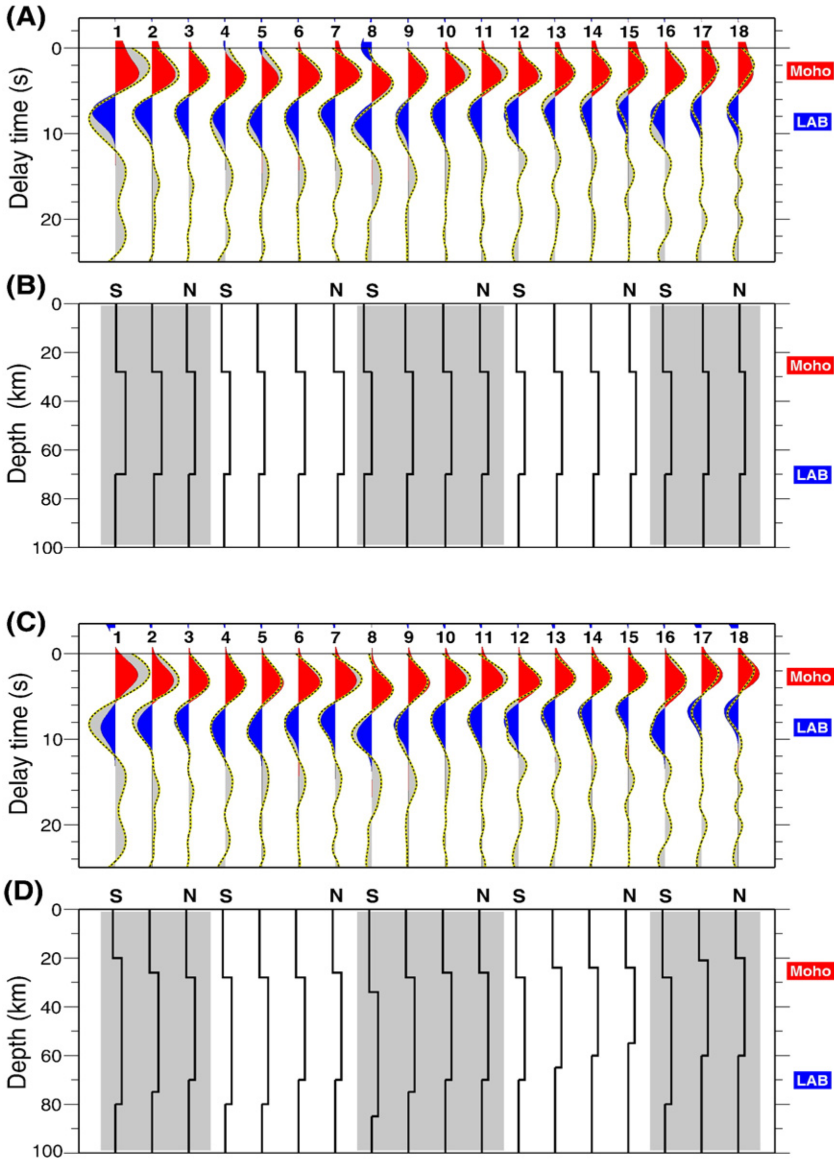

Fig. 5. (A,C) Synthetic S receiver functions (RFs) based on S velocity models in (B,D). For comparison, observed S and SKS RFs (yellow line with black dots; see Fig. 3A) are superimposed. Zero is the arrival time of the S phase. Marked phases are Sp conversions from the Moho (positive to the right) and the lithosphere–asthenosphere boundary (LAB; negative to the left). Numbers atop the figure indicate sectors (for locations see Fig. 2) used for the summation of the receiver functions. (B,D) Shear-wave velocity models, consisting of Moho and LAB, are grouped from west to east (different shading for sectors 1–3, 4–7, 8–11, 12–15 and 16–18) to indicate lateral velocity (B) or lithospheric thickness (D) variations from south to north.

LAB beneath northern Ireland lies well above the regional average of 80–100 km depth for the North Atlantic [1].

3.4. Determination of model stability

In order to estimate the range of possible velocity models that allow us to explain the observed S RF waveforms (e.g. Figs. 3A and 5A), a number of tests were conducted. First, Moho and LAB depths were fixed at 28 and 70 km depth respectively, and only S velocities (vs) within the lithosphere and asthenosphere were varied (Fig. 5A,B). With these constraints, it was very difficult to get a satisfactory fit between synthetic and observed Rfs. The best-fit average crustal velocities (3.2–3.9 km/s) deviate up to 17% from the 3-D crustal velocity model of Landes et al. [25], and reach values that are very similar to (lithospheric) upper mantle velocities (3.85–4.65 km/s) if compared with the iasp91 model. Also, modelled variations of asthenospheric velocities vs in the range of 3.25–4.05 km/s seem unrealistically large. Nevertheless, when comparing sectors to the north and south of the ISZ in Fig. 2, the magnitude of the resulting (average) lateral velocity variations is compatible to those proposed by Masson et al. [17]. In order to explain large teleseismic P-wave traveltime residuals recorded during the VARNET-96 experiment in southern Ireland, they modelled a steeply south-dipping (compositional) boundary within the lithosphere (30–106 km depth) separating two continental blocks juxtaposed during the Caledonian orogeny. In their model, Masson et al. [17] assumed that the LAB has no topography and variations in traveltime residuals across the ISZ are generated by a velocity contrast of 3%.

According to Kumar et al. [1], large changes in lithospheric thickness prevail for the North Atlantic region. In a second test, a better S RF waveform fit was achieved by varying the depths of the Moho and LAB only, while keeping average velocities fixed at 3.65 km/s (crust), 4.35 km/s (lithospheric mantle) and 3.85 km/s (asthenosphere) (Fig. 5C,D). However, this resulted in a Moho topography with unrealistic large undulations (20–34 km depth) in comparison to the 3-D model of Landes et al. [25], which indicates a rather flat Moho topography in the depth range of 28–32 km depth.

Therefore a combination of lateral shear-wave velocity and Moho depth variations, within the limits of the 3-D velocity model of Landes et al. [25], and lateral thickness variations of the lithosphere was deemed necessary. The different combinations for the depths of the Moho (26–32 km) and the LAB (55–85 km), as well as vs-velocities in the crust (3.7–3.85 km/s; compatible with Landes et al. [15], if converted to vp-velocities assuming a Poisson ratio of 0.25) and the mantle lid (4.3–4.65 km/s), resulted in an excellent agreement between the observed and synthetic S RF waveforms (Fig. 3B). Variations of the thickness of the Moho and LAB showed that these boundaries may be sharp first order discontinuities, but could also be of transitional nature (maximal ± 10 km) given our period range of 4–20 s. However, within the limits of resolution and reasonable vs values, the absolute depth of these discontinuities, as determined by the time delays of the maxima in the Rfs, is accurate to within ± 5 km. Results from P Rfs (period range 0.5–15 s; [18]) put a more solid constraint on the Moho depth with an uncertainty in the range of about ± 1 km.

In conclusion, we find it necessary to infer a variation in LAB depth to explain the observed S Rfs (Figs. 3 and 4). An explanation of the observed variations in Sp delay times for the LAB by lateral velocity changes only is not supported.

4. Interpretation and discussion

Lithospheric thickness variations underneath Ireland, as summarised in Fig. 4, may occur for several reasons. One option would be that the LAB variation resembles the transition from continental to oceanic lithosphere. In the case of Ireland, however, this transition is located offshore, much further to the northwest than our anomaly [3,27]. Second, there is also no indication for lithospheric thinning due to continental rifting onshore Ireland [28]. Whether the observed topographic variations of the LAB are related to the proposed location of the ISZ (Fig. 2) remains speculative, as there is no obvious relation between the Caledonian continent–continent collision and lithospheric thinning. Teleseismic Rfs from the western part of the ISLE network showed that delay times for Ps conversions from the mantle transition zone (P410s, P660s) are in agreement with the standard iasp91 model. No distinct variations in depth of these discontinuities were observed, thus ruling out the present-day existence of a thermal anomaly at the mantle transition zone beneath Ireland [18].

Alternatively, thermal erosion by upwelling of hot asthenospheric material is suggested to enable partial alteration and removal of the subcrustal lithosphere [13]. We prefer the latter option to explain the LAB variations in Fig. 4, because northern Ireland geographically belongs to the British Tertiary volcanic province (BTVP), which is supposed to be directly related to the proto-Iceland plume of early Tertiary age [11–14].

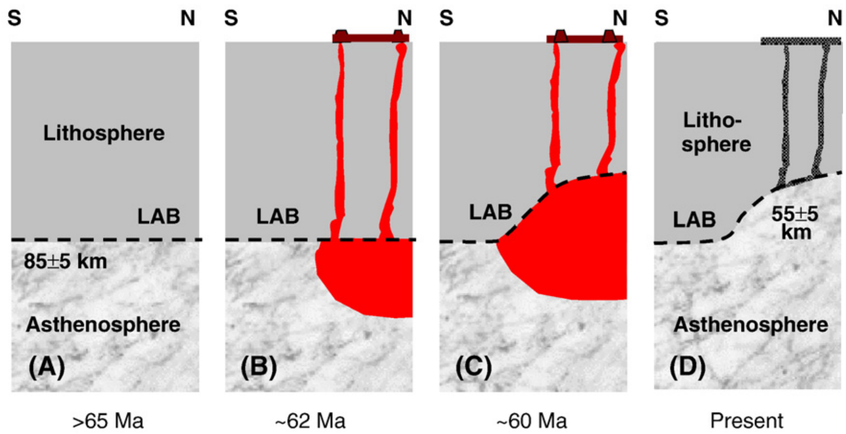

Fig. 6. Cenozoic evolution of the deep lithosphere (grey) in Ireland due to interaction with the head of the proto-Iceland plume (dark grey). (A) About 85 km thick lithosphere before the arrival of the plume head. (B) Start of plume–lithosphere interaction at approximately 62 Ma. (C) Rapid thinning of the lithosphere in the northern part of Ireland due to thermal erosion (62–60 Ma). (D) Present state with still elevated asthenosphere (light grey pattern) in this region.

The wide-spread activity in the BTVP started about 62 Ma ago [29,14]. The main eruption period, which occurred during the early stages and lasted for a period of 3–4 Ma only [30], involved the western coast of central England and Scotland [4], and most of northern Ireland generating numerous dykes, intrusions as well as the basaltic Antrim Plateau with the Giant's Causeway (Fig. 4 [14,30]). With the contemporaneous flood–basalt volcanism in Greenland [3], geochemical evidence relates the source of the BTVP to the spreading head of the proto-Iceland plume [11,14]. Isotope data support the hypothesis of a deep origin of the melts, e.g. An ancient plume head under northern Ireland and the Irish Sea [14,28]. Geochemical modelling of lavas from Mull (West Scotland) revealed that a rapid thinning of the lithosphere from at least 60–70 km to 55 km thickness occurred at 60 Ma which is best explained by plume-related thermal erosion [13]. A similar mechanism was proposed for the Irish Antrim Plateau [28].

Even today, teleseismic tomography [31] finds anomalous low-velocity material in the upper mantle below the Irish Sea and western Britain. Al-Kindi et al. [9] concluded from the interpretation of both seismic refraction and gravity data that a hot convective sheet of Icelandic plume material underplates the same region. Our study presents detailed seismic evidence for asthenospheric upwelling beneath the northern part of Ireland. Although the main magmatic activity occurred between 62 and 43 Ma [14], this event is still preserved by the elevated LAB (Fig. 4). In addition, mantle S-velocities of less than 4 km/s cannot be explained without the presence of an elevated temperature and considerable partial melt, even if the effect of anisotropy on seismic velocities is considered [32], and a high water content is assumed [33].

From our observations, the Cenozoic plume–lithosphere interaction below Ireland can be described as follows (Fig. 6). The original base of the lithosphere at about 80–85 km depth [28] was heated by the head of the proto-Iceland plume at or slightly before 62 Ma. The short period of excessive volcanism [30] and the remarkable change in the geochemical composition [13,28] suggest rapid lithospheric thinning due to thermal erosion from 62–60 Ma [13]. As a result, the thinnest lithosphere of 55–60 km was achieved beneath northern Ireland (Figs.4 and 6). This is also documented by low-pressure melting of mantle peridotites which formed the Causeway tholeiites [28]. While plume-related magmatism lasted at least until 43 Ma [14], the unusually low S-velocities of 3.8–3.95 km/s in the asthenosphere (Fig. 3C) indicate that the ambient temperature is still elevated, as was also found by teleseismic P-wave tomography to the east of Ireland [31]. The observed low velocities are unlikely to extend as deep as the mantle transition zone, as no structural changes were observed at these discontinuities [18]. To the south, further away from the former plume head, the lithosphere preserved its original thickness of about 80–85 km which compares well with other S RF studies in the North Atlantic region [1]. In conclusion, we suggest that the upwelling of the asthenosphere beneath northern Ireland is caused by the dynamic support of the proto-Iceland plume in conjunction with the thermal erosion of the lower lithosphere.

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