Lithos 189 (2014) 49–64

Contents lists available at ScienceDirect

Lithos

journal homepage: www.elsevier.com/locate/lithos

The lithosphere–asthenosphere system beneath Ireland from integrated geophysical–petrological modeling II: 3D thermal and compositional structure

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ARTICLE INFO

Article history:
Received 23 March 2013
Accepted 27 September 2013
Available online 9 October 2013

Keywords:
Lithosphere–asthenosphere boundary
Geophysical–petrological modeling
Mantle composition
Xenoliths
Ireland
Seismic tomography

ABSTRACT

The lithosphere–asthenosphere boundary (LAB) depth represents a fundamental parameter in any quantitative lithospheric model, controlling to a large extent the temperature distribution within the crust and the uppermost mantle. The tectonic history of Ireland includes early Paleozoic closure of the Iapetus Ocean across the Iapetus Suture Zone (ISZ), and in northeastern Ireland late Paleozoic to early Mesozoic crustal extension, during which thick Permo-Triassic sedimentary successions were deposited, followed by early Cenozoic extrusion of large scale flood basalts. Although the crustal structure in Ireland and neighboring offshore areas is fairly well constrained, with the notable exception of the crust beneath Northern Ireland, the Irish uppermost mantle remains to date relatively unknown. In particular, the nature and extent of a hypothetical interaction between a putative proto Icelandic mantle plume and the Irish and Scottish lithosphere during the Tertiary opening of the North Atlantic has long been discussed in the literature with diverging conclusions. In this work, the present-day thermal and compositional structure of the lithosphere in Ireland is modeled based on a geophysical–petrological approach (LitMod3D) that combines comprehensively a large variety of data (namely elevation, surface heat flow, potential fields, xenoliths and seismic tomography models), reducing the inherent uncertainties and trade-offs associated with classical modeling of those individual data sets. The preferred 3D lithospheric models show moderate lateral density variations in Ireland characterized by a slightly thickened lithosphere along the SW-NE trending ISZ, and a progressive lithospheric thinning from southern Ireland towards the north. The mantle composition in the southern half of Ireland (East Avalonia) is relatively and uniformly fertile (i.e., typical Phanerozoic mantle), whereas the lithospheric composition in the northern half of Ireland (Laurentia) seems to vary from moderately depleted to fertile, in agreement with mantle xenoliths erupted in northwestern Ireland.

1. Introduction

Ireland has remained relatively stable since the mid-Eocene times. However, the geology of the island reveals an interesting tectonic history since, at least, Proterozoic times to late-Paleozoic times then again more recently (Eocene times). That includes closure of the Iapetus Ocean and hence welding of the two terranes, Laurentia and Easter Avalonia, of which present-day Ireland is composed, as well as later extrusion of large scale flood basalts in northeastern Ireland at around 42 Ma. An essential key to understanding the geodynamic evolution of a region is to define, with reasonable accuracy, the present-day thermal and compositional structure of the lithosphere as an end-point for geodynamic scenario modeling. The lithosphere–asthenosphere boundary (LAB) divides the outermost, cold, relatively rigid layer of the Earth (the lithosphere) from the warmer and rheologically weaker sub-lithospheric or asthenospheric mantle. The LAB's nature, as based on a number of different geophysical and geochemical parameters (e.g., seismic velocities, seismic anisotropy, temperature, composition, electrical resistivity), is still a matter of intense debate in the solid Earth academic community (e.g., Plomerova et al., 2002; Eaton et al., 2009; Romanowicz, 2009; Rychert and Shearer, 2009; Griffin et al., 2009; O'Reilly and Griffin, 2010; Jones et al., 2010; Schmerr, 2012). The topology of the LAB controls to a large extent the temperature distribution within the upper mantle and is, therefore, a crucial parameter in any quantitative lithospheric model.

The crustal structure in Ireland and neighboring offshore areas is reasonably well known after three decades of seismic experiments (e.g., Chadwick and Pharaoh, 1998; Hauser et al., 2008; Hodgson, 2001; Jacob et al., 1985; Landes et al., 2000; Lowe and Jacob, 1989; Masson

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http://dx.doi.org/10.1016/j.lithos.2013.09.014

et al., 1998), with the exception of Northern Ireland, where no constraints on large-scale crustal structure are currently available. However, a number of issues regarding the Irish mantle still remain relatively unclear. In particular, the nature and extent of a hypothetical interaction between a putative proto Icelandic mantle plume and the Irish (and, in a broader context, the British Isles) lithosphere during the Tertiary opening of the North Atlantic has long been discussed in the literature with diverging conclusions (e.g., Al-Kindi et al., 2003; Davis et al., 2012; Landes et al., 2007; O'Donnell et al., 2011). Furthermore, the European-scale seismic tomography images upon which much of the discussion is based do not agree as to whether a clearly fast or slow sub-lithospheric mantle (i.e., depths >100 km) exists beneath Ireland (e.g., Piromallo and Morelli, 2003; Schivardi and Morelli, 2011; Legendre et al., 2012).

In this work we use geophysical and petrological data to constrain a 3D model of the thermal and compositional structure of the Irish lithosphere based on both geophysical (elevation, potential fields, surface heat flow, seismic tomography) and petrological (mantle xenoliths) constraints. We use the geophysical–petrological modeling tool LitMod3D, that solves simultaneously the heat transfer, thermodynamic, rheological, geopotential and isostasy equations in the lithosphere and sub-lithospheric upper mantle (Afonso et al., 2008; Fullea et al., 2009), in a trial-and-error forward modeling. Key mantle properties (e.g., seismic velocities and density) are determined as a function of the pressure, temperature and compositional conditions that vary with depth. Such a thermodynamically self-consistent approach allows us to combine comprehensively a large number of geophysical and petrological observables, reducing the inherent uncertainties and trade-offs associated with classical modeling of those data sets individually. Allowable models from individual data are eliminated when other data are included. In this way, we can explore acceptable model space and define models that satisfy all data.

2. Geological setting: Ireland

The present day Irish lithospheric structure is largely the result of the combined action of tectonic and magmatic processes acting since Proterozoic times. The Iapetus Suture Zone (ISZ) is a major SW-NE trending tectonic boundary that divides Ireland into two main tectonic terranes: Eastern Avalonia in the southeast and Laurentia in the northwest (Fig. 1). Both terranes were accreted during the oblique closure of the Iapetus Ocean in early Paleozoic times (e.g., Chew and Stillman, 2009). A phase of granitic magmatism associated with the end of oceanic subduction and terrane accretion is represented by the widespread Newer Granite series and associated volcanic rocks (late Silurian–early Devonian). Elsewhere along the ISZ, granites north of it are less radiogenic and older than those in southern Ireland (Brown et al., 2008; Conliffe et al., 2010; Stone et al., 1997). The Caledonian closure of the Iapetus Ocean involved large-scale sinistral transpression, i.e. accretion of predominantly Ordovician and Silurian turbidite and volcaniclastic sediments, and led to a predominant SW-NE structural pattern in Caledonian rocks (e.g., Morris, 1987; Ryan and Dewey, 1991). This SW-NE trend defined by the surface geology of the upper crust (Woodcock and Strachan, 2000) and the gravity field fabrics (Readman et al., 1997) is also observed in the lower crust today according to seismic anisotropy studies (Polat et al., 2012). The latter means that the accretion history of the Irish crust was essentially complete by the early Carboniferous, with little modification thereafter (O'Reilly et al., 2012).

During the subsequent Variscan orogeny a number of sedimentary basins were produced (mid-late Devonian–early Carboniferous) and later inverted due to the reactivation of old faults in the Caledonian basement during the later-Carboniferous–early Permian NNW-SSE compression (Readman et al., 1997; Williams et al., 1989). The deformation in the basins is important in the southern part of Ireland, leading to an E-W trend in the Variscan cover sequence (Williams et al., 1989).

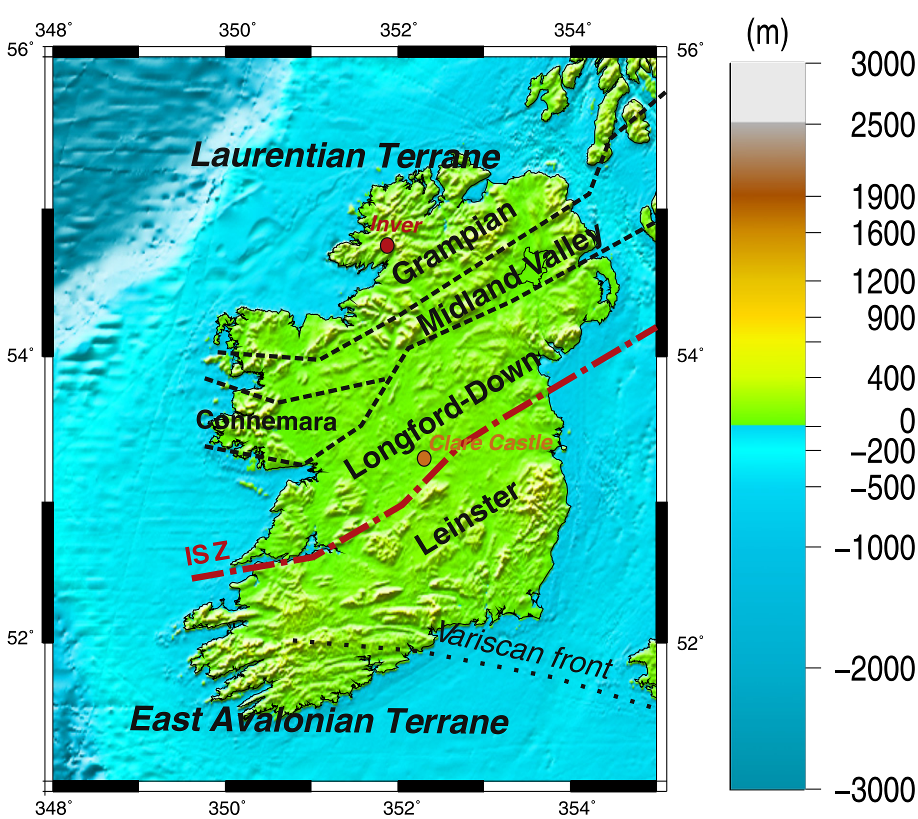

Fig. 1. Elevation map of Ireland from the DEM ETOPO2 (Smith and Sandwell, 1994, 1997). The divisions for the main tectonic terranes are shown as black dashed lines. The thick red dash-dotted line is the Iapetus Suture Zone (ISZ), a major tectonic boundary between the Laurentian (north) and the Eastern Avalonian (south) plates. The thin dotted line is the Variscan deformation front. Orange and red points are lower crustal and mantle xenolith localities in Clare Castle and Inver respectively (see text for further details).

Next, associated with the subsequent Phanerozoic tectonic and magmatic phase, the lithosphere is stretched and several Mesozoic basins are developed in the Irish offshore region (O'Reilly et al., 2010) as well as beneath onshore Northern Ireland. Subsequently, Cretaceous sea-floor spreading in the North Atlantic, development of Cenozoic volcanic continental margins in the North Atlantic and, most recently, the extrusion of large scale flood basalts (Northern Ireland and Scotland mainly) followed (Preston, 2009; Welford et al., 2012).

Much of present-day Ireland is covered by Pleistocene fluvioglacial deposits that obscure the details of the underlying geological structure (Mitchell, 1981). Rocks related to the Proterozoic to Caledonian orogenic cycles are present in the west and NW margins (i.e., Dalradian) and eastern Ireland (Paleozoic).

3. Geophysical and petrological setting

3.1. The Irish crust

A number of crustal and upper-mantle seismic studies have been carried out since the mid-1960's aimed at imaging the lithospheric structure of Ireland and surrounding offshore areas. Landes et al. (2005) present a review of the seismic reflection/refraction profiles recorded in the period 1969–1999 (e.g., the ICSSP 1982, COOLE 1985, VARNET 1996 and LEGS 1999 surveys, see Fig. 2A for their locations). Landes et al. (2005) present a Moho depth map of central and southern Ireland based on their compilation of the 2D seismic studies. The topography of the Moho shows smooth variations in central and southern Ireland with the thickest crust present along the eastern and southwestern margins of Ireland (32 km), and a gentle thinning towards the NW (28.5 km) (see Landes et al., 2005, and references therein).

Consistent with this, a compilation of offshore deep seismic reflection data in the UK and surrounding areas (BIRPS data set, Chadwick and Pharaoh, 1998, see Fig. 2A for locations) shows a reflection Moho at TWTT (two-way travel time) of 10.5 s in the southern part of the Irish Sea between Ireland and the UK, 10 s in the center, and 9.5 s in the northern part (North Channel). For a standard 6 km s^{-1} migration velocity, these times convert to depths of 31.5 km, 30 km, and 28.5 km respectively. According to Chadwick and Pharaoh's (1998) seismic reflection compilation there is also a relatively thinned crust along the NW Atlantic margin of Ireland (23–29 km). A similar crustal pattern in the north Irish margin is observed in the seismic P-wave velocity model compiled by Kelly et al. (2007) from wide-angle reflection and refraction profiles. A recent P-to-S (Ps) receiver function study of the British Isles reports a fairly constant Moho depth of 30–31.6 km in central and NE Ireland with a moderate to low average crustal Vp/Vs ratio (1.64–1.73) (Davis et al., 2012). However, one station located in NW Ireland in Davis et al.'s (2012) study (station GNP) yields a crustal thickness of around 27 km and a relatively high average crustal Vp/Vs ratio (1.86). An earlier Ps receiver function study by Shaw Champion et al. (2006) yielded a crustal thickness of 29.6 km and a relatively high average crustal Vp/Vs ratio (1.86) for that station. However, as Shaw Champion et al. (2006) acknowledge, the form of the GNP receiver functions is unusual, dominated by reverberations from a strong shallow converter, likely due to GNP sitting on top of the Donegal granite. The presence of the strong reverberations prevents any reliable Ps crustal identification in GNP

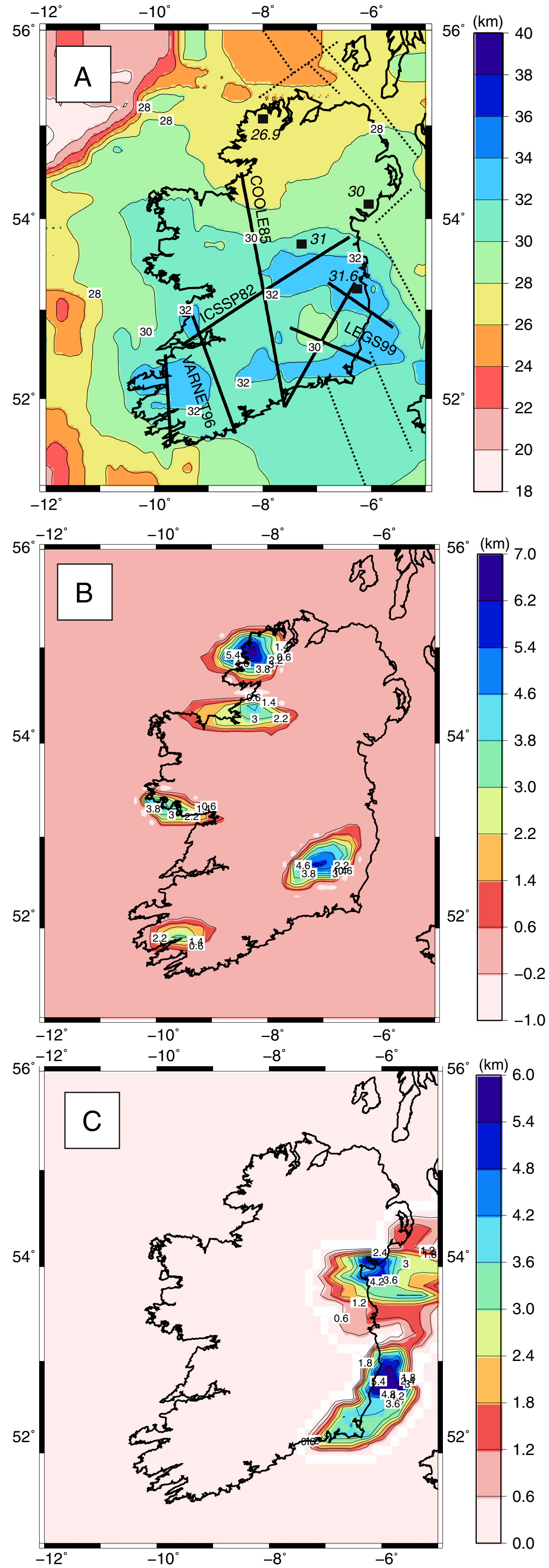

Fig. 2. Crustal model. A) Moho depth map. Black lines are seismic refraction lines used as constraints: VARNET96 (Hauser et al., 2008; Landes et al., 2000; Masson et al., 1998), LEGS99 (Hodgson, 2001), COOLE85 (Lowe and Jacob, 1989) and ICSSP82 (Jacob et al., 1985). Dotted lines are reflection profiles collected by the BIRPS group (Chadwick and Pharaoh, 1998 and references therein). Black squares are seismic stations with Moho depth estimates (in km) based on receiver functions (Davis et al., 2012). B) and C) thickness maps of the New Series granite and mafic volcanic crustal bodies respectively (see Table 2 for their geophysical parameters).

(Shaw Champion et al., 2006). The stations along the west Scottish margin show a Moho depth ranging from 25 km to 30 km based on receiver functions (Davis et al., 2012). In particular, stations located on tectonic terranes in Scotland (e.g., the Northern Highlands and Central Lowlands terranes) that have their natural continuation in Northern Ireland are all characterized by Moho depths <27 km (Davis et al., 2012).

Seismically, the crust in Ireland can be divided into three layers according to seismic refraction studies: upper, mid and lower-crust, characterized by Vp velocities of 5.7–6.2 km/s, 6.3–6.7 km/s and 6.8–7.2 km/s respectively (e.g., Landes et al., 2005; Hauser et al., 2008; O'Reilly et al., 2010, 2012 and references therein). A reinterpretation of the unusually high quality S-wave data set from the VARNET seismic experiment (Landes et al., 2000; Masson et al., 1998), indicates an anomalously felsic nature of the crust in the southern part of Ireland (i.e., the E. Avalonian terrane), particularly in the lower crustal layer: Vp/Vs = 1.76, SiO2 ≈67% (compared to the global continental average of 62%, e.g., Christensen and Mooney, 1995; Christensen, 1996) (Hauser et al., 2008).

These seismic observations are in agreement with the petrophysical and geochemical signature of lower crustal metapelitic xenoliths from central Ireland, close to the ISZ (van den Berg et al., 2005). The felsic, silica-rich nature of the crust seems to continue north of the ISZ into the Laurentian terrane, where all the active source seismic and receiver function studies suggest Vp/Vs values < 1.75, with the notable exception of the Irish continuation of the Northern Highland Scottish terrane (Assumpçao and Bamford, 1978; Davis et al., 2012). In spite of this observed compositional homogeneity in both the E. Avalonian and Laurentian terranes, the fine-scale structure of the crust presents some differences across the ISZ. Based on a full waveform analysis of both P- and S-wave coda from VARNET, O'Reilly et al. (2010) found that the reflectivity of the lower crust is significantly higher in E. Avalonia than that in the north of the ISZ, in the southern margin of the Laurentian continent (e.g., the late Paleozoic Clare Basin). O'Reilly et al. (2010) explain the reflectivity of the E. Avalonian lower crust in terms of a laminated velocity structure related to the partial melting of a metapelitic accretionary wedge ('incipient delamination model').

3.2. The Irish lithosphere

3.2.1. Seismic studies

At a regional, European scale, the lithospheric P- and S-wave velocities in Ireland are in general faster than global velocity models according to body-wave seismic tomography (e.g., Marquering and Snieder, 2007; Bijwaard et al., 1998; Piromallo and Morelli, 2003; Amaru et al., 2008; Schivardi and Morelli, 2011). The fast P-wave velocities seem to continue deeper into the sub-lithospheric mantle beneath Ireland (e.g., Piromallo and Morelli, 2003). However, a recent multimode inversion of surface- and S-wave forms in Europe points to a distinct low shear velocity anomaly in the mantle (with respect to global model AK135) at sub-lithospheric depths of 80–200 km below the northern portion of Ireland (Legendre et al., 2012, as illustrated in Fig. 3A). In most of these studies however, Ireland is at the very edge of the modeling domain, so care must be taken when their results are interpreted from an Irish perspective. Local scale (i.e., at the scale of Ireland or the British Isles) P-wave tomography studies show relatively moderate lateral variations in the seismic velocities across the island (e.g., Arrowsmith et al., 2005; O'Donnell et al., 2011; Wawerzinek et al., 2008). In spite of the discrepancies between the local seismic studies, some features are consistently imaged by all the models in the lithospheric depth range 80–100 km (as exemplified in Fig. 3B). Namely, positive mantle anomalies are visible i) in the E. Avalonian plate east of Dublin (Irish Sea), and in the SE Irish margin, and ii) in a NW-SE zone within the Laurentian terrane from the Galway area to central Ireland. On the contrary, the mantle seems to be slow in central Ireland (north of the ISZ), and in SW Ireland, in the E. Avalonian terrane (Arrowsmith et al., 2005; O'Donnell et al., 2011; Wawerzinek et al., 2008). The surface- and S-wave model of Legendre et al. (2012) shows a similar pattern of relative velocity anomalies to that depicted by local P-wave tomography except for the negative anomalies in the SW Avalonian plate and the positive anomalies in the Irish Sea, east of Dublin (Fig. 3B). For this model however, station coverage is poor in Ireland as only two stations were used, one in the SW coast (VAL, Valentia), and one in the Dublin mountains (DSB, Dublin).

Based on S receiver functions computed from the ISLE teleseismic network data, Landes et al. (2007) inferred a lithospheric thickness of around 85 km in the E. Avalonian terrane and a conspicuous thinning towards the north (LAB depth as shallow as 55 km). Landes et al. (2007) relate the lithospheric thinning in the Laurentian terrane to the interaction with the proto-Iceland mantle plume and the Cenozoic magmatism.

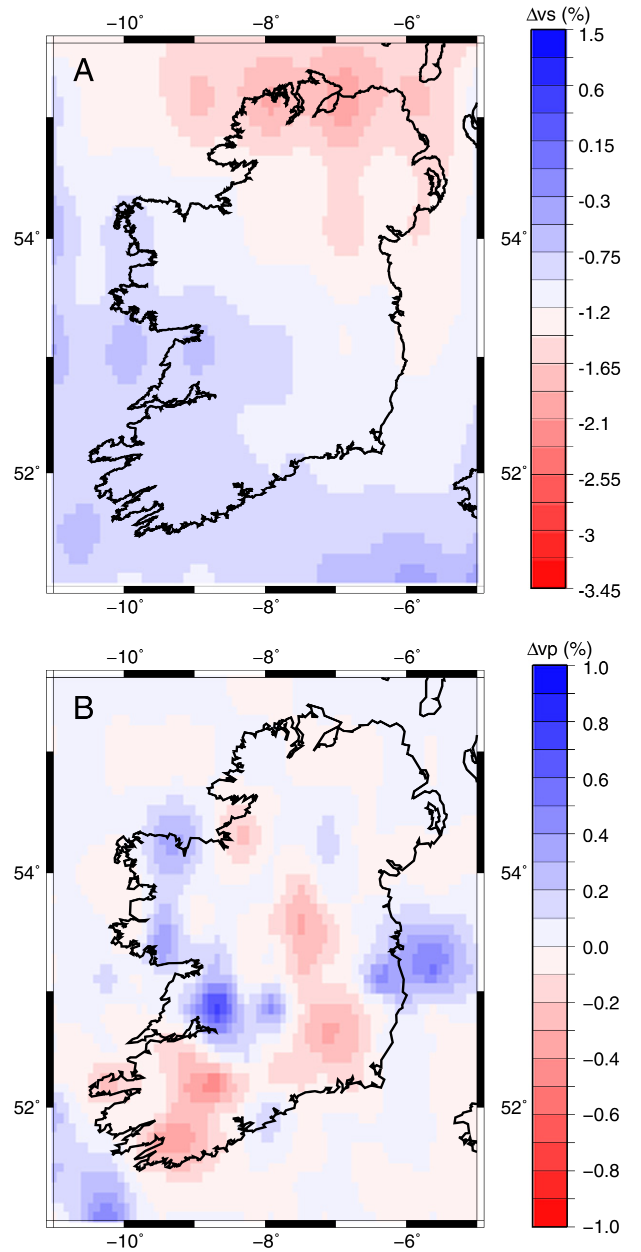

Fig. 3. Seismic tomography models in Ireland. A) Multimode inversion of surface- and S-wave forms in Europe from Legendre et al. (2012). Slice at a depth of 80 km, the reference velocity is 4.38 km/s. B) Local P-wave tomography from O'Donnell et al. (2011). Slice at 85 km depth.

Table 1
Bulk mantle compositions used in this work from xenolith suites and peridotite massifs.

	1) Av. tecton gnt. perid. (wt.%)[a]	2) Harz. inver (wt.%)[b]	3) Lherz. inver (wt.%)[b]	4) Derbyshire av. (wt.%)[c]	5) Fidra av. (wt.%)[d]	6) PUM M&S95 (wt.%)[e]
SiO_2	45	41.7	45.84	43.4	44.31	45
TiO_2	0.16	–	–	0.02	0.15	0.201
Al_2O_3	3.9	2.65	3.92	2	3.5	4.45
Cr_2O_3	0.41	–	–	0.51	–	0.384
FeO	8.1	8.32	7.19	7.4	8.6	8.05
MnO	0.07	–	–	0.22	0.14	0.135
MgO	38.7	44.86	38.05	44.5	38.5	37.8
CaO	3.2	0.77	2.72	1.5	3.3	3.55
Na_2O	0.28	0.05	0.21	0.08	0.3	0.36
NiO	0.24	–	–	0.32	–	–
Total	100.06			99.95	99.8	99.93
Mg#	89.5	90.58	90.42	91.47	87.73	89.3

[a] Global average taken from Griffin et al.(2009).
[b] Inver xenoliths suite (NW Ireland) 2) and 3) are sample 6 and 3 from Shaw and Edgar (1997) respectively.
[c] Xenolith data averaged from Calton Hill, Derbyshire (England) in Donaldson (1978).
[d] Fidra xenolith suite average (Midland Valley of Scotland) from Downes et al. (2001).
[e] PUM stands for Primitive Upper Mantle, M&S95 refers to McDonough and Sun (1995).

3.2.2. Mantle xenoliths

The only known mantle xenolith locality in Ireland is at Inver, Donegal, on the NW margin of the island (see location in Fig. 1) (Shaw and Edgar, 1997). The peridotite xenoliths from Inver are spinel lherzolites and harzburgites hosted in alkali basalt, similar to Permo-Carboniferous alkaline igneous xenolithic rocks from the Scottish Midland Valley (Gallagher and Eldson, 1990; Upton et al., 1984). The presence of devitrified glass (i.e., infiltrated melt) and coronas in some minerals (e.g., sieve-textured clinopyroxene) suggests a melt–mineral interaction during the ascent of the xenoliths (Shaw and Edgar, 1997). The composition of the primary minerals in the Inver xenolith suite falls within the range of fertile peridotite, although the low percentage of modal clinopyroxene is a feature typical of depleted mantle (Shaw and Edgar, 1997). From the six xenoliths at Inver analyzed by Shaw and Edgar (1997), one is a spinel harzburgite (sample 6) which shows little melt interaction (i.e., small amounts of glass and no sieve-textured

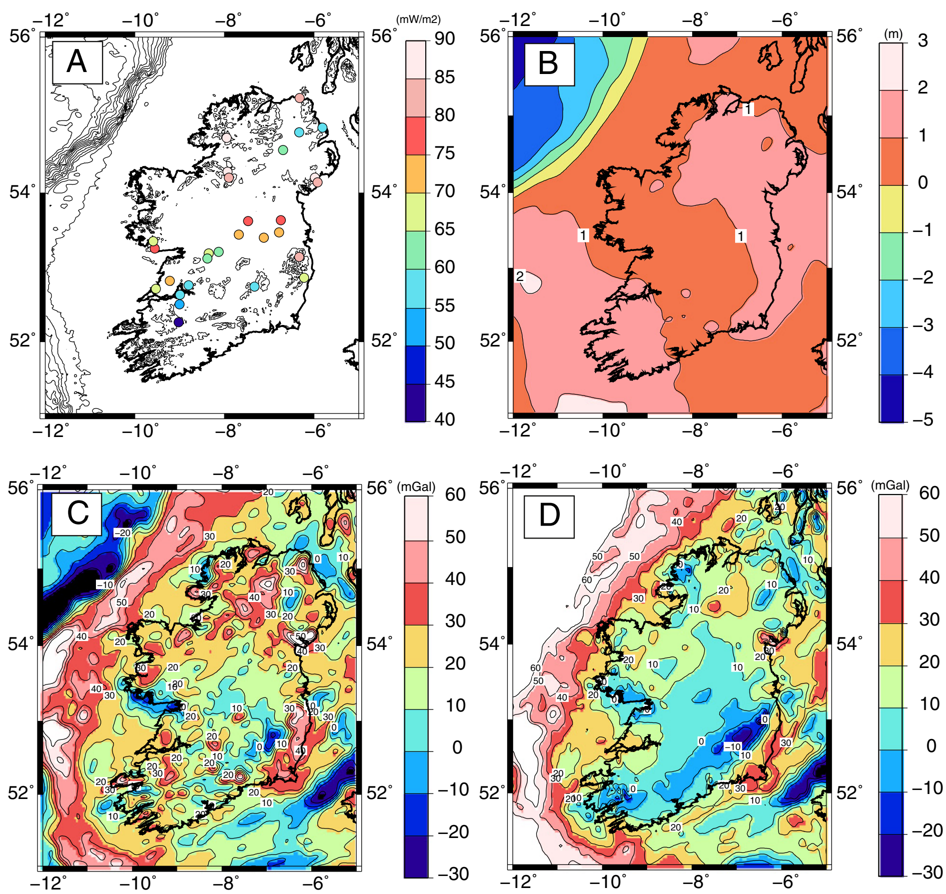

Fig. 4. Geophysical observables in Ireland. A) Measured surface heat flow (SHF) in Ireland from Brock and Barton (1984) and Brock (1989). B) Geoid anomaly from the global Earth model EGM2008. Long wavelengths (n<9) have been removed to retain the lithospheric signal. C) Free air anomaly from satellite data (EGM2008). D) Bouguer anomaly: onshore data measured and corrected by the Dublin Institute for Advanced Studies and the Geological Survey of Northern Ireland, offshore data were computed applying the complete topographic correction to free air satellite data. E1–E6) Gravity gradients from the satellite-only (mission GOCE) global Earth model GOCO03S (http://www.goco.eu/) computed up to degree and order 220 at 255 km above the sea level (GOCE mission perigee height) using a spherical harmonics synthesis code (Prof. Z. Martinec, DIAS, personal communication). The sub-indexes of V denote second derivative of the Earth's potential (i.e., gravity gradient) along the south-north, east-west, and vertical directions (x, y, z respectively, in the LNOF, see Section 4 for more details).

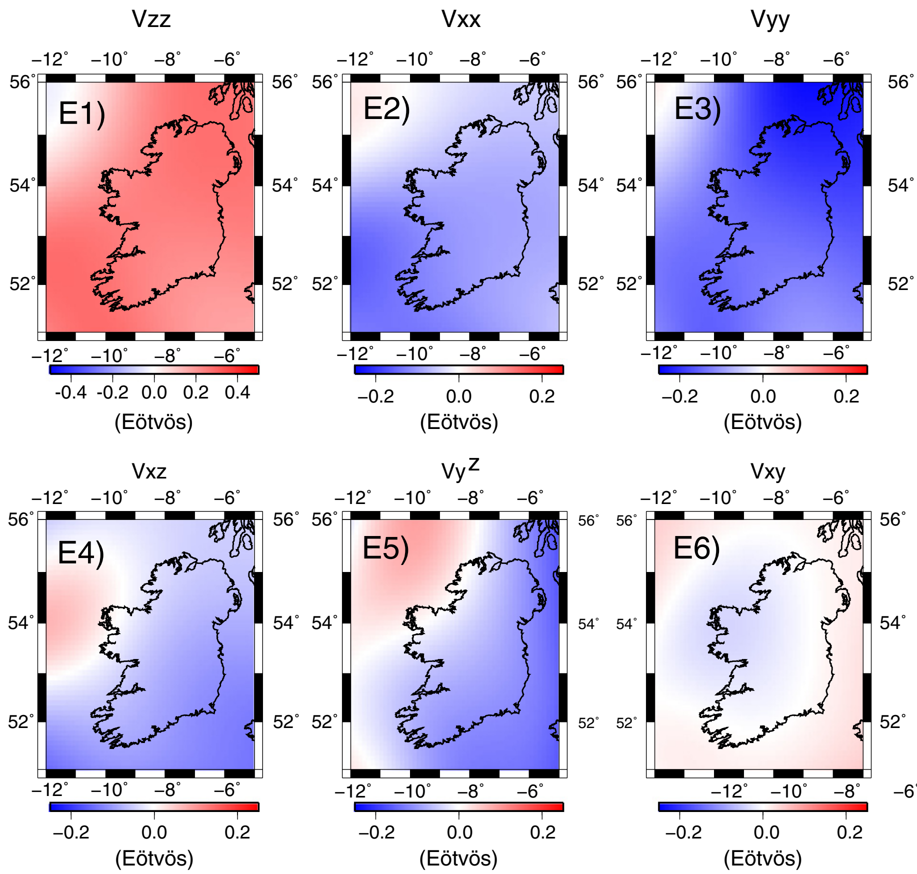

Fig. 4 (*continued*).

cpx) and the others are spinel lherzolites with variable degree of interaction with the melt. Among the spinel lherzolites, sample 3 stands out over the rest in that its modal orthopyroxene is clearly higher (≈38%) than the average of the group (≈15%) (Shaw and Edgar, 1997, Table 1).

Although not directly located in the study area, there are several xenolith localities in Britain lying in tectonic terrains which have their natural continuation in Ireland (e.g., Hunter et al., 1984; Menzies and Halliday, 1988). In the Laurentian terrane, several spinel lherzolite xenoliths from Fidra (Midland Valley of Scotland) entrained in a Permian basanite sill were analyzed by Downes et al. (2001). In terms of the major oxides, the xenoliths are similar to those of sample 3 at Inver (Shaw and Edgar, 1997) except for the significantly higher amount of iron (see Table 1). In the E. Avalonian plate, the xenolith locality closest to Ireland is at Calton Hill, Derbyshire (England). The suite in Derbyshire includes spinel lherzolites and harzburgites hosted in an ankaramite lava derived from about 45 km depth and a temperature of 950 °C (Donaldson, 1978). The average composition of the Derbyshire xenoliths lies within global averages for spinel lherzolites, with the particularity of having CaO and NaO_2 oxide contents in the lower part of the global range, and the MgO oxide content at the maximum of the global range (Donaldson, 1978). The MgO, CaO and Al_2O_3 contents of the Derbyshire xenoliths are similar to that of sample 6 (harzburgite) at Inver (Shaw and Edgar, 1997); in contrast, the Derbyshire xenolith FeO content is closer to that of sample 3 (lherzolite) at Inver (Table 1).

4. Geophysical observables

Regional geophysical datasets (elevation, gravity, gravity gradients, surface heat flow, geoid anomaly, Fig. 4) were taken from different sources. The elevation data, i.e., topography and bathymetry, are taken from the ETOPO2 Global Data Base (Fig. 1) (Smith and Sandwell, 1994, 1997). Geoid and free-air anomalies were obtained from the global Earth model EGM2008, which includes spherical harmonic coefficients up to degree and order 2190 (Fig. 4B and C) (Pavlis et al., 2008). The geoid signal has been filtered to remove long wavelengths of deep origin (i.e., >4000 km, degrees 2–9) and retain the effects of density anomalies shallower than ~400 km depth (Bowin, 2000). Bouguer anomalies onshore were measured and corrected by the Dublin Institute for Advanced Studies and the Geological Survey of Northern Ireland (Readman et al., 1997, and references therein). Offshore data were computed applying the complete topographic correction to free air satellite data (from the global Earth model EGM2008) using the software FA2BOUG and assuming a reduction density 2670 kg/m^3 (Fullea et al., 2008) (Fig. 4D). The compilation of surface heat flow data used in this work comes from different sources (Brock, 1989; Brock and Barton, 1984) (Fig. 4A). For further details on the surface heat flow data compilation see the extensive discussion in a companion paper by Jones et al. (2013).

Gravity gradients were taken from the recent satellite mission GOCE (e.g., Pail et al., 2011). In particular, we used the satellite-only global Earth model GOCO03S (http://www.goco.eu/) and computed the gradients up to degree and order 220 (lateral resolution of about 90 km) at the satellite height (255 km) using a spherical harmonics synthesis code (Prof. Z. Martinec, DIAS, personal communication) (Fig. 4E1–6). The GOCE gravity gradients are referred to either the gradiometer reference frame (GRF) or the local-north oriented frame (LNOF). LNOF is a right-handed local Cartesian system with its X axis pointing N, its Y axis pointing W, and its Z axis pointing radially outwards, defined with respect to spherical coordinates (Fuchs and Bouman, 2011). LitMod works in a user-defined Cartesian reference frame (MRF) that assumes a specific map projection (e.g., Mercator, UTM). In particular, the gravity gradients are computed in the MRF and projected back onto a geocentric spherical system. However, as recently pointed out by Bouman et al. (2013), the difference in orientation between LNOF and MRF has to be taken into account.

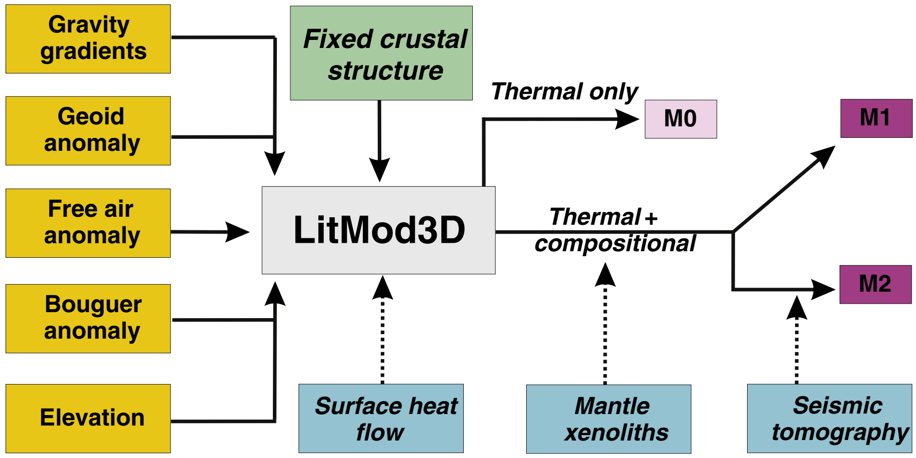

Fig. 5. Scheme illustrating the modeling process. Yellow and blue boxes show input data and data included in the models respectively. Output lithospheric models are shown by the light (thermal only, M0) and dark (thermal and compositional, M1 and M2) purple boxes, see the text for further details.

Therefore, the original data from GOCO03S model are rotated from LNOF to MRF assuming a UTM-based Cartesian system (x➔E, y➔N, z➔up) following Bouman et al. (2013), and then used as input data in the models.

5. The method: integrated geophysical–petrological modeling of the lithosphere

While the method used in the present study has been described in detail elsewhere (Afonso et al., 2008; Fullea et al., 2009), here we present a general overview of the fundamentals for completeness, with a special focus on the topics relevant to our study. A general scheme of the method is shown in Fig. 5.

5.1. The lithosphere–asthenosphere boundary

The lithosphere–asthenosphere boundary (LAB) has been characterized according to different geophysical and geochemical parameters: seismic velocities, seismic anisotropy, temperature, composition, electrical resistivity (e.g., Plomerova et al., 2002; Eaton et al., 2009; Muller et al., 2009; Fischer et al., 2010; Jones et al., 2010; Fullea et al., 2011, 2012; Kind et al., 2012; Yuan and Romanowicz, 2010). It divides the outermost, cold, relatively rigid layer of the Earth (lithosphere) from the warmer and rheologically weaker sub-lithospheric or asthenospheric mantle. A number of different reasons have been proposed to explain the weak rheological nature of the asthenosphere, most notably partial melt and/or the presence of relatively small amounts of water (e.g., Anderson and Sammis, 1970; Hirth and Kohlstedt, 1996; Karato, 2012; Kawakatsu et al., 2009; Schmerr, 2012). In the context of this paper we adopt the definition of the LAB based primarily on the temperature and compositional distributions. Therefore, we assume that the lithospheric mantle is defined: i) thermally, as the portion of the mantle characterized by a conductive geotherm, and ii) compositionally, as the portion of the mantle characterized by a, generally, different (normally, more depleted) composition with respect to the fertile primary composition in the sub-lithosphere (i.e., PUM in Table 1).

5.2. The geotherm

The lithospheric geotherm is computed under the assumption of steady-state heat transfer in the lithospheric mantle, considering a P-T-dependent thermal conductivity (Afonso et al., 2008; Fullea et al., 2009). Between the lithosphere and sub-lithospheric mantle we assume a “transition” region (a buffer layer) with variable thickness and a continuous linear super adiabatic gradient (i.e., heat transfer is controlled by both conduction and convection, see Fullea et al. (2009) for details). Below the buffer layer the geotherm is given by an adiabatic temperature gradient forced to be in the range 0.35–0.6 °C/km. When this condition is not held, i.e. for thick (>160 km) or thin (<60 km) lithospheres, the temperature at the base of the model (400 km) is allowed to vary laterally in order to keep the thermal gradient within the prescribed range. This condition typically translates into maximum lateral temperature variations at 400 km depth of ~120 °C, in agreement with predictions from seismic observations regarding the topography of the 410-km discontinuity (e.g., Chambers et al., 2005, and references therein).

5.3. Thermodynamic framework

Stable mineral assemblages in the mantle are calculated using a Gibbs free energy minimization as described by Connolly (2005). The composition is defined within the major oxide system NCFMAS (Na_2O–CaO–FeO–MgO–Al_2O_3–SiO_2). All the stable assemblages in this study are computed using a modified/augmented version of the Holland and Powell's (1998) thermodynamic database (Afonso and Zlotnik, 2011). The density and seismic velocities in the mantle are determined according to the elastic moduli and density of each end-member mineral as described by Connolly and Kerrick (2002)

Table 2
Geophysical properties of the different crustal bodies used in the 3D model.

Layer	Density (kg/m^3)	Heat production (W/m^3)	Thermal conductivity (W/m K)
Newer series granites	2670	$4 \cdot 10^{-6}$	3.5
Mafic volcanics	2920	$0.5 \cdot 10^{-6}$	2.3
Upper–middle crust	2780	$1 \cdot 10^{-6}$	2.5
Lower crust	3050	$1 \cdot 10^{-6}$	2.5

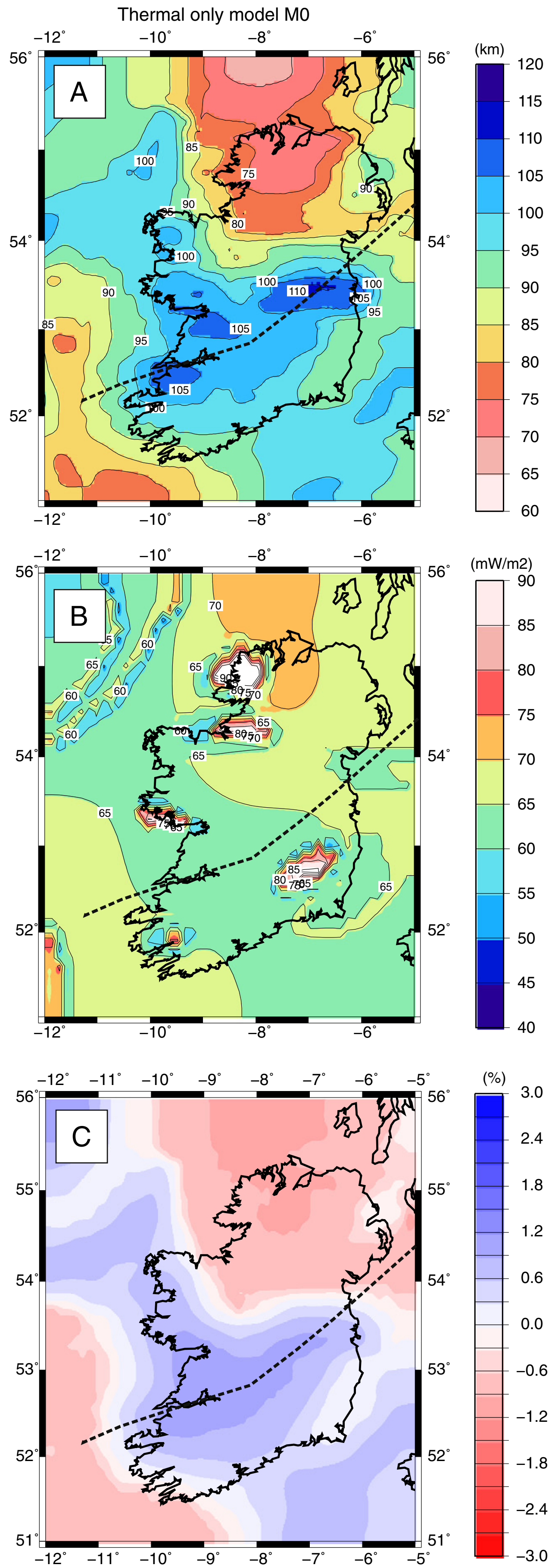

Fig. 6. Lithospheric model M0. A) Lithosphere–Asthenosphere Boundary depth. B) Predicted surface heat flow. C) Synthetic P-wave velocity anomaly map at 80 km depth. A reference average velocity of 8 km/s has been subtracted to compute the anomalies. Anelasticity effects are included in the synthetic seismic velocity anomalies considering the dominant frequencies of the seismic tomography study (~1–0.5 Hz) and an average grain size of 1 cm (see text for further details). The thick dashed line is the Iapetus Suture Zone dividing Laurentia and E. Avalonia.

and Afonso et al. (2008). Anelasticity effects are of primary importance, particularly at high temperatures (e.g., Karato, 1993; Sobolev et al., 1996; Goes et al., 2000; Cammarano et al., 2003; Afonso et al., 2010). We compute anelasticity as a pressure–temperature-dependent correction to the anharmonic output velocities (e.g., Afonso et al., 2005; Karato, 1993; Minster and Anderson, 1981). The effects of other parameters (i.e., grain size, activation volume) on the anelasticity correction applied here have been explored elsewhere (Fullea et al., 2012).

5.4. Forward modeling: geophysical observables

The coupled bulk density and pressure distributions in the mantle are computed following an iterative procedure (see Appendix B in Fullea et al., 2009). Once the temperature and density distributions within the 3D model domain are derived, we calculate a variety of geophysical observables: surface heat flow, elevation, geoid anomalies, gravity anomalies and gravity gradients. These theoretical/predicted values are then compared to measured data in order to validate lithospheric models. Therefore, LitMod works on a trial-and-error, forward modeling basis.

The predicted surface elevation in each model column is determined according to local isostasy by integrating the crustal and mantle densities over a 400-km-thick column from the surface down to the base of the model. We do not explicitly consider dynamic topography, i.e., contributions to topography related to mantle convection are assumed negligible (Fullea et al., 2009).

Synthetic gravity gradient and gravity and geoid anomaly data are computed by adding the individual contribution of each of the vertical prisms of constant or linearly varying density in which the model is discretized. Surface heat flow values are determined according to the local model geotherm and crustal heat production. For more details regarding the formulae and the discretization scheme the reader is referred to Fullea et al. (2009). The numerical approach adopted in this work is based on a planar approximation (Cartesian reference frame) that assumes a specific map projection (e.g., Universal Transverse Mercator or Mercator projection). This is generally a good approximation to model gravity and geoid data referred to the surface of the Earth if the lateral dimensions of the domain are relatively reduced. In the case of gravity gradients at the satellite height the spherical geometry of the Earth should be considered in relatively large regions (e.g., Bouman et al., 2013). However, for the purposes of this work (relative differences between alternative lithospheric models) and the relatively small size of our study region, a planar approximation is sufficient. Effects related to the spherical nature of the Earth represent only a relatively small (same order of magnitude as global Earth model GOCO03S at the satellite height, <1 mE) and constant systematic error in our models. The lateral and vertical resolution of the models in this work is of 10 km and 2 km respectively, corresponding to a grid of $50 \times 50 \times 200$ nodes in x, y and z axis respectively.

6. Results

In order to isolate the contribution to the geophysical observables from the lithospheric mantle (i.e., the thermal and compositional lateral variations), we have assumed a fixed crustal structure based on previous geophysical and petrological studies (see Section 3.1). Fixing crustal structure allows us to explore a range of different thermal and compositional lithospheric mantle models, and to compare their outputs against measured data (i.e., geophysical observables, Section 4) and other

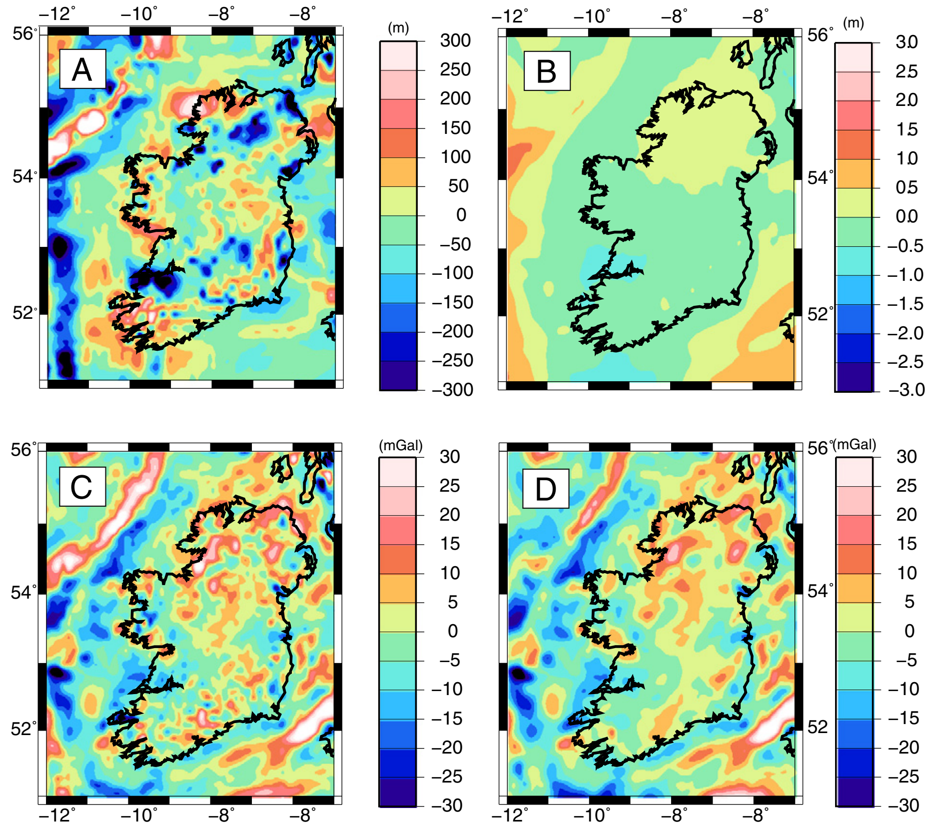

Fig. 7. Residuals for the lithospheric model M0 (see Table 3) A) Elevation. B) Geoid anomaly. C) Free air anomaly. D) Bouguer anomaly.

geophysical studies (i.e., seismic tomography, Section 3.2.1) (Fig. 5). To generate a sensible initial model we have used the 1D inversion of geoid anomaly and elevation data presented by Fullea et al. (2007). The method proposed by these authors assumes a two-layer lithospheric model composed of crust and lithospheric mantle in which the crustal density varies linearly with depth and the lithospheric mantle density is a function of temperature only. In spite of its limitations (i.e., 1D approach, no compositional effects), the method of Fullea et al. (2007) is a reasonable starting point for a full 3D model, as the lithospheric structure resulting from the inversion reproduces, to the first order, most of the long wavelength patterns observed in the geophysical data.

6.1. Crustal model

The crustal model considered in this study consists of two main layers: an upper–middle crust and a lower crust. Although there are fine-scale differences in the Irish crust north and south of the ISZ (e.g., O'Reilly et al., 2010), the bulk composition and vertical stratification in the crust seem to be rather constant across the Laurentian and E. Avalonian terranes (Hauser et al., 2008). The thickness of the lower crust is approximately one third of the total thickness, i.e., around 10 km (e.g., Landes et al., 2005 and references therein). The average Vp in the upper–middle crust is around 6.2 km/s (e.g., Hauser et al., 2008) which corresponds with an average density of 2780 kg/m^3 based on Vp-density empirical relationships (Christensen and Mooney, 1995). The lower crust in central Ireland is sampled by the Irish (Clare Castle) xenolith suite, which consists mainly of granulite facies metapelites (van den Berg et al., 2005, see Fig. 1 for location). The average measured density of the Irish xenoliths is 3100 kg/m^3 (van den Berg et al., 2005) but this value probably overestimates the actual density of the lower crust, which is not likely made up exclusively of metapelites. Other more fertile (and less dense) bulk compositions, like psammite or orthogneiss, may be underrepresented in the xenolith suite due to their less refractory nature making them less prone to assimilation by the entraining magma (van den Berg et al., 2005).

The initial Moho depth map obtained from 1D inversion of geoid anomaly and elevation data (Fullea et al., 2007) has been modified so as to match all existing seismic constraints in Ireland (see Section 3.1).

Table 3
Statistics of the residuals (geophysical observables) for models M0, M1 and M2.

Std. dev residuals	Thermal model M0 Comp. 1	Model M1[a] Dom. 4: comp. 2	Model M1[a] Dom. 4: comp. 3	Model M2[a] Dom. 4: comp. 1	Model M2[a] Dom. 4: comp. 5
ΔTopo (m)	106.4	110.7	115.9	110.6	113.6
ΔFA (mGal)	9.82	9.12	9.17	8.87	9.06
ΔBouguer (mGal)	8.69	8.94	8.94	8.62	8.83
ΔGeoid (m)	0.31	0.29	0.3	0.32	0.29
ΔUzz (mE)	33.5	25.8	37.9	27.5	24.1
ΔUxx (mE)	15.9	14.8	21.7	18.5	16
ΔUyy (mE)	42.1	25.8	27.1	21.7	24.7
ΔUzx (mE)	58.1	52.8	43.9	57	52.7
ΔUzy (mE)	51.4	31.8	41.7	34.3	28.7
ΔUxy (mE)	35.2	24.8	30.1	21.6	22.7

[a] In Models M1 and M2 lithospheric domains 1, 2 and 3 (see Fig. 8) have compositions 2, 1 and 2 respectively (see Table 1 and the text for further details).

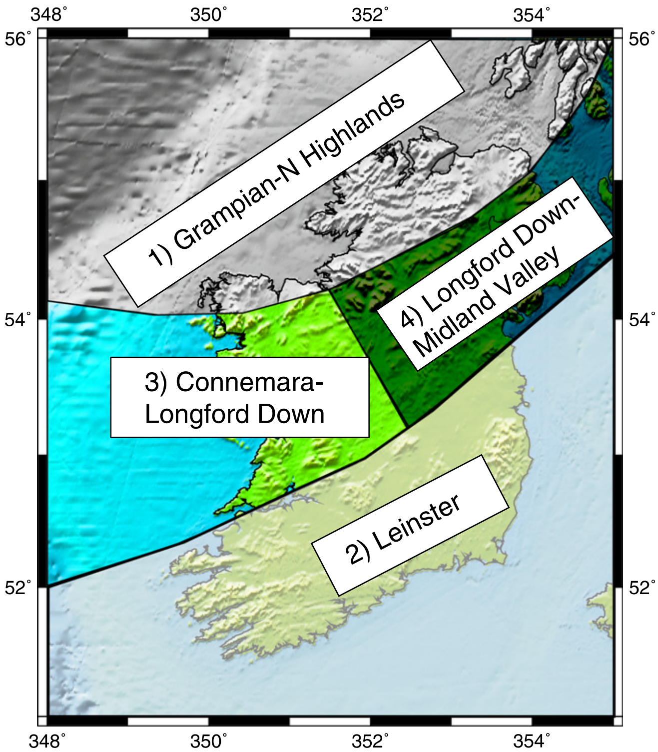

Fig. 8. Lithospheric mantle compositional domains based on crustal tectonics, petrological and geophysical considerations (see text for further details).

The Moho depth map together with the seismic constraints used to build it is shown in Fig. 2A. The crustal thickness and its internal structure defined in this way is considered to be an a priori constrained model feature in this study and is therefore kept fixed. The fixing of crustal structure allows us to isolate and focus on the lithospheric mantle temperature and chemical variations. Due to its large distorting effect in the potential fields (particularly in gravity anomalies), two additional crustal layers were added to represent the Newer Granite series (Leinster, Galway and Donegal granites) and mafic volcanic intrusions on the eastern Irish margin (Fig. 2B and C). The presence of these two aforementioned near surface geological structures, characterized by a density distinct from that of the background crust, is clearly visible by simple inspection of the gravity signal (granites observed as local minima and mafic intrusions as local maxima, see Fig. 4). Although not listed in the Newer Granite series in Ireland as such, the Killarney gravity low (located in the SW) and the Ox mountains complex (in the NW) have been interpreted and modeled as buried granitic bodies on the basis of their gravity and seismic (i.e., low velocity) signatures (e.g., Hauser et al., 2008; Lowe and Jacob, 1989; Readman et al., 1997). The physical properties of the crustal layers considered in this work are listed in Table 2.

6.2. Lithospheric models

The LAB depth derived from 1D inversion of geoid anomaly and elevation data (Fullea et al., 2007) is taken as a first approximation of the lithospheric geometry and is modified in successive steps to obtain a best fitting 3D model of the long-wavelength component of the geophysical observables. As a starting point, we define a purely thermal lithospheric model (M0) with no lateral changes in composition (i.e., a homogenous mantle composition). The mantle composition assumed for M0 in the whole lithospheric domain is based on garnet-bearing xenoliths and garnet xenocrysts (composition 1, Table 1). This composition is representative of the southwestern European Phanerozoic mantle (e.g., Fullea et al., 2010; Griffin et al., 2009) and is only moderately depleted compared to the sublithospheric mantle, which is assumed to have the PUM (Primitive Upper Mantle) composition of McDonough and Sun (1995) throughout the modeling process (composition 6, Table 1). The final lithospheric structure of model M0 is characterized by: i) a moderately thick lithosphere (≈105–110 km) with a SW-NE trend from the Clare basin to the eastern Irish margin near Dublin, roughly along the ISZ, and ii) a lithospheric thinning from central

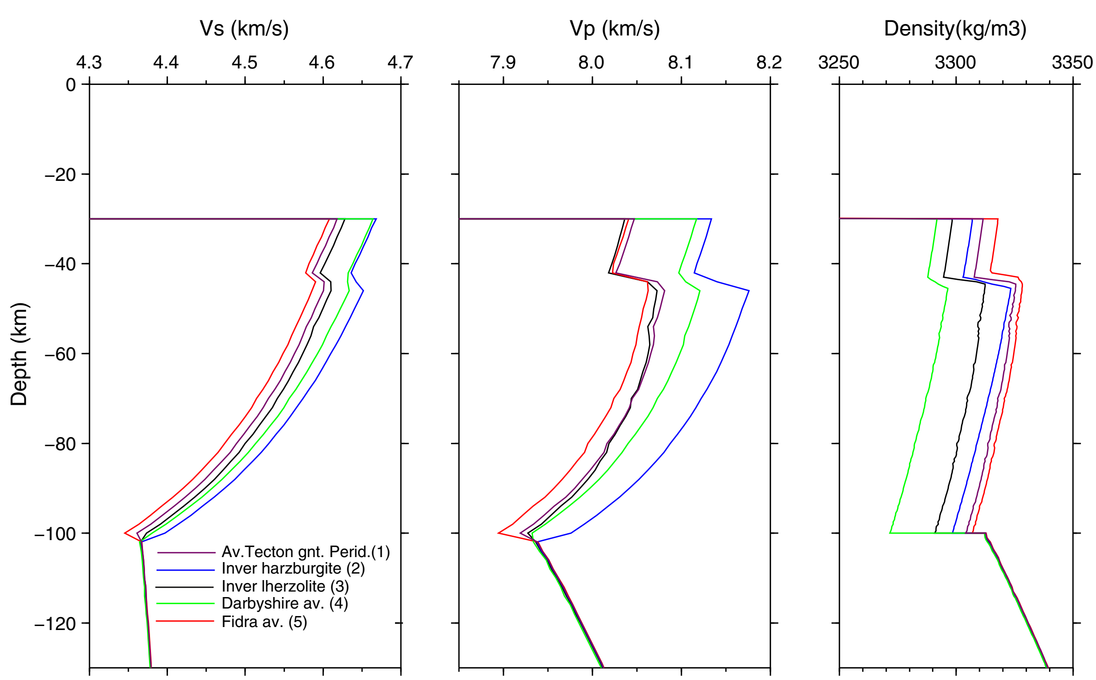

Fig. 9. Lithospheric vertical profiles of the seismic velocities and density associated with the mantle compositions discussed in this work. The temperature distribution corresponds with a 30-km-thick and 100-km-thick crust and lithosphere respectively (see Table 2 for the thermal parameters in the crust). The black dotted, blue, solid black, green and red lines correspond with lithospheric mantle compositions 1–5 in Table 1 respectively.

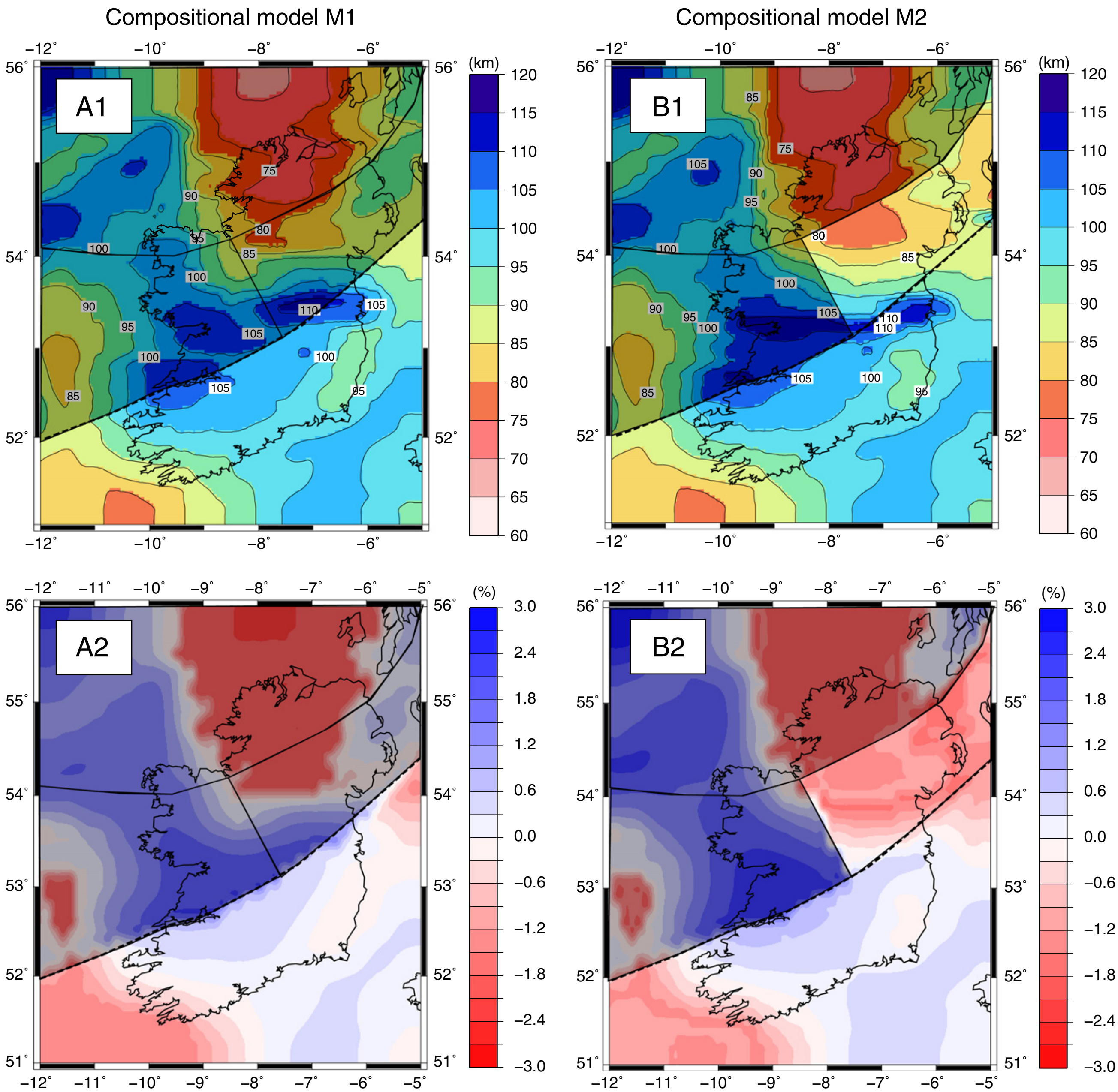

Fig. 10. Lithospheric models M1 (A1 and A2) and M2 (B1 and B2). A1) and B1) Lithosphere–Asthenosphere Boundary depth. A2) and B2) Synthetic P-wave velocity anomaly map at 80 km depth. A reference average velocity of 8 km/s has been subtracted to compute the anomalies. Anelasticity effects are included as discussed in Fig. 6. The mantle compositional domains defined in Fig. 8 are shown. Dark areas represent compositional domains with relatively depleted compositions (i.e., compositions 2 and 3 in Table 1). The thick dashed line is the Iapetus Suture Zone dividing Laurentia and E. Avalonia.

Ireland towards the NW Atlantic margin of Ireland (80–65 km thickness) (Fig. 6A). This model fits the long wavelength component of the geophysical observables reasonably well (Fig. 7, Table 3). The synthetic surface heat flow predicted by model M0 is shown in Fig. 6B. The overall trend observed in the synthetic data ranges from 65 mW/m^2 in the south to >70 in the NW, following the lithospheric thinning trend. Local maxima (80–95 mW/m^2) are associated with granite bodies with high radiogenic heat production (Fig. 2B, Table 2). The moderate S-N increase of <10 mW/m^2 in the synthetic surface heat flow is at odds with the map published by Goodman et al. (2004) based on a selection of measured and estimated data. Their map (Fig. 6 in a companion paper by Jones et al., 2013) shows a large positive gradient indicative of very high heat flow in northern Ireland (>80 mW/m^2), low heat flow in southern Ireland (<50 mW/m^2), and somewhat high values in the middle (~70 mW/m^2). However, Jones et al. (2013) have carefully revised the surface heat flow data available in Ireland, including measurements and estimates (see their Fig. 4), and have produced a new interpolated and filtered heat flow map of Ireland excluding anomalous surface heat flow estimates, all measurements in granites except for one on the Galway Granite, and the Portmore borehole on the northern coast (i.e., their Fig. 5). Their map suggests a moderate increase in heat flow in the center of Ireland from some 55–60 mW/m^2 to some 70 mW/m^2, in general agreement with the surface heat flow predicted by model M0 (Fig. 6B). We note, however, that although the surface heat flow observations are useful as broad constraints, they are not heavily weighted as constraining

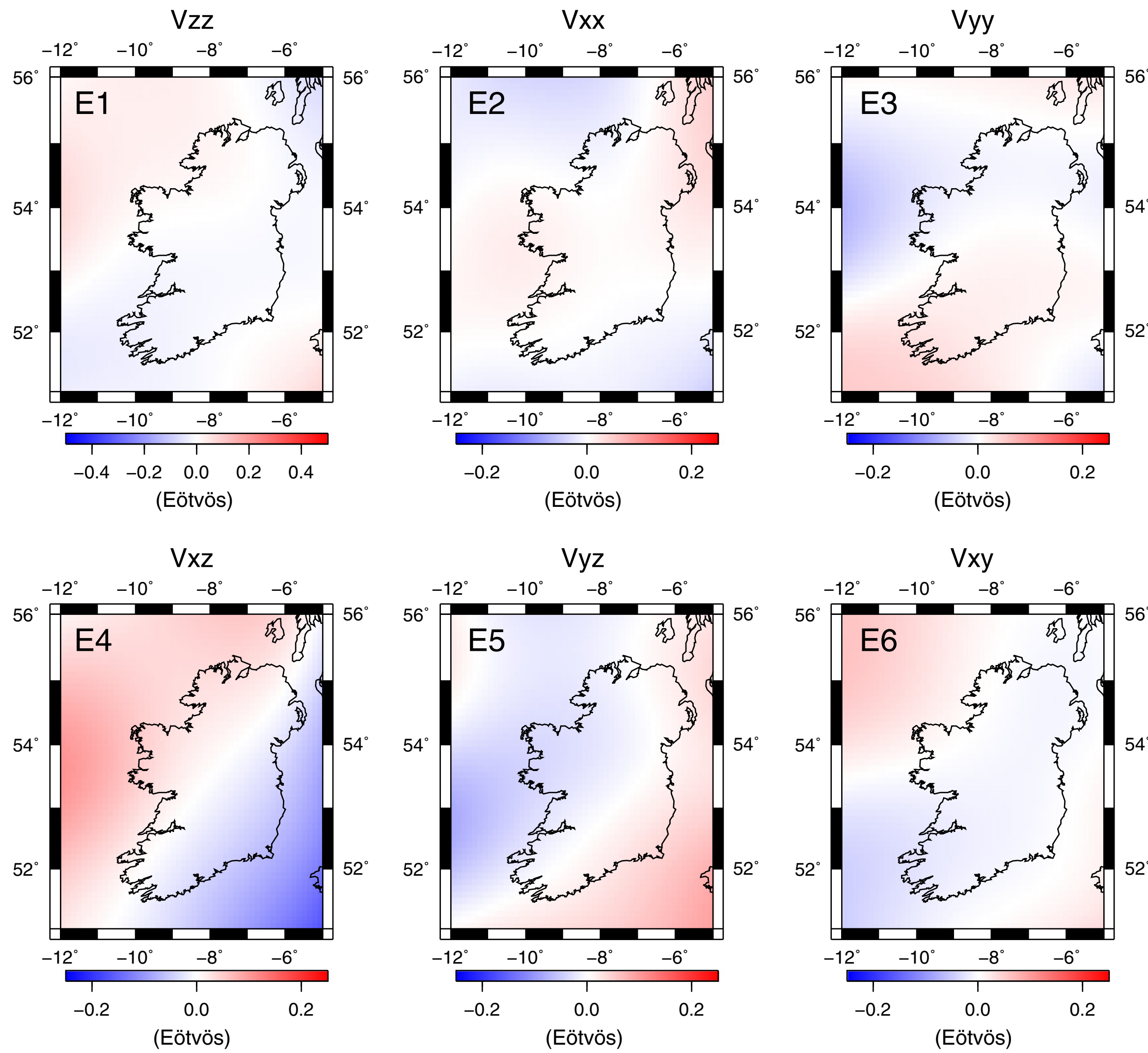

Fig. 11. Residuals for lithospheric compositional model M2 (composition 5 in domain 4, Fig. 8 and Table 1). A1–A6) Gravity gradients computed at 255 km above the sea level (GOCE mission perigee height). The sub-indexes of V denote second derivatives of the Earth's potential (i.e., gravity gradient) along the west-east, south-north and vertical directions (x, y, z respectively, in the MRF, see Section 4 for more details).

data in this study due to i) the sparsity of the data available, and ii) the local geological variations (e.g., in thermal conductivity) that might be affecting them and that are out of the scope of this regional-scale study.

In order to compare the lithospheric models with seismic tomography studies (Fig. 3) seismic attenuation effects need to be taken into account. To compute anelasticity we use a pressure- and temperature-dependent formula (e.g., Afonso et al., 2005; Karato, 1993; Minster and Anderson, 1981) for the relevant reference periods (i.e., 1 s and 50 s for P- and surface-wave tomography models, respectively) and a reference grain size of 10 mm. For further discussion on the effects of different grain sizes (and other standard parameters) on seismic attenuation the reader is referred to Fullea et al. (2012). Fig. 6C shows a slice of the synthetic seismic velocity anomalies predicted by model M0 at a depth of 85 km. The seismic velocity distribution predicted by M0 depends only on the thermal structure and clearly reproduces some of the main features observed in published tomography models: high velocities in central western Ireland (Galway and Clare basin) and Irish Sea near Dublin, and low velocities in NW Ireland (Fig. 6C). The model, however, fails to reproduce the negative velocity anomalies observed in central and SW Ireland (cf. Figs. 3A and 6C). In the following we explore compositional variations within the lithospheric mantle as a potential source of the seismic velocity structure observed in Ireland (Fig. 5).

6.2.1. Compositional domains

To study potential compositional heterogeneities within the Irish lithospheric mantle we define four broad domains based on crustal tectonics, petrological and geophysical considerations: 1) Grampian-N Highlands, 2) Leinster, 3) Connemara–Longford Down, and 4) Longford Down–Midland Valley (Fig. 8). While the E. Avalonian plate in the south is solely represented by domain 2 (including the areas affected by the Variscan orogeny), the Laurentian terrane is split into 3 domains (i.e., 1, 3 and 4) some of which have their counterparts in Scotland, east of the Irish Sea (e.g., Grampian, Midland Valley, Longford Down terrane corresponds with the Southern Uplands in Scotland). Some of the boundaries of domains 3 and 4, in contrast to those of domains 1 and 2, do not exactly correspond to the usual tectonic divisions as defined by crustal geology. Instead, the NW-SE boundary between domains 3 and 4 has been set to capture a geophysical feature common in most seismic studies: the contrast between fast (west) and slow (east) mantle in central Ireland (Wawerzinek et al., 2008; O'Donnell et al., 2011; Legendre et al., 2012, Fig. 3). Domain 1 is the only one where direct sampling of the mantle composition is available through xenoliths: those represented by the Inver harzburgites and lherzolites (compositions 2 and 3 in Table 1 respectively). In the other domains (i.e., 2, 3 and 4) reasonable assumptions regarding the bulk composition have to be made due to the lack of direct constraints. In domains 3 and 4

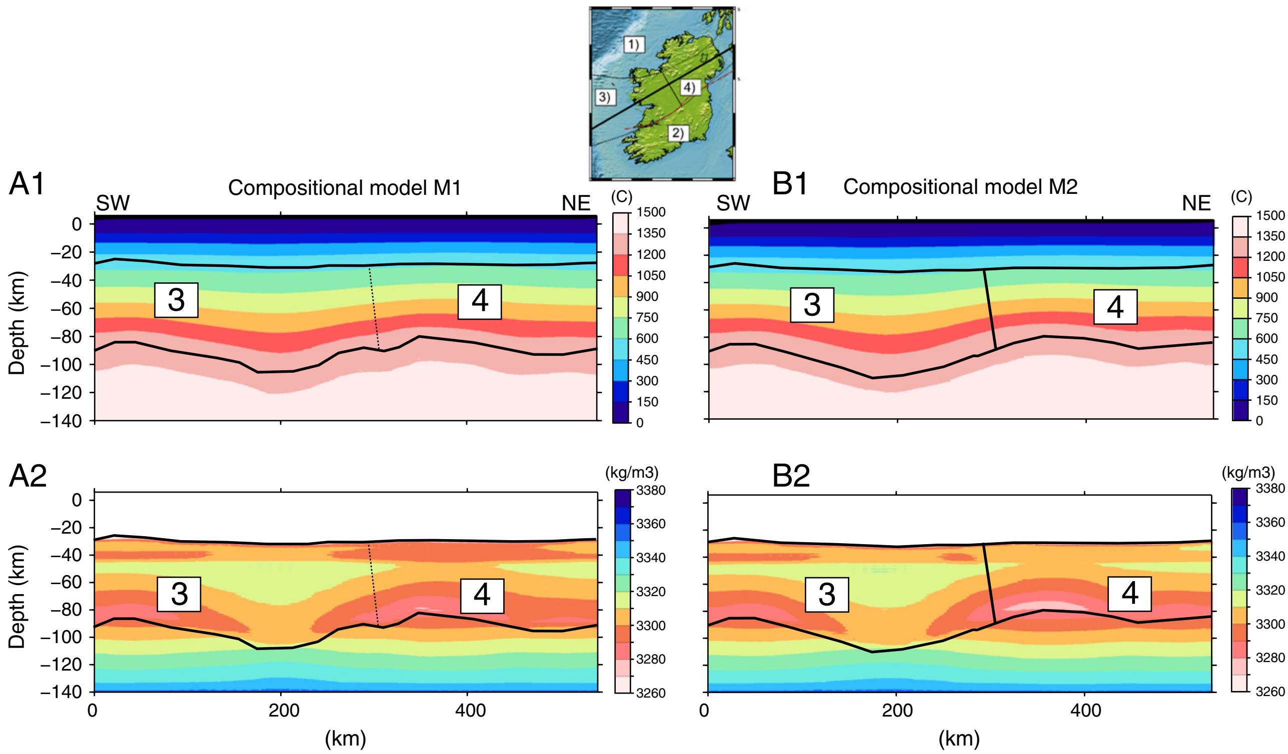

Fig. 12. SW-NE lithospheric cross-sections of compositional models M1 (left) and M2 (right). A1 and A2) temperature distributions. B1 and B2) Mantle density distributions. Solid black lines show the Moho and LAB depth for each model. The quasi-vertical black solid and dotted lines represent a boundary between two mantle domains: Connemara-Longford Down (3, left) and Longford-Down–Midland Valley (4, right).

(Connemara–Longford Down–Midland Valley terranes) we tested compositions 2 and 3 based on geographical proximity (to the Inver xenolith locality), as well as composition 1 (average tecton garnet peridotite) and composition 5 (xenoliths from the Midland Valley terrane in Fidra, Scotland) (Table 1). In the Leinster terrane (domain 2) two compositions were evaluated: a generic Phanerozoic composition taken from global averages (composition 1, Table 1), and that derived from the Derbyshire xenolith suite (England), the closest Avalonian xenolith locality to Ireland (composition 4, Table 1).

A number of parallel numerical experiments have been run to assess the mantle composition of domains 1, 2, 3 and 4 (Fig. 8). In each of those experiments the crustal structure was kept fixed whereas the lithospheric thickness (i.e., the thermal structure) was allowed to vary laterally so as to attain the best possible match with the long wavelength component of the geophysical observables (Fig. 4). In the Grampian terrane (domain 1), composition 2 (harzburgite) leads to an overall better fit to the measured geophysical data than composition 3 (lherzolite). In particular, composition 3 (Mg# = 90.42, Table 1) produces a lighter lithospheric mantle than composition 2, which is slightly more depleted (Mg# = 90.58, Table 1). Therefore, if mantle composition 3 is assumed in the Grampian terrane, then the lithospheric thickness must be increased (i.e., to increase the average mantle density) to keep the predicted topography and potential fields within the range of observed values. However, even if the lithospheric thickness (i.e., temperature) is allowed to change (without modifying the crustal structure), the final fit to the measured gravity field is slightly worse. In the southern half of Ireland (Leinster terrane, domain 2) composition 4 (Derbyshire, Mg# = 91.47, Table 1) predicts a rather low lithospheric density (Fig. 9) that translates into a synthetic topography several hundred meters above the real value. In such a model scenario (composition 4 in lithospheric domain 2), the LAB depth would have to be considerably increased in order to increase the average lithospheric mantle density. That would depress the mantle geotherm, leading to a positive seismic velocity anomaly underneath the Leinster domain that is at odds with the images depicted by local seismic tomography studies (Wawerzinek et al., 2008; O'Donnell et al., 2011). In contrast, a relatively fertile (and dense) composition (e.g., 1 or 5, Table 1) would be a more suitable candidate to represent the average mantle composition in the Leinster domain than the depleted composition 4. We acknowledge that the mantle compositional structure outlined above relies heavily in the fundamental assumption of an a priori fixed crustal structure. Trade-offs between the crustal and mantle structures (temperature and composition) are an inherent limitation in most lithospheric models, particularly in those based on potential field data. To address this issue (specifically, if composition 4 is a reasonable average in domain 2), tests were conducted allowing the a priori crustal structure to vary within permissible bounds. The main conclusion of these tests is that it is not possible to simultaneously fit all the geophysical observables without changing significantly the crustal structure derived from seismic data (Fig. 2A). It is therefore concluded that composition 4 cannot be a representative average of the mantle compositions in the Leinster lithospheric domain. For a detailed sensitivity analysis describing how the synthetic elevation and surface heat flow are affected by varying the crustal parameters within acceptable bounds, the reader is referred to the companion paper by Jones et al. (2013).

From the many tests conducted varying the bulk composition of the different mantle domains as well as the thermal structure (i.e., the LAB depth), two end-member lithospheric models stand out. The first type of models, M1 hereafter, are characterized by a SW-NE trending chemical boundary (along the ISZ) that separates a relatively depleted (i.e., compositions 2 or 3, Table 1) Laurentian terrane (i.e., domains 1, 3 and 4) from the fertile (i.e., compositions 1 or 5) E. Avalonian plate (i.e., Leinster domain). The second type of models, M2 hereafter, are defined by a "z-shaped" compositional boundary dividing Ireland into a northwestern depleted zone (i.e., compositions 2 or 3, Table 1) within the Laurentian plate (i.e., domains 1 and 3), and a southeastern fertile (i.e., compositions 1, 5 and 6, Table 1) area comprising the E. Avalonian plate (Leinster terrane) and the Midland-Valley Laurentian terrane

(Fig. 10). Both model families M1 and M2 match the long wavelength component of the topography, geoid and gravity anomalies with comparable accuracy (cf. residuals of M0, M1 and M2 in Table 3) and predict a surface heat flow distribution very similar to that of Model M0 (Fig. 6B). However, models M1 and M2, characterized by a N-S compositional change, match most of the gravity tensor components computed at the satellite height better than the purely thermal, compositionally homogeneous model M0 (Fig. 11, Table 3). Although models M1 and M2 show a similar misfit for all the geophysical observables used here, differences in their compositional structure are appreciable. In terms of the mantle density, models M1 and M2 are rather similar, with only small variations of the thermal LAB depth (Fig. 12). Models M2 are characterized by a moderate lithospheric thinning in the Midland-Valley domain, with respect to M0 (Fig. 6), and a negative velocity anomaly in NE Ireland and the North Channel between Northern Ireland and Scotland (Figs. 10B and 12). However, the situation in the model family M1 is the opposite, i.e., slight lithospheric thickening in the Midland-Valley and a neutral to positive velocity anomaly in NE Ireland (Fig. 10A). Furthermore, the E (negative)-W (positive) seismic velocity anomaly contrast in central Ireland (cf. Fig. 3) is more clearly visible in models M2 than in models M1. Such a velocity contrast is related to chemical and thermal variations within the lithosphere from Connemara to Midland Valley in the Longford-Down terrane (Figs. 8 and 10). The compositional boundary imposed in models M2 to reproduce published seismic anomalies in central Ireland (O'Donnell et al., 2011; Wawerzinek et al., 2008) implies a relatively depleted mantle composition in Connemara (e.g., composition 2, Table 1) compared to that of Midland Valley (Fig. 12). However, this moderately depleted composition it is not as depleted as composition 4, which is comparable to composition 2 in that the contents of the MgO, CaO and Al_2O_3 oxides are similar for the two compositions, but the former has a lower amount of FeO (Table 1). We note that assuming composition 4 (Table 1) in Connemara (domain 3) would result, as is the case for composition 2, in mantle velocities faster than those associated with fertile compositions in the Midland Valley (e.g., 1 or 5, Table 1) (Fig. 9). However, that would predict very low mantle densities in Connemara thus preventing the model to match the observed data without significant changes in the crust. This is not the case for composition 2, which shows the fastest seismic velocities of all the compositional range explored in this study and, at the same time, density values comparable to that of fertile compositions (i.e., 1 or 5, Table 1) (Figs. 9 and 12).

7. Discussion and conclusions

The geophysical–petrological best fitting models (M0, M1 and M2) depict a present-day Irish lithospheric structure characterized by moderate lateral density variations. This is consistent with the rather flat pattern observed in the geophysical data modeled in this work (Figs. 1 and 4). From a thermal point of view, the main structural characteristics of the best fitting models are (1) a slightly thickened lithosphere along the ISZ, and (2) a progressive lithospheric thinning from central Ireland towards the north (Figs. 6 and 10). In terms of mantle composition, the lithosphere south of the ISZ in the E. Avalonian plate is relatively and uniformly fertile (i.e., typical Phanerozoic mantle). In contrast, the lithospheric composition in the Laurentian plate seems to vary from moderately depleted to fertile in some of the preferred models (M1 and M2) (Figs. 8 and 10). In particular, the Grampian and Galway domains (Fig. 8) are defined by moderately depleted, Fe-rich compositions such as those derived from the Inver xenolith suite (see location in Fig. 1, Table 1). In contrast, the lithospheric mantle in the northeastern Midland Valley domain (Fig. 8) has to be more fertile, with a composition likely similar to that derived from xenoliths in the neighboring Midland Valley of Scotland (e.g., Fidra xenolith suite, Table 1), if the seismic velocity anomalies from local tomography in central Ireland (O'Donnell et al., 2011; Wawerzinek et al., 2008) are to be matched (i.e., models M2).

All the best fitting models are able to reproduce, to first order, the main features of the geophysical observables used here, although models with an N-S compositional variation (M1 and M2) show an overall better fit to gravity gradients computed at the satellite altitude (255 km) (Figs. 7 and 11, Table 3). Compositional models M2 and M3 reproduce the data with similar accuracy and a rather similar lithospheric density structure (Fig. 12, Table 3). However, models M2 are able to reproduce the seismic velocity pattern observed in Ireland better than M1 (Figs. 3 and 10). Therefore, compositional models M2 are preferred over the others (M0 and M1) on the basis of their global fit to the constraining data used in this work (Fig. 5).

The thermal and compositional structure of the present day stable Ireland appears to reflect its long term geological record. The closure and suturing of the Iapetus Ocean during the Caledonian orogeny involved large amounts of sinistral transpression and moderate crustal and lithospheric shortening. A long-lasting result of this soft collision process could be the moderate lithospheric root imaged along the ISZ in our models (Figs. 6 and 10). Subsequent Cenozoic magmatic activity, with the extrusion of large scale flood basalts mainly in Northern Ireland and Scotland, could be related to the imaged lithospheric thinning in northern Ireland and the compositional field within the Longford-Down terrane (Figs. 8 and 10). A plausible hypothesis, as proposed by many authors on the basis of various observations (e.g., Al-Kindi et al., 2003; Arrowsmith et al., 2005; Kiristein and Timmerman, 2000; Landes et al., 2007; Wawerzinek et al., 2008), is that a lateral branch of the Iceland mantle plume was active in Ireland (and Britain) during the Tertiary. The action of the plume impinging at the base of the lithosphere could have triggered mantle melting and therefore potential refertilization of the lithosphere in the Midland Valley domain. That would imply a change from the moderately depleted composition of the Inver xenoliths to the fertile composition of the Fidra xenolith suite. It is worth noting that, due to the Fe-rich nature of the depleted mantle in northern Ireland (composition 2, Table 1), the hypothetical refertilization process would not have strong implications in terms of lithospheric density variations (which then would be almost entirely controlled by the thermal structure), but it would certainly have an impact on the seismic velocities (i.e., decreasing them, Figs. 9 and 10). This could explain the apparent paradox of having a significant seismic velocity contrast in central Ireland and, at the same time, almost no expression in the density-dependent geophysical observables (i.e., elevation, potential fields, Figs. 1 and 4).

Acknowledgments

JF was supported by the JAE-DOC programme (CSIC-Spain) cofounded by ESF, and by the Spanish Ministry of Economy and Competitiveness grants CGL2009-13103 and CGL2012-37222. Funding for this work comes from a Science Foundation Ireland grant (10/IN.1/I3022) for the IRETHERM project (www.iretherm.ie) to AGJ that supported JF. AGJ visited JCA in Macquarie, funded by a Short Term Travel Fellowship from Science Foundation Ireland (10/IN.1/I3022-STTF11). AGJ wishes to thank JCA and the GEMOC group for their generous hospitality.

The work of JCA has been supported by two Australian Research Council Discovery Grants (DP120102372 and DP110104145). This is contribution 327 from the Australian Research Council Centre of Excellence for Core to Crust Fluid Systems (http://www.ccfs.mq.edu.au) and 892 in the GEMOC Key Centre (http://www.gemoc.mq.edu.au).

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