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Shape and size of the starting Iceland plume swell

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Abstract

Emplacement of a large igneous province is usually accompanied by kilometre-scale uplift over an area of $\sim 10^6$ km^2. We have developed a method for mapping the dynamically supported swell associated with the North Atlantic Igneous Province by inverting palaeo-topographic information from continental margins. In the forward model, latest Palaeocene palaeo-topography around Britain and Ireland is calculated by correcting present-day topography for global sea-level change, denudation and dynamic support. We initially assume a Gaussian, axially symmetric dynamic support profile. Modelled coastlines are compared with palaeo-coastlines mapped on 2D and 3D reflection seismic data. In the inverse model, the amplitude, width and centre of the dynamically supported swell are determined by minimising misfit between modelled and observed coastlines. Uncertainties associated with global sea-level variation and denudation have little effect on this calculation. The best-fit dynamic support profile from inverting palaeo-coastline positions is in good agreement with dynamic support estimates from subsidence anomalies measured in extensional sedimentary basins fringing Britain and Ireland. However, a circular planform of dynamic support cannot simultaneously account for palaeo-coastlines, subsidence anomalies and the spatial extent of the North Atlantic Igneous Province. In combination, these data suggest that the swell was more irregular in planform. This inference can be tested in future by modelling stratigraphic data from offshore Norway, Greenland and Canada. The large areal extent and short time interval for inflation of the dynamically supported swell are best explained by rapid convective emplacement of an abnormally hot mantle layer horizontally beneath the lithosphere, during the starting phase of the Icelandic convective system. We emphasise the need to interpret the igneous record jointly with the dynamic support history when discussing models of large igneous province formation and mantle convection.

Keywords: large igneous province; North Atlantic; Iceland; dynamic support

1. Introduction

A range of models has been proposed to account for the formation of large igneous provinces. A popular class of models holds that large volumes of melt are generated when hot mantle is emplaced beneath the lithosphere by mantle convection, although the nature of the convection is debated. Hot mantle may be supplied by an

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axisymmetric mantle plume conduit or by vertical sheets [1,2]. The depth of origin of these features is unclear and it may not be the same for all hotspots [3]. Rapid melting may reflect rapid upwelling of a hot starting plume head [4,5] or decompression of an existing plume head beneath a rift [6]. A starting plume head may flatten relatively slowly as it impinges on the base of the lithosphere [5], or may be injected relatively rapidly beneath the lithosphere if there exists a low-viscosity asthenosphere channel [7] or if the mantle has a non-Newtonian rheology [8]. The relative importance of thermal and compositional anomalies in the melting source is unclear [9]. Alternatively, large igneous provinces may not be related to large-scale, deep-seated convection in the mantle. A hot mantle reservoir may incubate below thick continental lithosphere [10,11]. Unusually high melt volumes may be generated by small-scale convection driven by rifting leading to continental break-up [12,13], or by variation in the upwelling process beneath mid-oceanic ridges [14], without the necessity for unusually hot mantle.

Perhaps unsurprisingly, many models for formation of large igneous provinces concentrate on explaining observations of the igneous record. However, some contentious issues may be resolved by considering both the igneous record and the record of accompanying regional uplift through space and time. Here, we present a method for estimating spatial patterns of dynamic support over large areas of the continents. We concentrate on the North Atlantic Igneous Province, which is often associated with the developing Icelandic convective system. First, we describe a method for reconstructing palaeo-topography, and show how this calculation can be inverted to yield a regional estimate of dynamic support. We develop this model in the region encompassing Britain and Ireland and then discuss implications for the whole North Atlantic domain. Next, we compare dynamic support estimates derived from NW European stratigraphy and planforms of uplift inferred from the spatial distribution of the North Atlantic Igneous Province. We conclude that this province is most likely related to the developing Icelandic convective system, and that regional uplift is best explained by rapid injection of abnormally hot mantle horizontally beneath the lithosphere.

2. Dynamic support from continental stratigraphy

Temporal and spatial patterns of dynamic support are recorded indirectly by the Cainozoic stratigraphy of sedimentary basins on the NW European continental margin. Large amounts of seismic and well data cover this region following years of hydrocarbon exploration. Dynamic support is usually recognised when post-rift sedimentary successions within extensional basins, which are expected to subside monotonically through their post-rift phase [15], preserve evidence for transient regional uplift [16,17]. These effects were not caused by global sea-level change since they were of finite spatial extent. The flexural strength of the lithosphere (<5 km in elastic thickness [18–20]) is too low to support uplift across such a large region of the NW European shelf ($\sim 10^3$ km in wavelength). Since uplift was transient, it was not entirely generated by emplacement of the large igneous province itself. Therefore regional uplift was most likely dynamically supported by unusually high temperatures in the upper mantle [16,17].

Coeval uplift and igneous activity began close to the Cretaceous–Palaeocene boundary but were confined until mid Palaeocene times to an axis stretching from Disco Island through central Greenland toward the Irish Sea [21,22]. In latest Palaeocene times, rapid and more widespread transient uplift was accompanied by more voluminous igneous activity, in particular with the development of volcanic passive margins between Europe and Greenland. Here, we map the south-eastern quadrant of the dynamically supported swell at the time of its greatest inflation. The timing of this event is constrained by detailed biostratigraphic studies of the sedimentary successions that record the regional regression [23]. Subsidence modelling of volcanic passive margins and of extensional basins indicates that the swell decayed through Early Eocene time [16, 17,24].

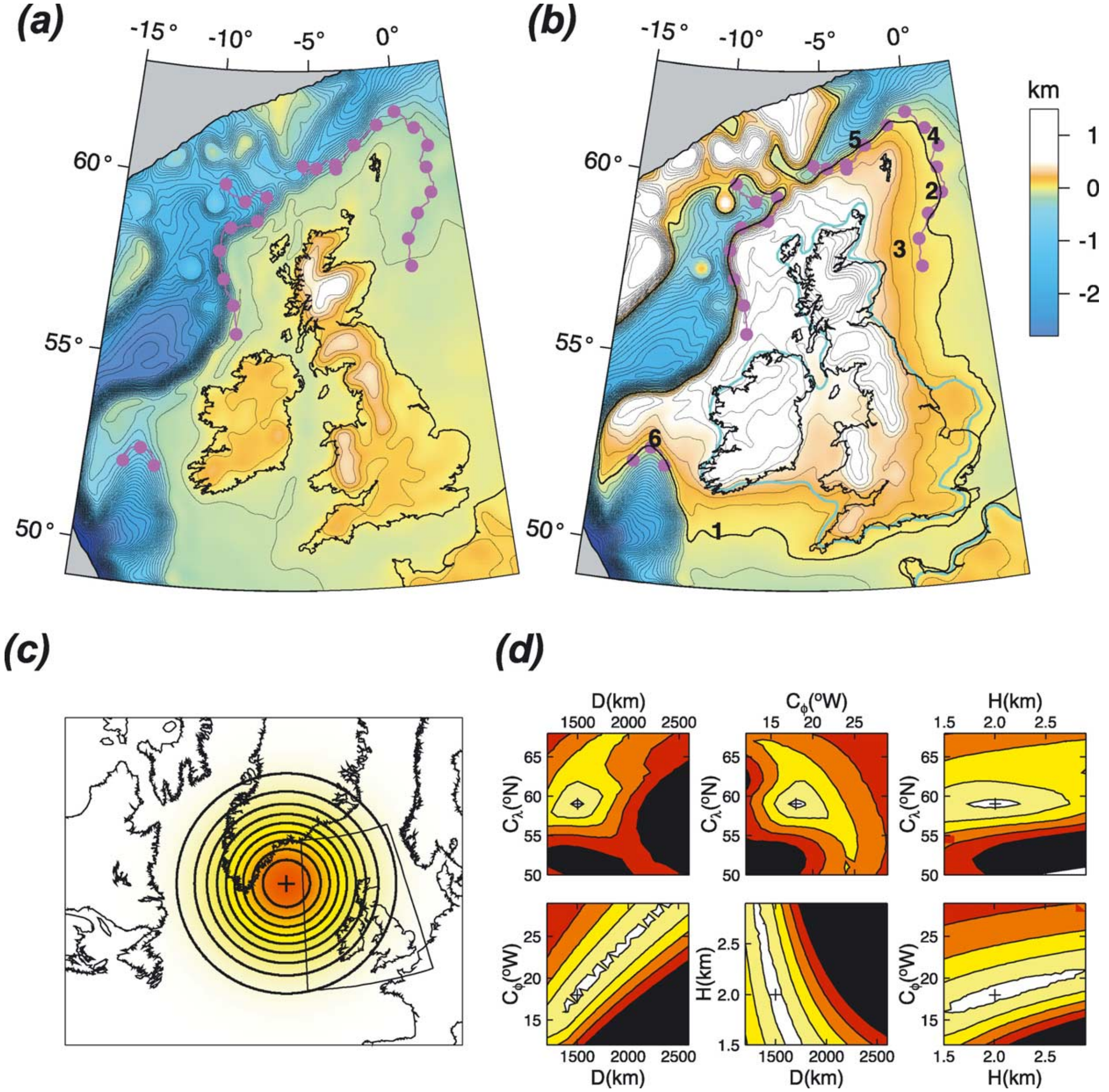

Fig. 1. (a) Present topography (from ETOPO5), low-pass filtered at 50 km to leave isostatically compensated topography; contour interval 100 m. Pink spots = latest Palaeocene coastline recognised as break between delta topsets and foresets on seismic and well-log data ([17,34,35] and unreleased BP datasets). Typical seismic reflection profiles through the deltas given in [36,47]. Grey = oceanic crust. Projection is stereographic. (b) Latest Palaeocene topographic reconstruction calculated by applying three corrections to present topography: (i) global sea-level rise of 100 m; (ii) denudation correction given by $T_D = T_0 + D(\rho_a - \rho_s)/\rho_a$ where T_D is corrected topography, T_0 is starting topography, D is denudation (Fig. 2), $\rho_a = 3.2$ Mg m^{-3} is the asthenospheric density and $\rho_s = 2.4$ Mg m^{-3} was the average density of rock now eroded; (iii) dynamic support estimate shown in (c) required to match observed and predicted coastlines. Black line = predicted coastline after applying all three corrections; blue line = coastline predicted by applying sea-level and denudation corrections only (i.e. no dynamic support). (c) Best-fitting circular Gaussian dynamic support profile used in (b), contoured at 200 m intervals from 200 m ($H = 2.0$ km, $D = 1500$ km, $(C_\lambda, C_\phi) = (59°N, 18°W)$, $\mu = 40$ m). (d) Slices through misfit function $\mu(H, D, C_\lambda, C_\phi)$, contoured at 100 m intervals from 100 m. μ measures the mean model topography per unit length at the observed coastlines.

2.1. Topographic reconstruction

Palaeo-topography can be estimated by correcting present-day topography for three effects: (a) global sea-level change; (b) variations in lithospheric thickness and density; and (c) dynamic support. This calculation forms the forward problem in this study. We estimate global sea-level change to be the sea-level rise of around 100 m that would result if the global icecaps melted. Next, we consider three groups of processes that have potentially altered lithospheric thickness and density since latest Palaeocene times: horizontal plate motions; addition of magma; and denudation.

Lack of significant Cainozoic normal faulting both onshore and adjacent offshore Britain and Ireland suggests that negligible Cainozoic rifting has occurred [25]. The most recent major rifting event to affect these regions occurred in Late Jurassic time [26–28]. This rifting finished over two lithospheric thermal time constants ago, so Cainozoic thermal subsidence can be neglected to first order during reconstruction. For the same reason, much of the rift-related topography in the base of the lithosphere should have decayed by Cainozoic time. Any ponding of hot asthenosphere by the subdued sub-lithospheric topography [29] should not significantly affect the planform of dynamic support (this issue is discussed further in Section 2.3). Structural inversion provides evidence for localised Cainozoic lithospheric shortening, particularly in southern England and offshore south of Ireland, but these structures reflect insignificant regional uplift [30]. Thus, effects of horizontal plate motions can be neglected to first order during topographic reconstruction of Britain, Ireland and the surrounding shallow-water shelf, the region containing the stratigraphic data we model here.

A second mechanism for altering the thickness and density of the lithosphere is addition of igneous rock. Palaeocene onshore volcanism was accompanied by kilometre-scale magmatic underplating of the crust beneath Britain and Ireland [30,31]. However, radiometric dating suggests that the greatest volume of onshore magmatism occurred during Palaeocene time, prior to the time of our latest Palaeocene reconstruction [31,32]. A final mechanism that has thinned the crust by up to several kilometres since the Eocene is denudation that occurred in response to regional uplift (Fig. 2) [33]. We conclude that denudation is probably the only cause of significant variation in the lithospheric template that needs to be accounted for when reconstructing latest Palaeocene topography.

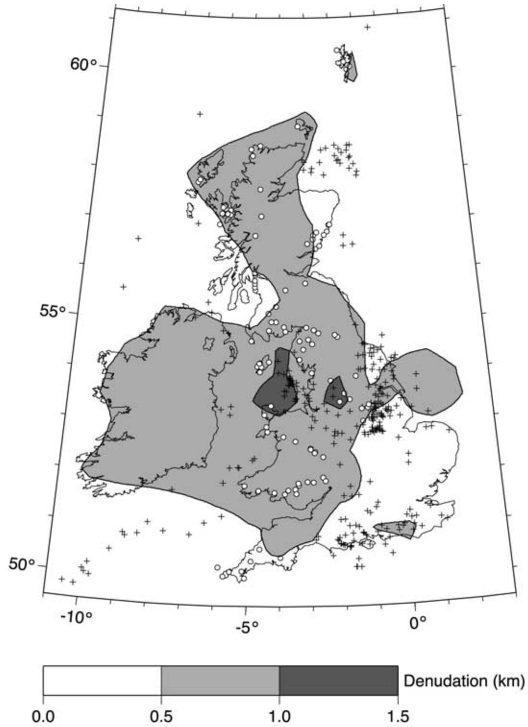

Fig. 2. Estimate of Cainozoic denudation of Britain and Ireland from modelling subsidence histories of well sections (crosses) and apatite fission track length distributions (circles). See [33] for further details.

When the present-day topography is corrected both for a global sea-level change of 100 m and for denudation, the modelled coastline lies close to the present one (Fig. 1b). The discrepancy between this modelled coastline and the latest Palaeocene coastline observed on seismic datasets is a measure of contemporary dynamic support. In the inverse problem, the dynamic support profile is estimated by minimising misfit between observed and predicted coastlines. Note that although reconstructed topography depends

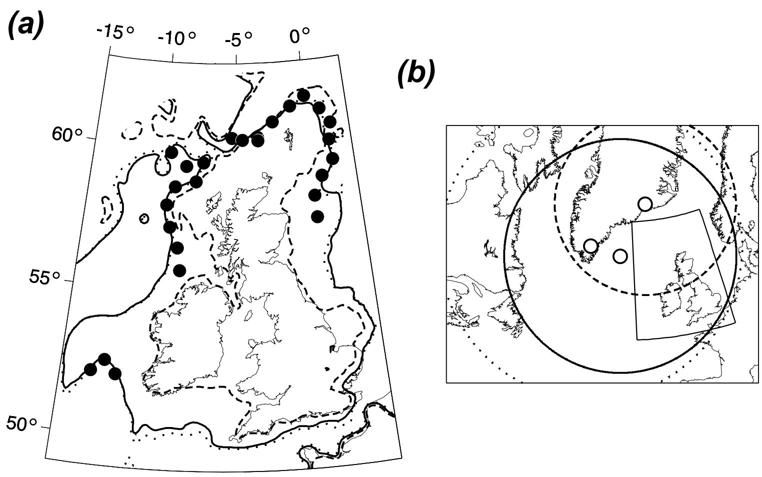

Fig. 3. Sensitivity of model coastline predictions to variations in the height, width and centre of the dynamically supported swell. (a) Observed coastlines (black spots) and three reconstructed coastlines (heavy lines). (b) Corresponding model swell planforms, drawn at the 30 m dynamic support contour, plotted on a latest Palaeocene continental reconstruction. Solid line: best-fitting swell (H = 2.0 km, D = 1500 km, (C_λ, C_ϕ) = (59°N, 18°W), μ = 40 m). Dotted line: alternative swell with very similar misfit, illustrating trade-off between H, D and C_ϕ (H = 2.7 km, D = 1900 km, (C_λ, C_ϕ) = (60°N, 25°W), μ = 42 m). Dashed line: swell with central position same as today, similar to prediction in [6] (H = 1.6 km, D = 1200 km, (C_λ, C_ϕ) = (65°N, 11°W), μ = 245 m). Open circles mark model plume centres.

strongly on the denudation correction, the predicted position of the coastline is independent of denudation (Fig. 2). This situation arises because negligible denudation can occur seaward of the coastline at maximum regression. Thus the denudation correction is significant in the prediction of Eocene onshore–offshore sediment transport pathways, but the dynamic support profile determined by inversion does not depend on the denudation correction.

In order to place reasonable bounds on the inverse calculation, we begin by assuming a radially symmetric pattern of dynamic support with a Gaussian profile, similar to the axisymmetric plume swell model proposed by White and McKenzie [6]. This assumption merely reflects an initial attempt to find the most parsimonious model that satisfies the data, and can be progressively relaxed as required. The results of the inverse calculation are shown in Fig. 1. We systematically co-varied the maximum amplitude H, diameter D (measured at height H/e) and central position (C_λ, C_ϕ) of the dynamic support model, and measured misfit μ between modelled and observed coastlines. The misfit function is defined as the mean model topography per unit length at the observed coastlines. We find it possible to match all the palaeo-coastlines surrounding Britain and Ireland using a single dynamic support model (Fig. 1b). A small, roughly circular minimum on all the misfit plots involving central latitude C_λ shows that this parameter is well constrained (Fig. 1d). However, parameters H, D and C_ϕ trade off so that a range of swells with larger radius and dynamic support situated further west, and vice versa, can explain the observed coastlines equally well (Figs. 1d and 3). This trade-off occurs because the swell is only constrained by data in its southeastern corner at present, and does not represent a problem with the inversion technique. The contour interval for the average data misfit in Fig. 1d is 100 m, showing that model coastlines can be dramatically affected by small changes in width, height and position of the swell (Fig. 3). The predicted North Sea coastline is most sensitive to changes in swell parameters, while the coastline in the Faroe–Shetland Basin provides relatively little constraint on the dynamic support model. In general, the most useful palaeo-coastline observations come from regions with relatively flat present-day topography.

2.2. Subsidence anomalies

The latest Palaeocene dynamic support profile obtained by inverse modelling of palaeo-coastlines can be compared with estimates obtained from anomalous subsidence histories of extensional sedimentary basins. Given the syn-rift subsidence history, post-rift subsidence can be calculated using the well-established lithospheric stretching model (Fig. 4). Post-rift marker horizons are then compared with the anticipated post-rift subsidence curve to reveal anomalous uplift and subsidence events. Dynamic support associated with the North Atlantic Igneous Province can be seen clearly in records of anomalous subsidence (Fig. 4a,b). Detailed biostratigraphic analysis of the corresponding regressive sedimentary succession shows that the bulk of the regional uplift occurred over 1–2 Myr or less during Late Palaeocene time [23]. Dynamic support peaked in latest Palaeocene times and decayed through the Eocene over a period of 10–20 Myr. Latest Palaeocene delta-top sediments of the Sele Formation (North Sea) [34], Flett Formation (Faroe–Shetland Basin) [35] and their lateral equivalents offshore Ireland [36] provide good constraints on the magnitude of peak dynamic support thanks to well-constrained depositional water depths. The same horizons constrain the position of latest Palaeocene coastlines used in the topographic reconstruction above, so the two different dynamic support estimates are constrained in the same spatial and temporal locations.

In general, dynamic support estimates from subsidence analysis provide a good match to the dynamic support model derived from topographic reconstruction, verifying both sets of calculations (Fig. 4c). In detail, the anomalous subsidence technique explicitly accounts for variations in

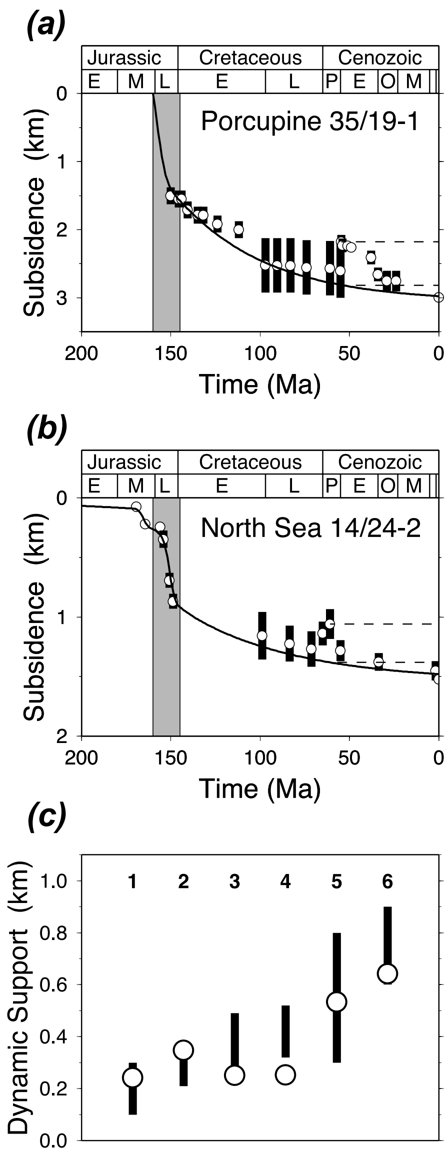

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Fig. 4. (a,b) Typical 1D subsidence plots to illustrate estimation of latest Palaeocene peak dynamic support from stratigraphy of Porcupine Basin and the Outer Moray Firth, North Sea Basin. Circles = backstripped stratigraphy, bars = uncertainty in depositional water depth. Theoretical subsidence curve fitted to syn-rift stratigraphy assuming lithospheric pure shear stretching; post-rift curve forward modelled assuming no further stretching (technique described fully in [17]). Grey shading = syn-rift period constrained by seismic reflection profiles. Dashed lines illustrate subsidence anomaly interpreted as a latest Palaeocene peak in dynamic support. (c) Comparison of latest Palaeocene dynamic support estimates from subsidence analysis with those from fitting coastlines. Black bars = estimates from subsidence analysis at numbered locations marked on Fig. 1b: 1 = Fastnet Basin [48]; 2 = South Viking Graben [16]; 3 = Witch Ground Graben [16]; 4 = North Viking Graben [16], 5 = Faroe–Shetland Basin [16]; 6 = Porcupine Basin [17]. Values normalised to Cainozoic global sea-level fall of 100 m. Circles = estimates from fitting coastlines at these locations, from Fig. 1c; errors discussed in the text.

sediment accumulation rate via backstripping, but these variations are a source of error in the topographic reconstruction technique. For example, sedimentary basins surrounding Britain and Ireland were starved of sediment in latest Cretaceous times, but sediment flux subsequently increased in response to regional uplift and to global climate change. Further analysis of subsidence records from Porcupine Basin suggests that the resultant sediment infill caused the depositional water depth to decrease by around 200 m through Cainozoic time. Thus our topographic reconstruction probably underestimates the dynamic support of Porcupine Basin by 200 m. Regions that have not accumulated Cainozoic sediment are not subject to this error source during topographic reconstruction. Notwithstanding this second-order problem, the main advantage of the topographic reconstruction technique is that it clearly shows regional palaeogeographies and drainage patterns that are likely correct to first order.

2.3. Extrapolation of topographic reconstruction

The effect of rifting needs to be considered when extrapolating the topographic reconstruction developed around Britain and Ireland into deep-water regions further west. Cretaceous rifting probably affected Rockall Trough and the Faroe–Shetland Channel and continental break-up between Greenland and Europe occurred in latest Palaeocene times [28]. Significant thermal subsidence following these events has occurred through Cainozoic time which is not accounted for in our reconstruction. Substantial topography at the base of the lithosphere during Palaeocene time was potentially infilled by the hot asthenosphere layer beneath that generated the regional uplift [29]. Variation in thickness of the hot layer would lead to a swell of irregular shape. However, a marked increase in viscosity associated with dehydration as mantle rises above the dry solidus is indicated by experimental results [37] and appears to be an important control on mantle flow beneath Iceland at present [38,39]. This effect should inhibit ponding of hot asthenosphere in lithospheric thin spots over the short timescale for swell inflation suggested by the stratigraphic data ($\sim$1 Myr). Unusually hot mantle was likely injected into an asthenosphere channel confined below the depth of the dry solidus, which is roughly constant.

3. Combining stratigraphic and igneous constraints

We previously argued that widespread regional uplift was caused by a thermal anomaly within the upper mantle. Both uplift and generation of the North Atlantic Igneous Province were likely associated with this same thermal anomaly. The distribution of igneous rocks can therefore be used to map the planform of the thermal anomaly.

White and McKenzie [6] presented a classic reconstruction of the North Atlantic Igneous Province at the time of break-up between Greenland and Europe. They suggested that a circle circumscribing the large igneous province provides a good estimate of the planform of the starting Iceland Plume head (Fig. 5b). However, this circular planform is incompatible with the stratigraphy of Britain and Ireland (Fig. 3). Conversely, the circular planform that best fits the stratigraphy of Britain and Ireland is located too far south to explain igneous activity between Greenland and Norway (Fig. 5b). Can both stratigraphic and igneous constraints on the shape and size of the dynamically supported swell be satisfied without resorting to a complex, irregular planform? A complete answer to this question requires analysis of a more geographically widespread stratigraphic data set. Nevertheless, some interesting conclusions can be drawn by comparing the preliminary stratigraphic constraint determined here with regional igneous constraints. We first review the spatial extent of the North Atlantic Igneous Province.

Location and volume of igneous activity is controlled by lithospheric as well as by mantle processes. The most prominent parts of the North Atlantic Igneous Province are the volcanic passive margins that form a linear belt just offshore eastern Greenland on a latest Palaeocene continental reconstruction (Fig. 5). The limits of this igneous belt have been interpreted to constrain the north-

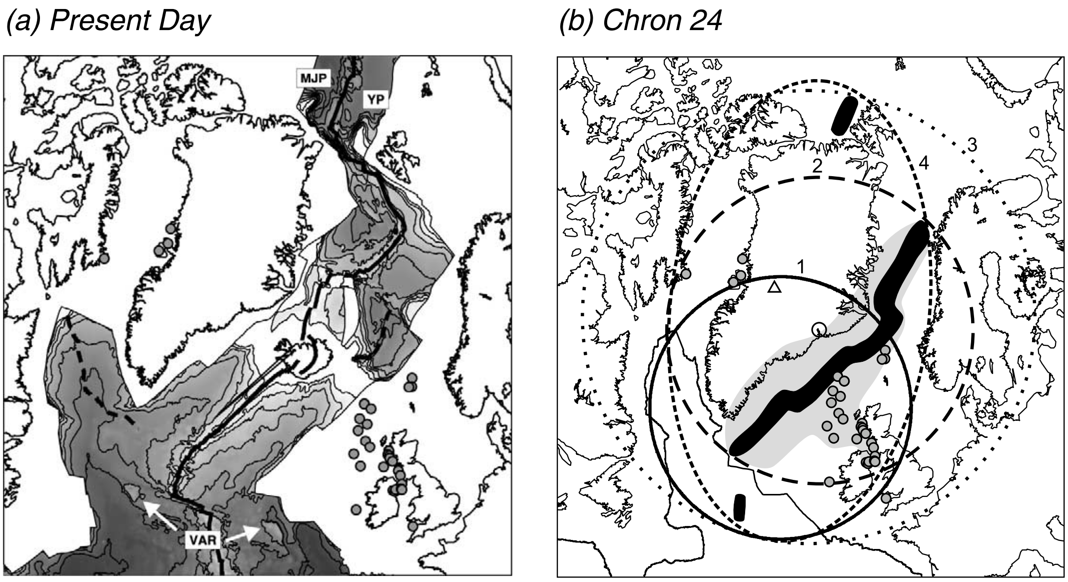

Fig. 5. (a) Present depth of the ocean (from ETOPO5) illustrating aseismic oceanic ridges and plateaux thought generated above a mantle thermal anomaly: MJR = Morris Jesup Rise, VAR = Vogt and Avery's [40] Ridge 'R' and conjugate, YP = Yermak Plateau. (b) Continental reconstruction at chron 24 time showing North Atlantic Igneous Province: incipient volcanic passive margins and oceanic ridges (black), flood basalts (grey), and central igneous complexes (grey circles). Numbered planforms: 1 = circular planform from inversion of British and Irish coastline data (same as Fig. 1c), 2 = White and McKenzie's [6] circular planform from plume-related igneous activity, 3 = circular planform encompassing all dynamically supported regions and plume-related igneous activity, 4 = elliptic planform that satisfies both stratigraphic constraints from around Britain and Ireland and igneous activity across the entire North Atlantic. Triangle = plume centre predicted from hotspot reference frame model of Lawver and Müller [49], circle = present-day plume centre in chron 24 reference frame. Projections are azimuthal equidistant, centre 64.5°N, 11°W in fixed European reference frame.

ern and southern limits of influence of the thermal anomaly [6]. But oceanic plateaux and ridges to the north and south of the belt suggest that the thermal anomaly may have been larger in planform. Southwest of Greenland in the Labrador Sea, seafloor spreading began in Late Cretaceous time. If the edge of the thermal anomaly extended southwards beneath this established spreading ridge, conjugate bathymetric ridges of over-thickened oceanic crust should have formed either side of the spreading axis. In fact, Vogt and Avery [40] recognised conjugate bathymetric ridges lying between chrons 24 and 23 (earliest Eocene) in the oceanic abyssal plains between Rockall Bank and Newfoundland, and suggested that they were generated by coeval excess asthenospheric temperature. The southern edges of these bathymetric ridges are not related to tectonic boundaries and are more likely to represent the true southern edge of the thermal anomaly (Fig. 5a). The Yermak Plateau and the Morris Jesup Rise lie offshore Svalbard and NE Greenland respectively (Fig. 5a). These features formed during Eocene time (chron 24–13) by seafloor spreading at the Gakkel Ridge in the Arctic Ocean. Although seismic data are scarce because of problems with icebergs, the anomalous elevation of both structures suggests unusually thick oceanic crust generated above hot asthenosphere [41,42]. The northern edges of these plateaux appear unrelated to tectonic boundaries and they may represent the edge of the regional thermal anomaly. If this interpretation is correct, then the widths of the plateaux suggest that northern edge of the thermal anomaly decayed through the whole of Eocene time, whereas the southern edge decayed more rapidly during earliest Eocene time.

When the revised North Atlantic Igneous Province is plotted on a contemporary continental reconstruction, we see that it is not possible to find a swell with a circular planform and Gaussian profile that satisfies both continental stratigraphic constraints from Britain and Ireland and igneous activity across the entire North Atlantic Ocean (Fig. 5b). A circular planform that encompasses all dynamically supported regions and the entire igneous province would over-predict dynamic support in the region of Britain and Ireland. An alternative simple dynamic support model with an elliptic planform that satisfies regional igneous constraints is also illustrated on Fig. 5b. This elliptic swell adequately satisfies stratigraphic constraints around Britain and Ireland. However, it is perhaps more likely that the initiating plume swell had a more irregular shape. Stratigraphic data from other North Atlantic margins is required to make progress.

4. Origin of the North Atlantic Igneous Province

The observed regional uplift implies unusually low densities in the upper mantle, which in turn implies a thermal anomaly. Recycled eclogite from ancient oceanic crust may contribute to melting anomalies [9], but a thermal anomaly is still required to produce uplift since eclogite is more dense than a standard peridotite source at the same temperature and pressure in the upper mantle. The requirement for a regional thermal anomaly also suggests that any small-scale convection generated by continental break-up is of secondary importance in development of a regional melting anomaly. This inference is supported by the results of numerical experiments that model both effects [43]. A complementary observation is that dynamic support reaches about 1000 km inboard of the NW European volcanic margin, whereas convection related to break-up is expected to be localised within a few hundred kilometres of plate boundaries [13,43]. The suggestion that the architecture of Europe–Greenland volcanic margins reflects variation in rate of upwelling of normal temperature asthenosphere beneath the Mid-Atlantic Ridge [8] is also problematic in view of the evidence for uplift.

The bulk of the North Atlantic Igneous Province, in particular the Europe–Greenland volcanic margins, was emplaced in 1–2 Myr or less [21]. The widespread regional uplift we map here developed over the same time interval [23]. This time interval is too small to admit either a thermal anomaly that develops by incubation [10,11] or a pre-existing plume head, so the thermal anomaly must have been emplaced by convection as uplift occurred. Coeval development of volcanic margins and regional uplift extending about 1000 km inboard suggests rapid lateral injection of a horizontal sheet of hot mantle beneath the lithosphere. Scenarios involving hot vertical sheets [2,24] cannot satisfactorily explain the widespread uplift. Beneath the break-up zone, the hot, horizontal sheet welled up principally in response to plate spreading, decompressed and generated a melting anomaly [6]. The anomalous melt thickness decreased as the hot layer incorporated into new oceanic plates, explaining observed thinning of oceanic crust outboard of both margins [24,44]. Beneath the continents, the sheet was probably initially confined below the dry solidus [37,38] and negligible upwelling occurred, explaining why uplift was more widespread than volcanism. An early model of a starting plume head that detaches from a thermal boundary layer, rises as a roughly spherical blob and gradually flattens as it impinges underneath the lithosphere predicted a relatively slow time interval for uplift of $\sim$10 Myr [5]. More recent starting plume models incorporating a general increase in viscosity with depth both predict more rapid surface uplift (around 5 Myr) and tend toward horizontal emplacement of the plume head [45]. Larsen et al. [8] show that a non-Newtonian rheology promotes very rapid ($<$1 Myr) horizontal emplacement of a thin hot layer beneath the lithosphere. However, it seems likely that a Newtonian convection model specifically incorporating a low-viscosity asthenosphere channel would share these characteristics, so the observations of uplift do not yet constrain mantle rheology. Unfortunately, it is also unlikely that planforms of dynamically supported swells can be used to determine the depth of the underlying convective system [1].

Tomographic, dynamic support, melt thickness and geochemical observations suggest that a hot, upwelling mantle plume exists beneath Iceland today [38,39,46]. The spatial co-location of Iceland and the latest Palaeocene uplift and magmatism described herein suggests that these events reflect the starting phase of the present-day Iceland Plume. However, there remains much work to be done to understand the time dependence of the Icelandic convective system. In particular, it is unclear whether the tectonically complex Greenland–Iceland–Faroes Ridge can be interpreted as a plume track that connects the large igneous province with the present plume centre. Another important question concerns how the Palaeocene, spatially confined phase of North Atlantic Igneous Province activity [21,22] (Fig. 5b) relates to the latest Palaeocene, spatially extensive phase mapped herein.

5. Conclusions

We have shown how a continental stratigraphic record can be used to constrain past spatial patterns of dynamic support. Palaeo-coastline and anomalous subsidence data from Britain and Ireland are compatible with a simple dynamically supported swell characterised by a circular planform and a Gaussian profile. However, circular planforms which satisfy the stratigraphic constraints do not agree with the spatial distribution of the North Atlantic Igneous Province, and vice versa. Together, stratigraphic and igneous data suggest an elliptic or likely more irregular dynamically supported swell $1\text{–}2\times10^3$ km in diameter during its maximum inflation in latest Palaeocene time. This inference can be tested in future by including stratigraphic data from offshore Norway, Greenland and Canada in the palaeo-topographic inversion. Alternatively, if a geographically widespread database of dynamic support estimates can be compiled from stratigraphic and igneous records, the swell can be directly contoured.

We also re-evaluated models advanced to explain the North Atlantic Igneous Province using both the continental stratigraphic and revised igneous records. The large spatial extent of latest Palaeocene regional uplift must have been supported by a thermal anomaly in the upper mantle. The short time interval for inflation of the plume swell and for emplacement of the bulk of anomalous magmatism is best explained by the rapid horizontal emplacement by convection of an anomalously hot sheet of mantle beneath the lithosphere.

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