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Gravity anomalies and magmatism along the western continental margin of the British Isles

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Abstract: Seismic data show that the western margin of the British Isles in the region of the Hatton Bank comprises a thin wedge of sediments, an extrusive volcanic sequence, stretched continental crust, and a thick high-velocity lower-crustal body. The sediments represent a load on the surface of the stretched crust, which should subside under their weight. The extrusive volcanic rocks defined by a seaward-dipping reflector sequence, were emplaced in the crust during the later stages of rifting. The lower-crustal body has been interpreted as underplated material which re-thickens the crust and causes uplift. Backstripping techniques have been used to evaluate the contribution of sediment loading and underplating to the observed crust and mantle structure and isolate the initial rift configuration of the margin. The contribution of these processes depends, however, on the elastic thickness, T_e, of stretched lithosphere. We have constrained T_e by computing the combined gravity anomaly of rifting, sedimentation and underplating and comparing it to the observed free-air gravity anomaly. Underplating gives rise to a distinct pattern of gravity anomaly 'highs' and flanking 'lows': the highs reflect the relatively dense uplifted crust and the lows less dense underplated material. The best fit between observed and calculated anomalies is for a model in which the T_e of stretched lithosphere is low (<5 km) and the width of the initial rift is narrow (<75 km). Since T_e is low, isostatic anomalies at the margin would be expected to be of small-amplitude and this is indeed the case at the Hatton Bank margin. Sensitivity studies show that large-amplitude isostatic anomalies would be expected, however, if T_e is high—with negative anomalies of up to 30 mGal over the underplated region and positive anomalies of up to 10 mGal in flanking regions. Such anomaly patterns are seen west of Ireland and Scotland suggesting that underplating may be a widespread feature of the British Isles margin.

Keywords: gravity, anomalies, magmatism, continental margin, lithosphere, flexure.

It has been recognized for some time that passive continental margins form by extension of the crust and lithosphere at the time of rifting. A number of models have been proposed to explain the response of the lithosphere to extension. They include kinematic models in which the lithosphere deforms either by 'pure shear' (McKenzie 1978) or 'simple shear' (e.g. Lister *et al*. 1986) and dynamic models in which the lithosphere deforms in response to a prescribed temperature, rheology and velocity structure (e.g. Braun & Beamount 1987; Dunbar & Sawyer 1989; Bassi *et al*. 1993). Both types of models make certain predictions concerning the symmetry, the role of detachment surfaces, and the amount of crustal thinning at margins which should be resolvable in seismic reflection and refraction data.

Despite major technological advances in acquisition systems during the past few years, it has proved difficult to use seismic data to distinguish between the different types of extensional model. One reason is that rift-type margins are influenced by a wide range of geological processes (e.g. Fig. 1) which obscure the crustal and upper mantle structure produced by rifting. Included in these processes are sedimentation, erosion and magmatism. There is evidence that margins are dominated by one process more than another. The Gulf of Mexico margin (Antoine *et al*. 1974), for example, has thick sediments, normal lower-crustal velocities (i.e. P wave velocity <6.8 km s^{-1}) and an absence of magmatism while the Vøring and Møre margins offshore Norway (Eldholm *et al*. 1995) have thin sediments, high lower-crustal velocities (>7.3 km s^{-1}), and evidence of magmatism in the form of seaward-dipping reflector sequences and underplating. Other margins such as the East Coast, USA (Kelemen & Holbrook 1995) appear to be associated with thick sediments *and* large amounts of magmatic material.

One approach to isolate the initial configuration of a margin is by backstripping (Watts & Ryan 1976). By comparing the predictions of extensional models to backstripped stratigraphic data, for example, it is possible to constrain the amount of extension at a margin. When applied to data at a well, backstripping only provides information on the amount of extension at a single point. Flexural backstripping of seismic reflection profile data (e.g. Watts & Ryan 1976; Sawyer 1985; Bessis 1986) has the potential to constrain variations in the amount of extension *across* a margin. The main problem is that the technique requires knowledge of the spatial variations in the long-term (i.e. $>10^6$ a) thermal and mechanical properties of the lithosphere. However, as shown by Watts (1988) it is possible to constrain these properties when the flexural backstripping technique is used in combination with gravity modelling.

Most previous backstripping and gravity modelling studies have been carried out at margins such as offshore East Coast, USA (Watts 1988) and Canada (Keen & Barrett 1981) where there are thick sediments. An outstanding problem is how to apply the technique to margins, such as offshore the British Isles, Norway and Greenland, which are known from seismic data to be dominated by magmatism. The purpose of this paper is to use a new compilation of gravity anomaly data along the western margin of the British Isles, together with backstripping and gravity modelling, to determine the gravity

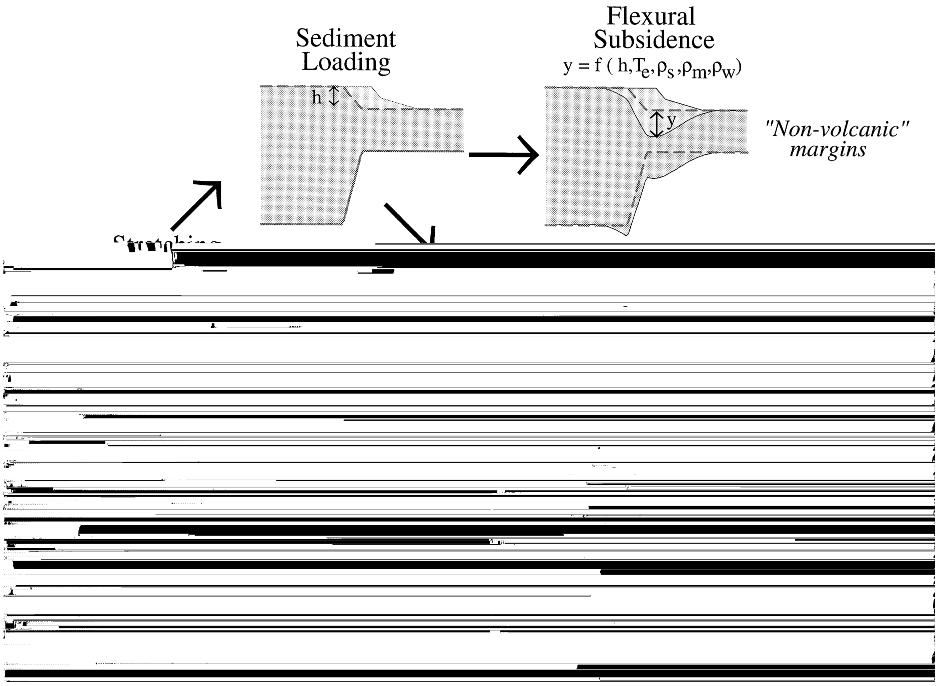

Fig. 1. Simple models for the development of passive continental margins. The models assume that margins are the consequence of stretching, sediment loading and underplating. 'Volcanic margins' are considered as those cases where stretching and underplating dominate; 'Non-volcanic' margins where stretching and sediment loading dominate. The stretched crust is assumed to be in Airy isostatic equilibrium. Sediment loading and underplating are assumed to modify the crust by flexure. The models are based on an initial crustal thickness of 30 km, a stretching factor of 3, a density of water, sediment, underplated material and mantle of 1030, 2400, 2900 and 3330 kg m^{-3} respectively, and a constant elastic thickness of the lithosphere, T_e, of 25 km.

anomalies that might be expected from magmatism. Our aim is to better understand (a) the role of magmatism in contributing to the gravity field, (b) the strength of extended continental lithosphere and, (c) the manner in which the continental lithosphere responds to extension.

Gravity, seismic and well data

The gravity anomaly data used in this paper is based on a compilation carried out by the University of Leeds company GETECH as part of its West–East Europe Gravity Project (WEEGP). The compilation includes (Green & Fairhead 1994) all available 'point' gravity anomaly data for Europe, including data acquired in onshore and offshore regions by the national surveys of the UK and Ireland. Data in offshore regions have been supplemented by commercial and academic data sources. In regions of poor surface ship coverage (e.g. Rockall Plateau) the marine gravity data set has been supplemented by gravity data recovered from sea surface height data acquired during the GEOSAT, ERS-1 and Topex/Poseidon satellite altimeter missions. Both land and marine data sets have been corrected for rogue points, inter-survey compatibility, and base station tie-in errors.

Figure 2 shows a raster image of a 5 # 5 minute gravity anomaly grid of the western continental margin of the British Isles. The free-air image (Fig. 2a) is dominated by an 'edge effect' anomaly that comprises a high over the outer shelf and a low over the flanking slope and rise. Offshore northern Scotland and southwest England, the edge effect appears to truncate older Caledonian and Hercynian structures. The edge effect is generally interpreted (Worzel 1968) as the consequence of the juxtaposition of thin oceanic with thick continental crust. At many margins (e.g. Nova Scotia; Keen & Barrett 1981), the form of the thickening is in general accord with the predictions of the Airy model. Indeed, when the gravity effect of an Airy model is subtracted from the free-air gravity anomaly, the resulting isostatic anomalies are generally small (Talwani & Eldholm 1973). Unusually, the British Isles margin shows quite large-amplitude isostatic anomaly 'highs' and 'lows' (Fig. 2b). The existence of these anomalies imply deviations from the Airy model (due either to changes in crustal density or crustal thickness, or some combination of these factors) which should be apparent in seismic data.

Seismic reflection and refraction studies of the British Isles margin are limited to two main 'transects': one in the Hatton Bank (Fowler *et al.* 1989) and the other in the Goban Spur (Horsefield *et al.* 1993) regions. They suggest (White 1992) that the structure of the margin can be characterized by two end member types: those that are associated with magmatism and those that are not. The Hatton Bank margin is an example of a volcanic margin with thick sequences of extrusive basalts and a thick, high-velocity, underlying lower-crustal body. Goban Spur, however, is a non-volcanic margin with tilted fault blocks and normal lower-crustal velocities.

The Hatton Bank transect (Fowler *et al.* 1989) crosses a small-amplitude positive isostatic anomaly which trends subparallel to the margin. The peak of the high is located between Expanding Spread Profile (ESP) mid-points A and G which define the high velocity underplated material. That a link might exist with the isostatic anomaly is not entirely unexpected since underplating modifies the density structure of the crust in a way that is not accounted for by the Airy model.

There is evidence that magmatism has modified other data types along the British Isles margin. Deep Sea Drilling Project (DSDP) data (Clift *et al.* 1995), for example, show that at Site 550 on the Goban Spur transect the tectonic subsidence is normal for the thermal age of the underlying oceanic crust. In contrast, at Site 552 SW of Hatton Bank the subsidence is up to 1.8 km *shallower* than expected. These differences have been interpreted as due to proximity of the Hatton Bank to the Iceland plume which re-thickened and uplifted the pre-existing crust. Other evidence has come from backstripping well data in the Porcupine (Tate *et al.* 1993) and the Faeroes–Shetland Channel (Turner & Scrutton 1993) basins.

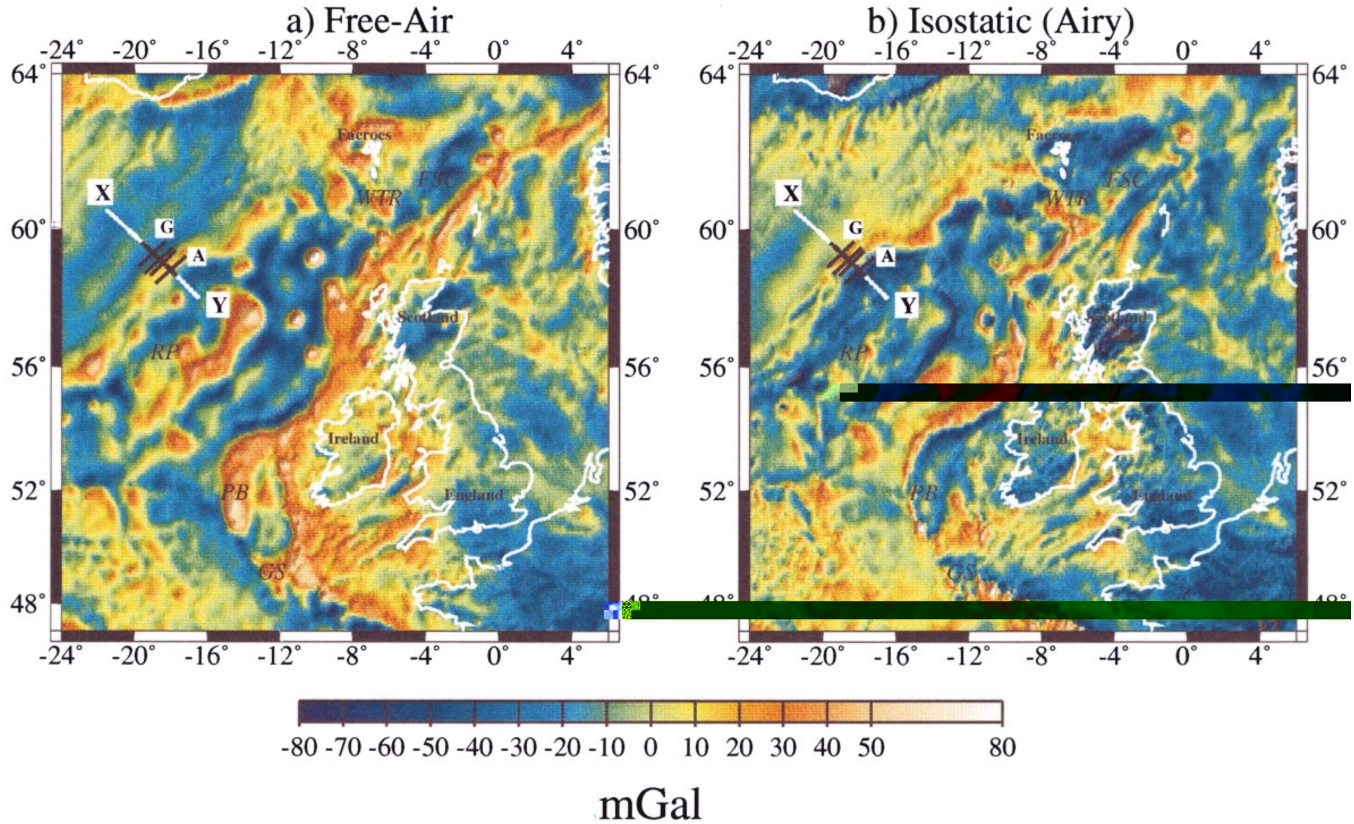

Fig. 2. Gravity anomalies at the continental margin west of the British Isles. The anomalies are based on a 5×5 minute grid of all available university, government and industry 'point' gravity data. The gridded gravity anomalies have had a long-wavelength gravity field (the OSU 91a field to degree and order 16; Rapp *et al.* 1991) removed from them. The resulting field resolves features of the gravity anomaly field with wavelength $\lambda < 2500$ km. Solid black and white lines indicate the location of the seismic transect of the Hatton Bank margin along which the gravity anomaly study has been carried out. PB, Porcupine Bank; RP, Rockall Plateau; GS, Goban Spur; FSC, Faeroes–Shetland Channel; WTR, Wyville–Thompson Ridge. (**a**) Free-air gravity anomaly. (**b**) Isostatic anomaly based on an Airy model with densities of the water, crust and mantle of 1030, 2850 and 3200 kg m^{-3} respectively and a mean thickness of 4.5 km deep oceanic crust of 7.5 km.

Data analysis

We analysed the data on the Hatton Bank transect in two steps. First, backstripping techniques were used to determine the relative contributions of sediment loading and underplating to the margin and isolate the structure that resulted from rifting. Second, the gravity anomaly due to the rift structure was calculated, combined with the anomaly due to sedimentation and underplating, and compared to the observed anomaly.

As pointed out earlier, magmatism at Hatton Bank is expressed as extrusive basalts in the uppermost crust and underplated material at the base of the crust (Fowler *et al.* 1989). The influence of the extrusive basalts on the loading history of the margin is not clear. According to White (1992), they represent sub-aerial lavas which have subsided because of loading and thermal contraction of the underlying stretched continental crust. Eldholm *et al.* (1995), however, have pointed out that the seaward-dipping reflectors that define the basalts on seismic reflection profiles terminate abruptly on their seaward side and may be bounded by a listric fault (see also Barton & White 1995). If this is the case, then the sequences are accommodated *within* the rift crust (as is the case for syn-rift sediments at non-volcanic margins) and for this reason we have not taken their loading effects in the initial models.

McKenzie (1984) has shown that if a thickness of underplated material, X, is added to the base of the crust then the corresponding uplift, u, is given by $X(\rho_a - \rho_x)/(\rho_a - \rho_w)$ where ρ_a, ρ_x, and ρ_w are the mean densities of the asthenosphere, underplated material and water respectively. At the Hatton Bank margin (Fowler *et al.* 1989) seismic data suggest up to 15 km of underplated material has been added which implies an uplift of 2.1 km, assuming densities of 3200, 2900 and 1030 kg m^{-3} for the asthenosphere, underplated material and water respectively.

Figure 3 shows a restoration of the Hatton Bank margin *at the time of rifting* which takes into account the modifying effects of sediment loading and underplating. The restoration was carried out in three steps. First, the sediments at the margin were backstripped in order to obtain the depth that basement would have been in the absence of sediment loading. Second, the uplift corresponding to the amount of underplated material determined seismically was computed and subtracted from the backstripped sediment depth. The result is the depth that basement would have been in the absence of sediment loading and underplating. The final step, was to compute the 'backstrip' Moho from the backstripped basement assuming Airy isostasy.

The restoration in Fig. 3 allows the gravity effect of the individual processes believed responsible for margin formation

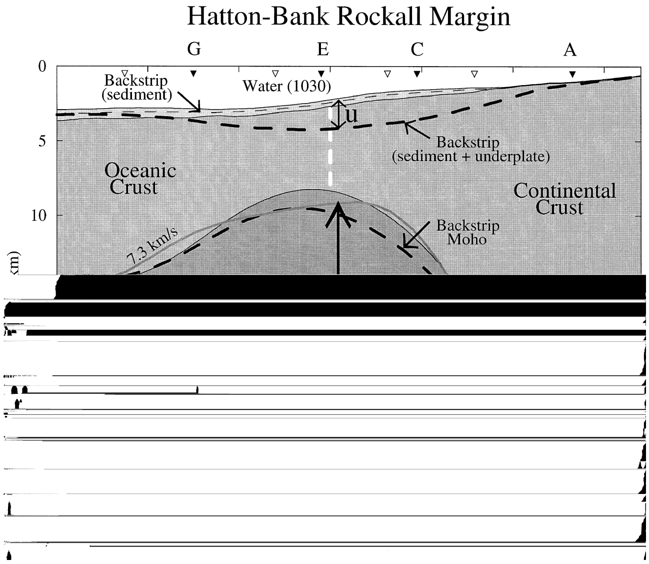

Fig. 3. Simple model for the crustal structure of the Hatton Bank margin. The grey solid lines show the observed depth to the base and top of the underplated material based on seismic refraction data (Fowler, *et al.* 1989). The heavy dashed lines show the restored structure of the crust. The top of the restored crust was obtained by first backstripping the sediments and then calculating the uplift, *u*, due to underplating, *X*. The bottom of the restored crust was computed assuming Airy isostasy. The densities in brackets show the densities assumed in the backstripping.

to be computed and compared with observed anomalies. The rift anomaly has the form (Fig. 4a) of an edge effect with a high at the transition between unstretched and stretched continental crust and a low over the extended region. The individual contributions to the gravity anomaly of underplating (Fig. 4b) and sediment loading (Fig. 4c) were computed from the differences between the rift crustal structure and the present-day topography. The underplating anomaly comprises a gravity 'high' which is flanked by 'lows': the high reflect the excess mass associated with the displacement of water by the higher density uplifted crustal rocks while the lows reflect the lower density of the underplated material compared to that of the surrounding mantle. There is, additionally, a small high caused by uplift of the Moho by the underplated material and the displacement of dense underplated material into the crust. The sedimentation anomaly comprises a small-amplitude high since sediments represent a mass excess which is flanked by a small amplitude low.

The 'sum' anomaly (Fig. 4d), obtained by adding the underplating and sedimentation anomalies to the rift anomaly, shows a broad high with two peaks: one centred over the original boundary between stretched and unstretched crust and the other over the region of greatest underplating and uplift. The high is flanked on the seaward side by a broad negative anomaly. Underplating is the principal contributor to the sum anomaly, widening the original rifting anomaly and reducing its overall amplitude.

We compare in Fig. 4 the sum anomaly to the available surface ship free-air gravity anomaly data in the region of the seismic transect. The observed data are based on two previous cruises to the region on RV Vema and RRS Shackleton both of which crossed the transect. These data agree to better than 5 mGal with some unpublished gravity data that were acquired by the University of Durham (Prescott 1988) along the original transect. The figure shows a reasonably good agreement between the observed and calculated anomalies, especially when it is considered that we have used such a simple density model for the crustal and upper mantle structure.

The calculation of the sum anomaly in Fig. 4d assumes that the stretched crust responds *locally* to the low density material that has been added to its base by underplating. The model therefore ignores any contribution that the long-term (>10^6 a) flexural strength of the crust may make in limiting this response. To investigate this effect we calculated the wavenumber parameter, ö(k), that when multiplied by the local isostatic response gives the flexural response to underplating. The parameter is given by:

$$\Phi(k)=\left[1+\frac{Dk^4}{(\rho_m-\rho_w)g}\right]^{-1}$$

where ñ$_m$ is the density of the mantle, g is average gravity, D is the flexural rigidity and k is wavenumber (k=2ð/ë). The sum anomaly for an elastic thickness, T_e, of 10 km (equivalent to a flexural rigidity of 8.9 # 10^{21} N m) differs from the anomaly based on local compensation in two main ways: the high decreases in width and the low increases in amplitude. The differences arise because as T_e increases, the uplift is reduced (the stretched crust appears stronger to the underplated material) and hence, more of the features of the rifting anomaly appear in the sum anomaly. The sum anomaly is more negative over the underplated region and more positive in flanking regions than the observed anomaly, which suggests that the flexural strength of stretched crust is probably not an important factor at the Hatton Bank margin.

Discussion

This attempt to restore crustal and mantle structure at the Hatton Bank margin has a number of implications for the development of the continental margin basins. These include the compensation during rifting, the strength of extended lithosphere, the extent of magmatic material along the western margin of the British Isles, and the mode of continental extension.

Local compensation during rifting

We assumed in calculating the rifting gravity anomaly (Fig. 4) that the depth to the base of the rifted crust can be computed from the backstripped tectonic subsidence and uplift profile using an Airy model of isostasy. While an Airy assumption may be quite an acceptable one, it does impose constraints on the mode of extension at a margin. For example, in the case of the density structure assumed in Fig. 4, it limits the amount of crustal thinning to about 5 times the depth of the overlying water-filled basin.

According to Braun & Beamount (1987) and Kooi & Cloetingh (1992), the mode of extension at a margin is a rheological problem which is determined by the 'depth of necking'. The depth of necking, in their models, corresponds to a depth in the crust that does not move vertically during extension. Shallow necking depths mean that compared to the predictions of the Airy model of isostasy, there will be a shallow basin and large amounts of crustal thinning whereas for deep necking there will be a deep basin and small amounts of thinning.

Unfortunately, we know of no seismic, or other evidence, for the depth of necking along the western margin of the British Isles. Our approach therefore was to see the extent to which the observed gravity and seismic data could be explained by assuming an Airy model for the rifting anomaly. The fact that we can explain the observed gravity anomaly does not, by itself prove that an Airy model is an acceptable one to describe rifting. However, when taken with the results from other margins (e.g. East Coast, USA, Valencia Trough) which are also based on an Airy model, it makes a persuasive argument.

Strength of extended lithosphere

We have shown in computing the uplift due to underplating that the best fit to the observed gravity anomalies is for a model in which the stretched lithosphere is of negligible strength. There is currently much controversy, however, concerning the strength of extended continental crust. Some workers believe that rifted lithosphere is relatively strong (Weissel & Karner 1989) while others (Fowler & McKenzie 1989; Watts 1988) consider it weak.

There is evidence in Africa (Ebinger *et al.* 1989) that continental rift systems are associated with relatively low values of the elastic thickness compared to those of surrounding cratonic regions. Hartley *et al.* (1996) argued that a negative correlation exists between elastic thickness and present-day surface heat flow over Africa and suggested that crustal composition (i.e. the distribution of the heat producing elements) is a major factor in controlling continental T_e. However, other factors such as faulting and magmatic processes are also likely to contribute (Ebinger *et al.* 1989) to the reduction in elastic thickness.

Continental margins appear to be highly segmented as regards their long-term strength. Around Africa, for example, Watts & Marr (1995) identified a number of strong zones which abut weak ones. They attributed the strong zones to offshore extensions of the rigid cratonic areas of Africa. The weak zones, it was found, had in common that they were regions of prolonged hotspot activity. This suggests that magmatism may be a contributing factor at margins that reduces the strength of lithosphere: a result that is in general accord with the results of this study.

An outstanding problem is why (e.g. Watts 1988) the stretched lithosphere at some margins (e.g. East Coast, USA) has remained weak for long periods of time following rifting. Oceanic lithosphere, for example, increases its strength as it cools following a heating event. The occurrence of weak zones at old margins suggests that either the strength of stretched crust recovers on much longer (>200 Ma) time-scales than the oceanic lithosphere or, that during rifting the strong uppermost part of the crust is somehow de-coupled from any support that it might otherwise have received from the cooling and strengthening of the underlying mantle.

Extent of underplating material along the UK margin

The sum anomaly that best fit the observations in Fig. 4 is based on a local model of compensation. Therefore, isostatic anomalies based on an Airy model would be expected to be small. This indeed seems to be the case along the Hatton Bank transect (Fig. 2b) which shows small-amplitude isostatic anomalies which are generally positive to the northeast and generally negative to the southwest of the transect.

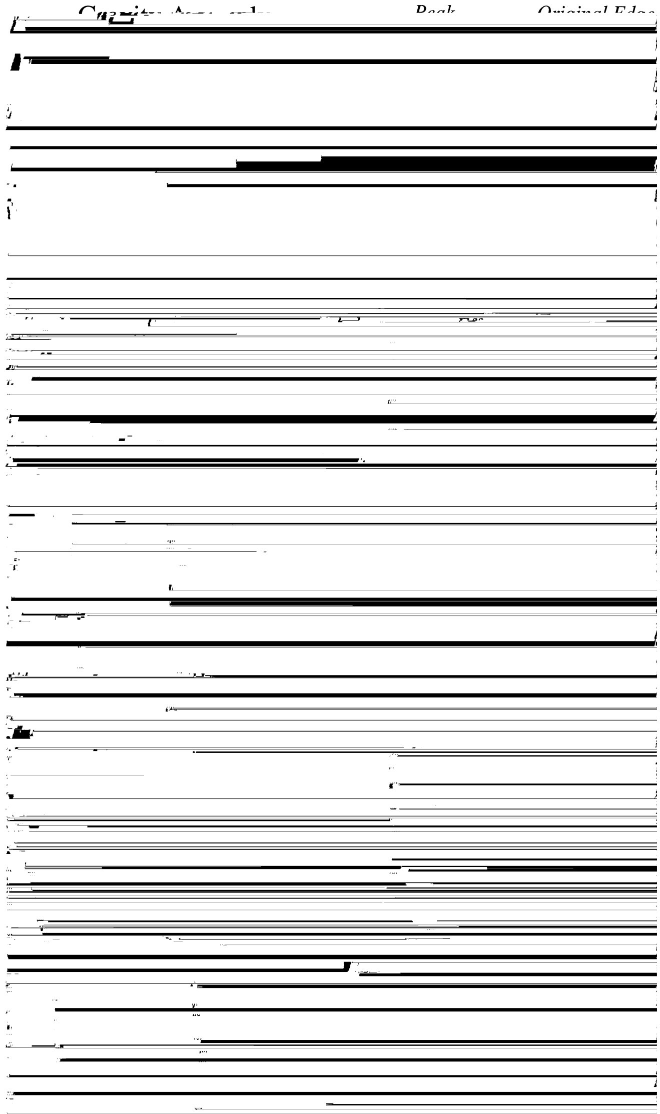

Fig. 4. Comparison of calculated and observed gravity anomalies along the Hatton Bank transect. The calculated anomalies are based on the crustal structure derived from backstripping. (**a**) Rifting anomaly, (**b**) underplating anomaly, (**c**) sedimentation anomaly and (**d**) sum anomaly. The solid lines have been computed for $T_e = 0$ km and densities of the underplated material of 2900, 2950 and 3000 kg m^{-3}. The thickness of the line shows the effect of using different densities for the underplated material. The dashed line shows the computed anomaly based on $T_e = 10$ km and a density of 2900 kg m^{-3} for the underplated material. The observed anomalies are based on available shipboard data in the region. The crustal model is based on Fig. 3. The depth to the 7.3 and 8.0 km s^{-1} velocity contours, which defined the underplated material seismically, are shown as white dashed and black dashed lines respectively.

Also shown in Fig. 4 is the sum anomaly that would be expected if underplating is limited by the flexural strength of the crust. In this case quite large-amplitude isostatic anomalies would be expected. For example, $T_e = 10$ km would be associated with isostatic anomalies with amplitudes of up to -30 mGal over the underplated region and up to +10 mGal in flanking regions.

Some segments of the western margin of the British Isles are characterised by large amplitude positive and negative isostatic anomalies (Fig. 2b) which suggests that underplating is not limited to the Hatton Bank margin. The positive anomalies form an 'outer' belt which extends along the margin northwest of Hatton Bank, north of Lousy, Bill Bailey and Faeroe Banks, to north of the Faeroes and an 'inner' one which extends west of the Porcupine Bank, west of Scotland to the Faeroes–Shetland channel. The two belts appear to connect along the trend of the volcanic Wyville–Thompson Ridge. The negative belts characterize the eastern margin of Rockall Plateau and the margin northwest of Ireland. The existence of a negative isostatic anomaly low that is flanked by two positives such as seen on the slope and rise west of Scotland and Ireland (Fig. 2b) is particularily diagnostic, and we speculate that this is a region of relatively strong stretched continental crust that is underlain by underplated material.

Modes of extension

We have used crustal restoration techniques to deduce the pattern of thinning of the continental crust across a rifted margin. As a number of workers have pointed out (e.g. Bassi *et al.* 1993; Buck 1991), the pattern of thinning of the continental crust at margins is one of the most useful constraints that we have on dynamical models of rifting.

The pattern of crustal thinning deduced at the Hatton Bank margin is compared to the results of backstripping, gravity modelling and crustal restoration studies at other margins in Fig. 5. These other margins include South Africa, Nova Scotia, Gabon, East Coast, USA, Brazil (Campos basin), Goban Spur, Carolina and the Valencia Trough (Western Mediterranean). The Valencia Trough and East Coast, USA curves are based on gravity modelling, flexural backstripping and crustal balancing studies (Watts 1988; Watts & Torné 1992). The South Africa curve is based on flexural back-stripping (Young 1992, fig. 5.15). of seismic reflection profile data. The Carolina, Brazil, Nova Scotia and Goban Spur curves were obtained by flexural backstripping previously published seismic reflection profile data (Beaumont *et al.* 1982; Hutchinson *et al.* 1983; Mohriak *et al.* 1990; Horsefield 1991).

Figure 5 shows that rifted margins can be divided into two main groups: wide (>250 km) and narrow (<75 km) rifts. Narrow margins are represented by Hatton Bank, Valencia Trough, Carolina and Goban Spur; wide margins by the East Coast, USA, Gabon, South Africa and Brazil. Interestingly, the only conjugate pairs in the examples studied are both examples of wide rifts. There is evidence from both the narrow (e.g. Carolina) and wide rifts (e.g. Gabon, Brazil, Nova Scotia) that the locus of crustal thinning is not necessarily confined to the continent/ocean boundary region but, may occur some distance landward of the boundary.

The reasons why the pattern of crustal thinning varies so much along the strike of rifted margins is not well understood. Buck (1991) has argued that the main control on the width of rifting are changes in buoyancy forces associated with crustal thinning and thermal heating. According to his model, narrow rifts form if the lithosphere is initially cold; wide ones where it is warm. Bassi *et al.* (1993) have shown, however, that in the case of large amounts of extension the patterns of rifting are a complex function of continental geotherm, lithospheric composition and strain rate. In their models, narrow rifts tend to form when plasticity is important (i.e. lower temperatures) whereas wide rifts form when strain hardening occurs during the cooling of initially warmer lithosphere. One effect of the

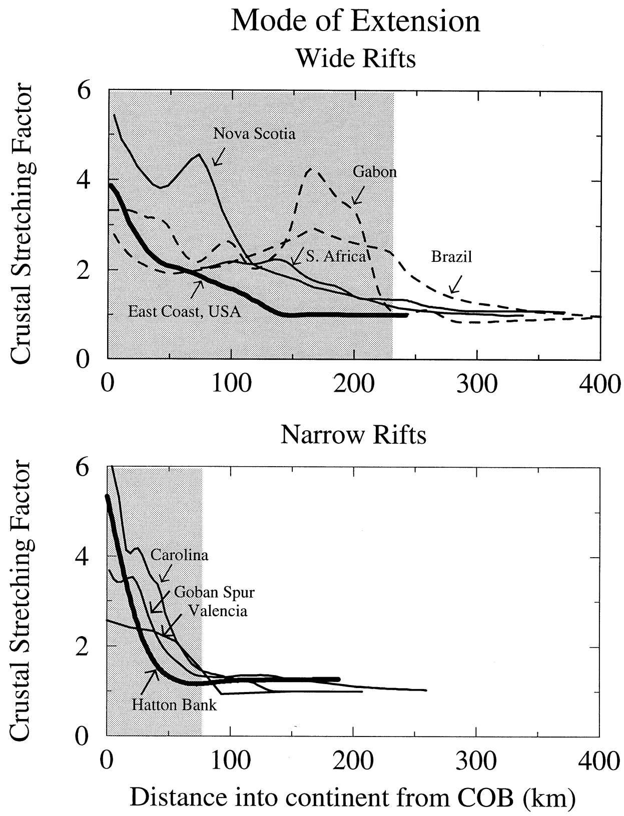

Fig. 5. The mode of extension at continental margins as inferred from crustal restoration studies. The heavy lines show the two studies based on flexural backstripping and gravity modelling: wide rifts—East Coast, USA; narrow rifts—Hatton Bank (this paper). The light lines are based on flexural backstripping studies at the Nova Scotia, Campos Basin, South Africa, Gabon, Goban Spur and Valencia margins. See text for details of data sources.

hardening during cooling is to shift the main locus of thinning from the continent/ocean boundary and into the continent and this may help explain the pattern of crustal thinning seen in Fig. 5 at the Gabon and Brazil margins.

Other factors may serve to focus rifting. The Hatton Bank margin is a narrow rift that is associated with large amounts of magmatism. Melting may therefore control the style of rifting at some margins. This view is supported by data from the Valencia Trough and Carolina margins. Both these margins are narrow rifts which are associated with magmatic activity. The main exception is Goban Spur which is a narrow rift with limited magmatism (Horsefield *et al.* 1993). The factors that control rifting therefore remain unanswered. Future work involving seismic reflection and refraction data acquisition on passive margins together with the development of more refined dynamical models offer the most promise, we believe, to address this issue in the future.

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