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Thinning of the Goban Spur continental margin and formation of early oceanic crust: constraints from forward modelling and inversion of marine magnetic anomalies

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SUMMARY

The deep seismic reflection profile Western Approaches Margin (WAM) cuts across the Goban Spur continental margin, located southwest of Ireland. This non-volcanic margin is characterized by a few tilted blocks parallel to the margin. A volcanic sill has been emplaced on the westernmost tilted block. The shape of the eastern part of this sill is known from seismic data, but neither seismic nor gravity data allow a precise determination of the extent and shape of the volcanic body at depth. Forward modelling and inversion of magnetic data constrain the shape of this volcanic sill and the location of the ocean–continent transition. The volcanic body thickens towards the ocean, and seems to be in direct contact with the oceanic crust. In the contact zone, the volcanic body and the oceanic magnetic layer display approximately the same thickness. The oceanic magnetic layer is anomalously thick immediately west of the volcanic body, and gradually thins to reach more typical values 40 km further to the west. The volcanic sill would therefore represent the very first formation of oceanic crust, just before or at the continental break-up. The ocean–continent transition is limited to a zone 15 km wide. The continental magnetic layer seems to thin gradually oceanwards, as does the continental crust, but no simple relation is observed between their respective thinnings.

Key words: continental margin, Goban Spur, magnetic anomalies, ocean–continent boundary.

1 INTRODUCTION

The Goban Spur continental margin is a non-volcanic rifted margin located southwest of Ireland (Fig. 1) and formed by crustal thinning between Europe and North America during the Early Cretaceous. The Goban Spur continental margin is linear over 200 km and is characterized by a few tilted fault blocks trending parallel to the margin. The deep seismic profile WAM (Western Approaches Margin, BIRPS & ECORS 1989) shot perpendicular to this margin is therefore representative of the whole margin. It crosses the ocean–continent transition in the vicinity of the westernmost topographic feature of the continental margin (Fig. 1), as revealed by gravity modelling (Louvel *et al.* 1992) and seismic data displaying characteristic oceanic basement hyperbolae west of this area (Masson, Montadert & Scrutton 1985). A narrow, high-amplitude (500 nT peak-to-peak) magnetic anomaly is observed between the two westernmost tilted blocks of the margin (Fig. 1) and is associated with a significant gravity anomaly (Louvel *et al.* 1992). On WAM, these anomalies correspond to a volcanic sill drilled at Deep Sea Drilling Project (DSDP) site 551 (de Graciansky & Poag 1985) at a depth of 4030 m, and lying between the two tilted block faults and located within the sedimentary section (Fig. 1). The thickness of the sill decreases eastwards, which suggests that the origin of the magma is located within or close to the westernmost tilted block.

The purpose of this paper is to perform forward modelling and inversion of magnetic-anomaly data along profile WAM in order to constrain the thickness of the magnetic layer along the profile. The results obtained by these two methods help to locate the ocean–continent transition and lead to a better understanding of the formation of the passive margin and the initial production of oceanic crust.

2 FORWARD MODELLING

A compilation of all available magnetic data (Fig. 2, Sibuet 1987) shows that a magnetic anomaly, approximately 25 km

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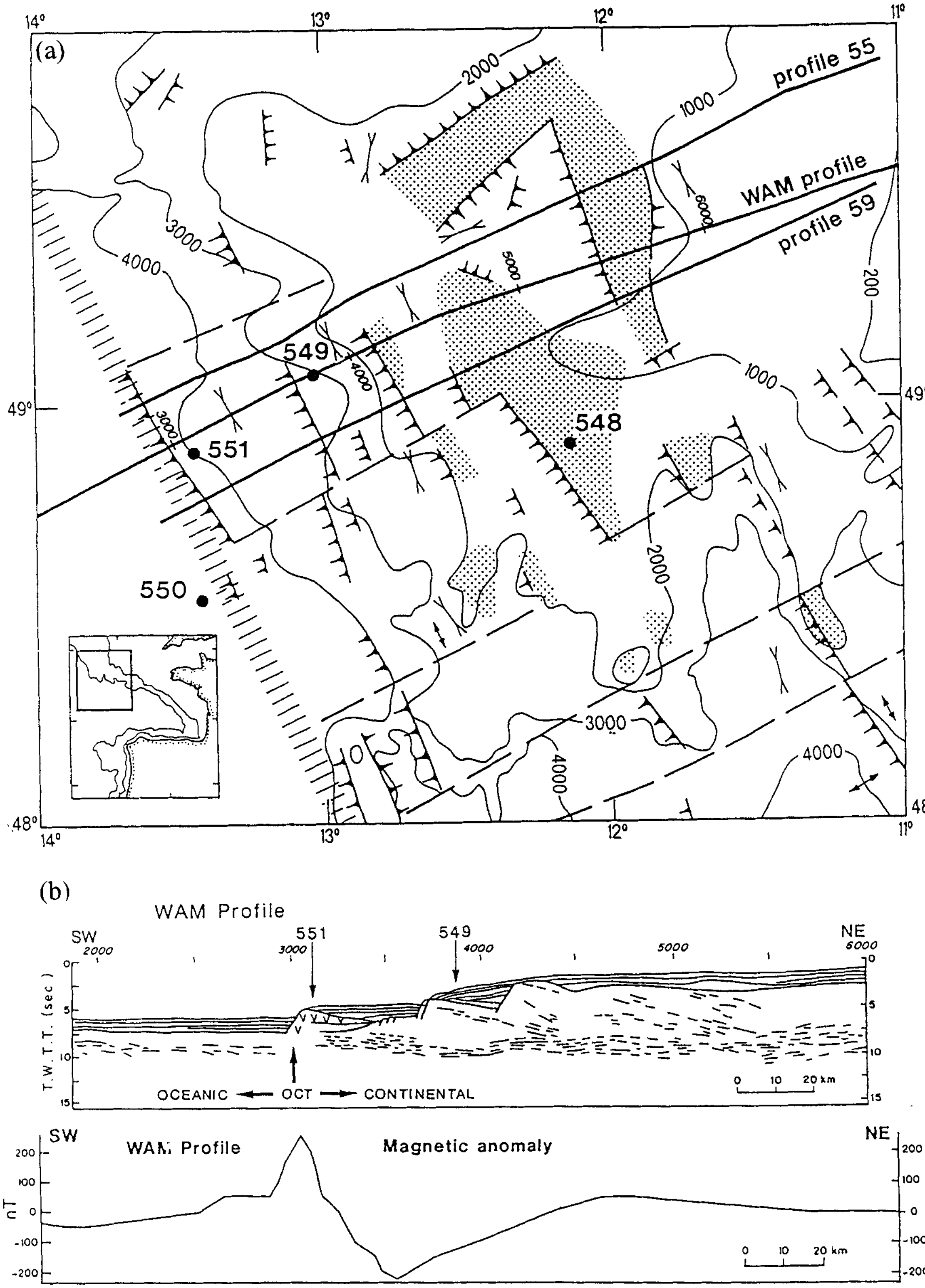

Figure 1. (a) Schematic structural map of the Goban Spur passive margin (Sibuet 1987) with the location of profiles WAM and Norestlante 55 and 59. DSDP sites 548–551 are denoted by black dots; long-dashed lines: transverse faults; lines with orthogonal ticks: normal faults; dotted pattern: lower Cretaceous eroded surfaces; dashed pattern: ocean–continent transition. (b) Line drawing of profile WAM and associated magnetic anomaly.

wide and 125 km long, is parallel to the base of the margin. Basalts drilled at DSDP site 551 (de Graciansky *et al.* 1985) suggest that this magnetic anomaly is created by a basaltic unit of similar dimensions. Seismic profiles show that the western portion of this basaltic unit overlays thinned continental crust, and thickens towards the ocean (Fig. 1; Pinet *et al.* 1991). The eastern portion of the basaltic unit is a sill located within the syn-rift sediments of the westernmost half-graben. According

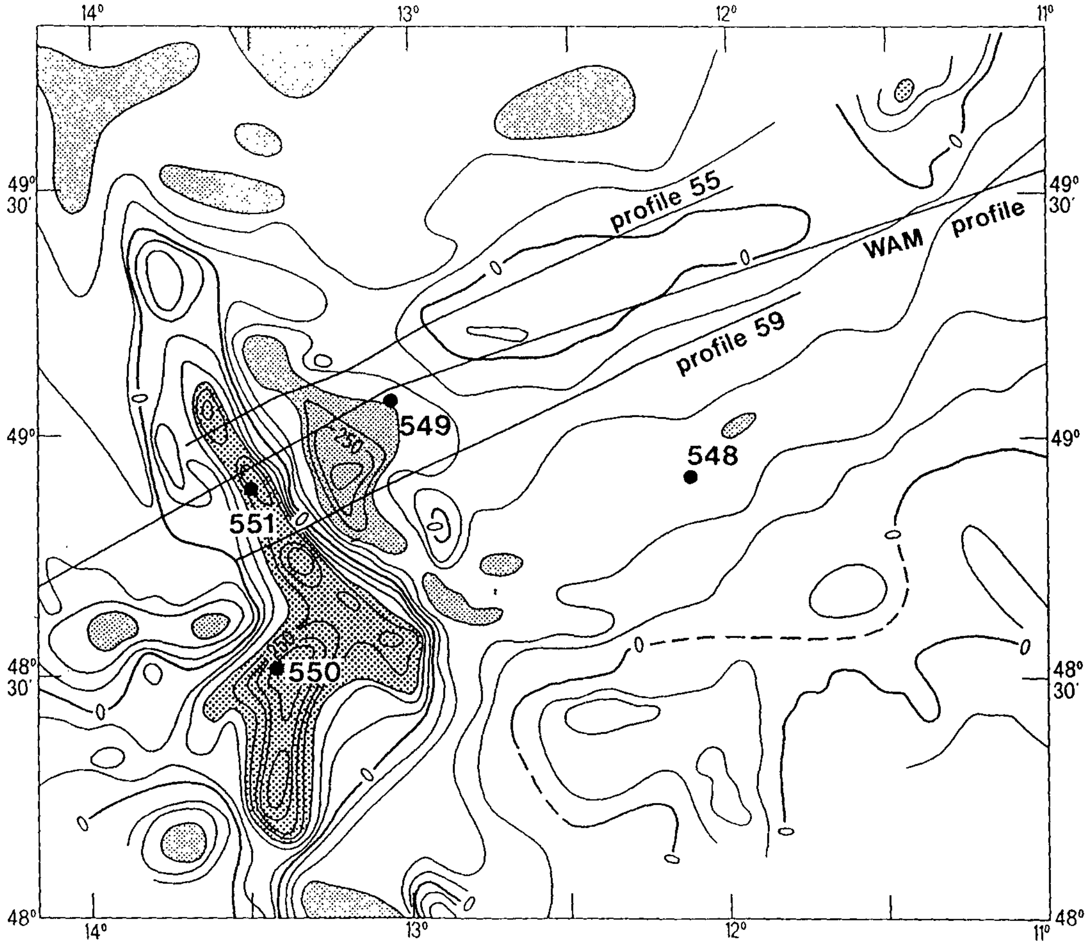

Figure 2. Magnetic-anomaly map of the Goban Spur area (Sibuet 1987) established from marine data. The contour interval is 50 nT. Dotted areas correspond to anomalies higher than 150 nT, fine dotted areas to anomalies lower than 150 nT. The horizontal extension of the volcanic sill located on the deepest tilted block is underlined by an elongated NW–SE anomaly.

to DSDP results and seismic correlations, this sill was emplaced about 106 Ma ago, i.e. mid-Cretaceous (Pinet *et al.* 1991). It corresponds to the first magmatic event occurring at the end of rifting between the Goban Spur and North Flemish Cap conjugate margins. Seismic data show that the surface of the basaltic sill is flat and that the sill fills the half-graben depression, suggesting the occurrence of a fluid magmatic event. The magnetic source is therefore clearly located within the westernmost tilted block, but uncertainties remain about its precise location and the volume of volcanic material. Most of the seismic sections are of poor quality, and many diffractions are present, making interpretation difficult. Gravity modelling for profile WAM does not provide significant constraints on the lower bound of the volcanic structure (Louvel *et al.* 1992). Here, magnetic modelling is used to constrain the location and geometry of this structure better and to confirm the location of the ocean–continent transition (OCT). These objectives can be addressed reasonably well because the sharp contrast in magnetic properties between volcanic rocks (oceanic crust and basaltic sill) and igneous and metamorphic rocks (continental crust) generates a strong magnetic anomaly.

Profile WAM cuts across the Goban Spur passive margin almost orthogonally. Although a second-order geological feature lying obliquely to the margin has been identified on a detailed tectonic map (Sibuet *et al.* 1992), the margin is basically a 2-D feature, as shown by the bathymetric (Sibuet *et al.* 1985), tectonic (Fig. 1), gravity (Sibuet *et al.* 1990) and magnetic-anomaly (Fig. 2) maps. Although 3-D modelling (and inversion) would be a more accurate way to study the magnetic anomalies of the Goban Spur margin, the use of 2-D magnetic modelling (and inversion) is justified because of the dominantly 2-D features observed on this margin. The simplifying assumptions of our modelling (and inversion) efforts, required by the fundamental non-uniqueness of the potential field solutions, prevent such kinds of exercises from providing more than a description of first-order effects. The elongated magnetic anomaly located between the two westernmost tilted blocks is clearly a 2-D anomaly across the profiles studied. Local details of the anomaly may correspond to second-order variations of the geometry and/or magnetization of the source bodies. The magnetic-anomaly profile for WAM (Fig. 1, bottom) was constructed from the magnetic-anomaly map (Fig. 2) and was not recorded while aquiring the deep seismic data.

In continental areas, magnetic anomalies are mostly due to induced magnetization, whereas, in oceanic domains, the effects of remanent magnetization acquired by the crust at spreading centres are clearly dominant. Adjacent to Goban Spur, the oldest recognized magnetic anomaly is anomaly 34 (Masson *et al.* 1985), located about 115 km west of the ocean–continent transition. The oceanic crust located between anomaly 34 and the margin was created during the long Cretaceous period of normal geomagnetic polarity, at approximately its present latitude (Dyment & Arkani-Hamed 1994). Its remanent magnetization is uniform and approximately parallel to the present geomagnetic field. Therefore, both remanent 'and induced magnetization in the oceanic part of Goban Spur are modelled by a single magnetization vector, associated with the 115 km wide domain of oceanic crust created between the OCT and magnetic anomaly 34 (83 Ma, Cande & Kent 1992). For simplicity, we assume that the basaltic sill was also created during the long Cretaceous period of normal geomagnetic polarity and that it presents similar magnetic properties. The values of declination and inclination used in the computation are $-10.4°$ and $64.5°$, respectively.

The effect of the basaltic unit on the magnetic anomaly is estimated by modelling a single body including both the oceanic crust and the basaltic sill (Fig. 3a), with the 5 A m^{-1} magnetization typical for oceanic basalts at this latitude (e.g. Bradley & Frey 1991), by using the alogorithm of Won & Bevis (1987). The oceanic basement is supposed to be flat, 6000 m deep, and the top of the basaltic unit has been defined from seismic profile WAM. The lower part of the body has been adjusted so that the computed anomaly fits the observed anomaly. As the magnetic anomalies are generated by magnetization contrasts, the average level of these anomalies is unimportant and only reflects the reduction of regional field effects. It is therefore not considered, and the average of the synthetic magnetic anomalies is zero. West of km 140, the fit between synthetic and observed anomalies is good (Fig. 3a). In the continental domain however, the discrepancy between the two curves underlines the need for non-zero magnetization in the thinned continental crust.

In a second attempt, a magnetization of 0.5 A m^{-1}, 10 times lower than for the oceanic crust, has been assumed for both the continental crust and the overlying sediments to account for the different magnetizations of continental and oceanic rocks. With respect to this low magnetization value, the sediments are not considered as an independent layer, and the top of the continental structure is assumed to correspond to the seafloor. This approximation does not significantly affect the results, and avoids uncertainties related to the definition of sedimentary-layer boundaries and magnetization values. To be consistent, the same magnetization has been assigned to the sediments in the oceanic domain. These assumptions require a continental magnetized body about 17 km thick in order to fit synthetic and observed anomalies (Fig. 3b). The

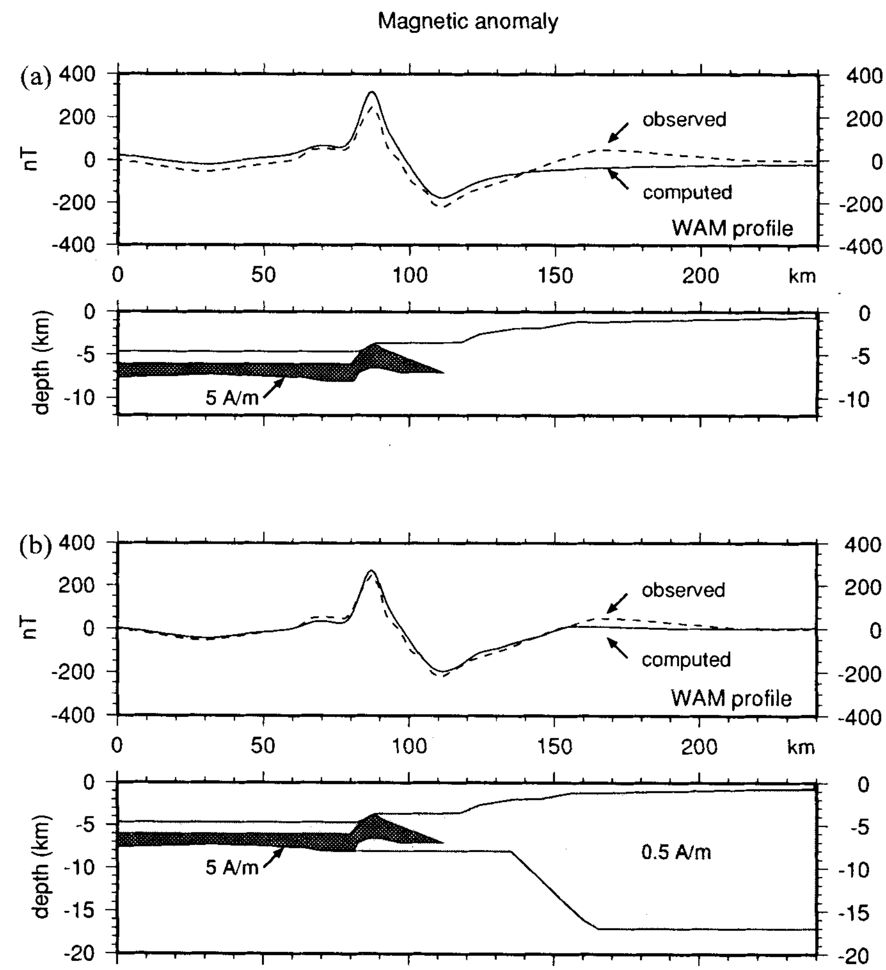

Figure 3. Forward modelling of magnetic anomalies along profile WAM, observed (dashed line) and computed (solid line). (a) a single magnetic body which represents the oceanic domain and the basaltic unit with a magnetization of 5 A m^{-1}, and (b) two magnetic bodies which represent the oceanic crust and basaltic unit with a magnetization of 5 A m^{-1} and the continental crust with a magnetization of 0.5 A m^{-1}.

thickness of the magnetized continental layer increases from km 80 to km 160, and remains constant eastwards, where the continental crust is not thinned (Fig. 3b).

This modelling has involved adjusting the shape of the lower part of the bodies to reach a reasonable solution. A less subjective approach, involving the inversion of magnetic anomalies in the presence of topography, is presented in the next section.

3 INVERSION

3.1 Method

The source of a magnetic anomaly is characterized by its geometry and magnetic properties. We have assumed that the anomaly is caused by a layer of constant magnetization and that the top of this layer is the basement topography. An expression for a magnetic anomaly created by 2-D rectangular prisms (i.e. infinite in the horizontal direction perpendicular to the profile) observed at a constant altitude is given by Parker (1972) in the Fourier domain. From this expression, Parker & Huestis (1974) have proposed an iterative spectral method for the inversion of magnetic anomalies in the presence of topography. They assume a magnetic layer of constant thickness and obtain lateral variations of the magnetization within this layer. The aim of our study is to define the source geometry, i.e. the shape of the lower boundary of the magnetic layer, instead of its magnetization. For this purpose, the method of Parker & Huestis (1974) has been modified to provide the thickness of the magnetic layer assuming a constant magnetization. The latter assumption is geologically reasonable: as previously shown, the oceanic crust and the basaltic sill can be modelled by using a single magnetization vector. Only the OCT is associated with a strong magnetization contrast. However, separate inversions can be performed assuming a constant magnetization for the oceanic crust and the basaltic sill and another constant magnetization for the continental crust.

The development of Parker's (1972) equation assuming a layer of constant magnetization M_0 and of varying thickness e leads to, after expansion of the exponential functions in series, rearrangement of summations, and extraction of the first term of the summations,

$$\mathrm{FT}(e)=\frac{1}{|k|}\frac{T(k)}{f(k)}+\frac{1}{|k|}\sum_{L=2}^{\infty}\frac{(-|k|)^L}{L!}[\mathrm{FT}((h+e)^L)-\mathrm{FT}(h^L)], \tag{1}$$

where FT is the Fourier Transform, k the wavenumber along the profile, $T(k)$ the Fourier transform of the magnetic anomaly, H a reference depth generally chosen as the minimum bathymetry, h the depth from H to the top of the layer, and

$$f(k)=2\pi M_0\left(\frac{\sin(k/2)}{k/2}\right)e^{-|k|H}, \tag{2}$$

This expression is applied iteratively, by using the ith estimate of e on the right-hand side of eq. (1) to compute the $(i+1)$th estimate of e. The 'initial thickness', a constant thickness adopted for the first iteration, is critical, as the inversion procedure leads to a family of solutions, each one corresponding to a given initial thickness. Values of the initial thickness that are not geologically reasonable lead to incoherent results. The iteration process ends when the root-mean-square (RMS) difference between two successive estimates of e is less than a given quantity, which should be choosen to ensure that the process ends after a small number of iterations. Our experience has shown that trying to reach an unreasonably low RMS misfit results in the appearance of ficticious high frequencies of e, and that convergence of the algorithm is usually obtained for a sampling interval similar to the reference depth H. Once the inversion process is complete, the solution is checked by forward modelling. The quality of the fit between the observed and recomputed anomalies provides an additional constraint for choosing the best value of the initial thickness.

3.2 Application to the Goban Spur magnetic anomaly profiles

In addition to profile WAM, the inversion procedure was applied to two parallel profiles, profiles Norestlante 55 and 59, on which single-channel seismic data are available (Figs 1 and 2). As for profile WAM, magnetic anomalies along profiles Norestlante 55 and 59 were extracted from the map (Fig. 2). We assume that the top of the magnetic layer in both oceanic and continental domains corresponds to the depth of basement obtained from seismic data. To avoid the high frequencies that are induced by the sharp limits of the continental tilted blocks, the depth of basement was smoothed by the application of a triangle convolution filter of width 11 km. The same low-pass convolution filter was applied to the magnetic anomaly. Magnetizations of 5 and 0.5 A m^{-1} are assumed for the oceanic and continental domains respectively. The sampling interval is 1 km, and the reference depths are 1500, 900 and 1800 m for profiles WAM, Norestlante 55 and Norestlante 59, respectively.

Fig. 4(a) shows the inversion result along profile WAM assuming a magnetization of 5 A m^{-1} and an initial thickness for the magnetic layer of 1500 m for the first iteration. The shape of the oceanic and volcanic features west of km 110 is in very good agreement with our previous forward models (Figs 3a and 3b). The thickness of the oceanic magnetic layer varies from 1200 to 1500 m. As shown by forward modelling, this layer is thicker between km 70 and km 80, just west of the OCT. The thickness of the basaltic unit increases oceanwards, up to 2350 m. In addition, the inversion suggests that the continental crust is slightly magnetized and that the thickness of the magnetic layer increases between km 110 and km 180 and is constant east of km 180. Obviously, the thickness of 2 km obtained by the inversion is unrealistic, and so is a magnetization of 5 A m^{-1} for the continental crust. Fig. 4(b) shows the inversion result assuming a magnetization of 0.5 A m^{-1} and an initial thickness for the magnetic layer of 13 km for the first iteration. The thickness of the magnetic layer is, of course, meaningless in the oceanic domain, as the magnetization is too low. The initial thickness of 13 km results in a good fit east of km 100. The geometry of the continental magnetic layer is similar to that previously obtained by forward modelling (Fig. 3b). Note that, for a given magnetization, the initial thickness of the magnetic layer used for the first iteration must be carefully chosen by using both inversion and forward modelling for a large set of possible values. The RMS residual between computed and observed anomalies has been calculated for a set of initial thickness values, assuming a magnetization value of 0.5 A m^{-1} (Fig. 5). For initial thicknesses less than or equal to 12 km, the inversion leads to a negative thickness around km 110 on the profile. The best realistic solution,

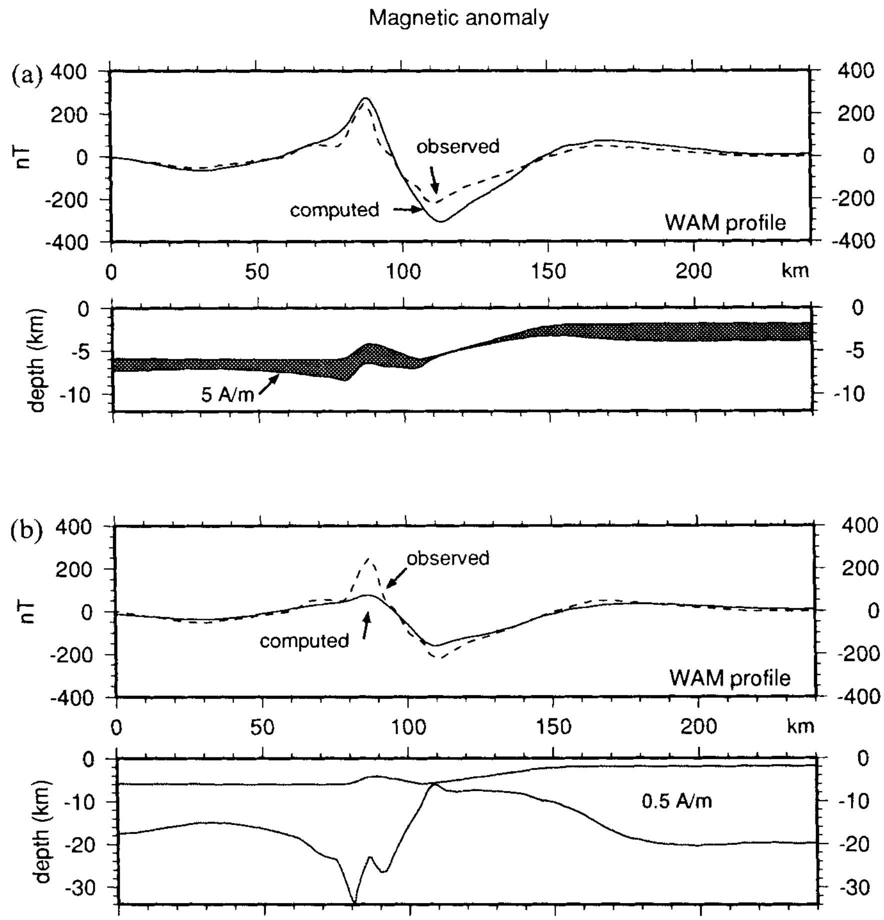

Figure 4. Inversion of magnetic anomalies along profile WAM. Top: observed (dashed line) and computed (solid line) magnetic anomalies. Bottom: magnetic body obtained by inversion assuming (a) a magnetization of 5 A m^{-1} and an initial thickness of the oceanic magnetic layer of 1.5 km; (b) a magnetization of 0.5 A m^{-1} and an initial thickness of 13 km.

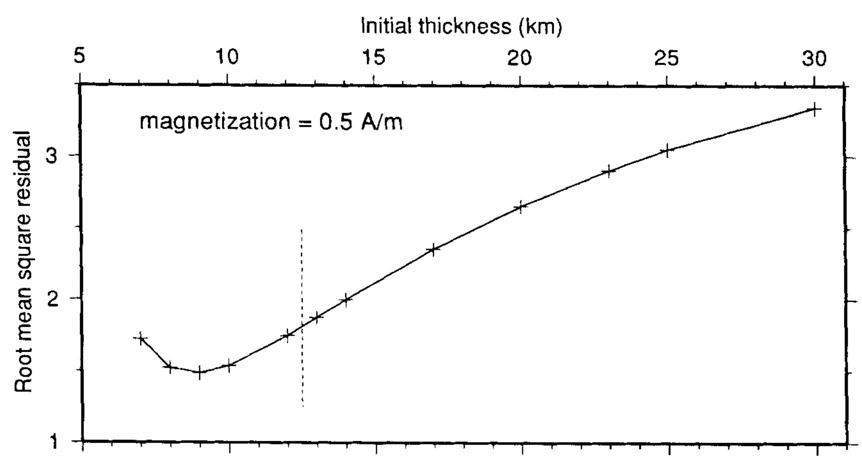

Figure 5. Root-mean-square residual between observed and computed magnetic anomalies versus the initial depth assuming a magnetization of 0.5 A m^{-1}. For initial thicknesses smaller than 12 km (vertical dotted line), the inversion process leads to negative thicknesses and its results are therefore discarded.

therefore, corresponds to an initial thickness of 13 km (Fig. 5). Models with a thicker magnetic layer tend to have worse fits where the observed signal has a high frequency content, as the deeper interface smoothes these high frequencies in a way similar to upward continuation. A composite model, taking into account realistic magnetizations for both the oceanic and continental magnetic layers, is displayed in Fig. 6. The continental magnetic layer probably extends under the basaltic unit west of km 112. Although its geometry cannot be constrained from our inversion scheme between km 80 and km 112, its contribution to the magnetic anomaly is probably negligible with respect to that of the basaltic unit.

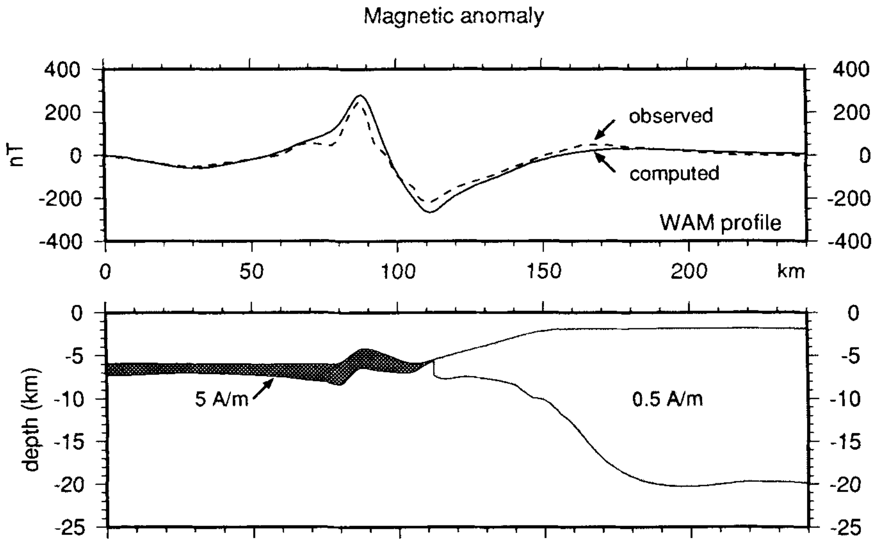

Figure 6. Composite model including results shown on Fig. 4(a) west of km 112 for the oceanic domain and on Fig. 4(b) east of km 112 for the continental domain. The resulting computed magnetic anomaly is in good agreement with the observed one.

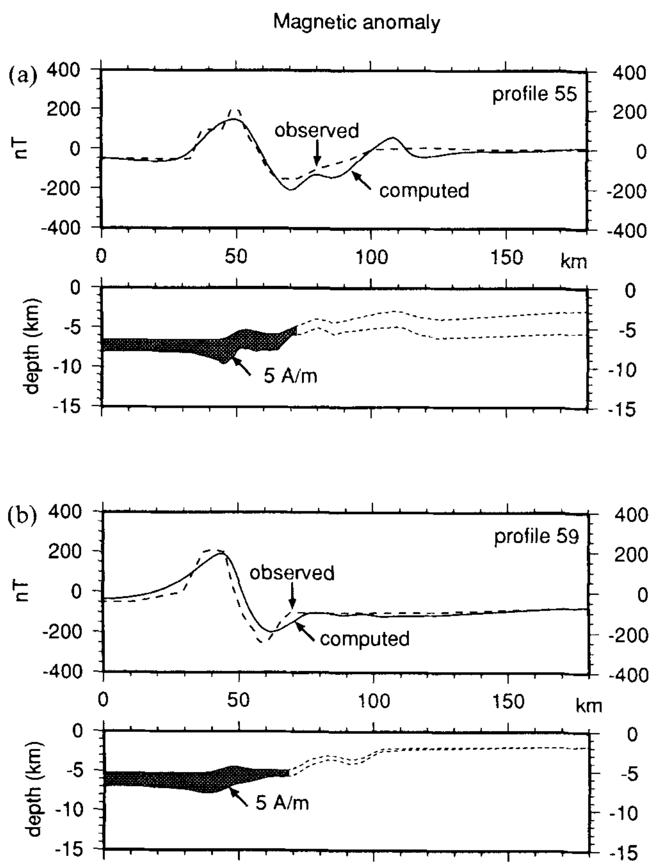

Figure 7. Inversion of magnetic anomalies along profiles Norestlante 55 (a) and 59 (b). Top: observed (dashed line) and computed (solid line) magnetic anomalies. Bottom: magnetic body obtained by inversion assuming a magnetization of 5 A m^{-1} and an initial thickness of the oceanic magnetic layer of 2.5 km (a) and 0.5 km (b). Dotted lines show limits of the resulting magnetic body in the continental domain, where a magnetization of 5 A m^{-1} is unrealistic.

The inversion of profiles Norestlante 55 and 59 leads to similar results to that from WAM for the oceanic domain, as shown in Fig. 7 for computations with a magnetization of 5 A m^{-1}. Initial thicknesses of 2500 and 500 m were respectively adopted in order to ensure a thickness of the oceanic magnetic layer similar to that computed from profile WAM in normal oceanic crust at the westernmost end of the profiles. Inversion along profile Norestlante 55 is consistent with landward thickening of the continental magnetic layer (Fig. 7a), as observed on profile WAM. Inversion with a magnetization of 0.5 A m^{-1} (not shown) yields thicknesses increasing from about 3 km east of the basaltic sill (km 70) to 20 km at the eastern end of the profile. The lower boundary of the continental magnetic layer is not as regular as along profile WAM, with local variations reflecting the basement topography. Inversion along profile Norestlante 59 leads to negligible thickness variations of the continental magnetic layer, whatever the magnetization and initial thickness (e.g. Fig. 7b). Such a result disagrees with those from profiles WAM and Norestlante 55, and emphasizes the limits of our method. This may be related to an erroneous interpretation of the poor seismic data along the Norestlante profiles, but more probably reflects the inadequacy of assuming a constant magnetization for continental areas.

4 DISCUSSION AND CONCLUSIONS

4.1 Oceanic crust

Because the oceanic magnetization is relatively well constrained (de Graciansky & Poag 1985), the geometry of the oceanic magnetic layer and the basaltic sill deduced from both forward modelling and inversion can be considered to be realistic. The thickness of the volcanic sill east of km 95 computed by inversion of the magnetic-anomaly data corresponds to that deduced from the seismic profile WAM. West of km 95, the magnetic analysis helps to constrain the shape of the volcanic body and the location of the OCT. The volcanic sill located between km 75 and km 110 (Figs 3 and 6) thickens towards the ocean, and displays a nearly horizontal lower boundary. At km 75, the volcanic unit seems to be in direct contact with the oceanic crust. Although this could be an artefact generated by grouping both the oceanic magnetic layer and the volcanic unit in a single structure, the magnetic layer is almost twice as thick between km 70 and km 80 (Figs 3 and 6), in the close vicinity of the basaltic unit, than at km 30, where it is supposed to be more representative of typical oceanic crust. At km 75, the thick oceanic magnetic layer and the volcanic unit have approximately the same thickness, and the oceanic magnetic layer progressively thins between km 70 and km 50. This suggests, if assumptions about similar magnetic properties are valid, some kind of continuity between the two features. The same conclusion can be drawn from the analysis of profiles Norestlante 55 and 59 (Fig. 7). Initiated either slightly before or at the continental break-up, the basaltic volcanism would therefore represent the first step of oceanic crust emplacement. Refraction data show that the oceanic crust adjacent to the Goban Spur margin is 5.4 km thick (Horsefield *et al.* 1994), which is slightly less than the mean crustal initial thickness of 6.0 km of normal oceanic crust (e.g. Chen 1992). Both forward modelling and inversion of magnetic-anomaly data show that the oceanic magnetic layer, which more or less corresponds to the basaltic layer, is thicker in the same area. This apparent contradiction vanishes if part of the thick oceanic basalts and the continental basaltic unit are indeed related to the same source. This assumption is supported by the geochemical anlysis performed on the basalts of the volcanic unit of DSDP site 551, which shows that they are typical of oceanic tholeiitic volcanism (de Graciansky *et al.* 1985). The melt was initiated during the rise of asthenosphere beneath the thinned area, and was erupted slightly before and at the break-up, forming the basaltic unit, and partly after the break-up, thickening the oceanic basaltic layer. In this respect, the Goban Spur continental margin can be considered as a slightly volcanic margin, intermediate between true volcanic margins such as the Vøring Plateau (e.g. Eldhom 1991) and, to a lesser extent, Hatton Bank (e.g. White 1992), and non-volcanic margins such as the Galicia margin (e.g Sibuet *et al.* 1987). The gradation of these northeastern Atlantic margins, from highly magmatic to purely tectonic processes, probably reflects the thermal structure of the asthenosphere at the time of continental break-up.

The ocean–continent transition is restricted to a zone 15 km wide located between km 75 and km 90, excluding the area where the volcanic sill is intercalated within the syn-rift sediments of the thinned continental crust. The onset of oceanic accretion takes place at km 80, where syn-rift sediments disappear (Pinet *et al.* 1991). It is of great interest to perform a similar magnetic analysis on the conjugate margin, namely the Northern Flemish Cap, to investigate whether an equivalent volcanic unit exists on this margin. Preliminary results from a recent cruise (Louvel 1995) suggest a symmetrical ocean–continent transition: a basement high located at the transition corresponds to a thicker magnetic layer, similar to the basaltic unit on profile WAM; to the east, an anomalous thick magnetic oceanic layer gradually thins oceanwards, and to the west, the magnetic layer abruptly thins towards the continent (assuming a $5\ \mathrm{A\ m^{-1}}$, oceanic-like magnetization). Shorter seismic and magnetic profiles on the Flemish Cap prevent further investigation on the continental domain.

4.2 Continental crust

As shown by the somewhat contradictory results obtained by the inversion along profiles WAM and Norestlante 55 on one hand, and Norestlante 59 on the other hand, the validity of our results is questionable in the continental domain. The major reason for this is that the inversion method assumes a constant magnetization, while the continental crust is made of a variety of terranes having different physical properties. Moreover, as our choice of magnetization of $0.5\ \mathrm{A\ m^{-1}}$ is poorly constrained, the thickness variations of the continental magnetic layer computed by forward and inverse methods across the Goban Spur margin should be examined in a qualitative way: a stronger (weaker) magnetization would result in smaller (larger) thickness variations. Profile Norestlante 59, located south of the area, may be affected by the proximity of the Celtic margin, which continues obliquely the Goban Spur passive margin to the south and was formed by the opening of the Bay of Biscay between Europe and Iberia. The landward thickening of the continental magnetic layer observed along profiles WAM and Norestlante 55 is similar to that, constrained by forward modelling of both surface and deep-tow magnetic-anomaly data (Sibuet *et al.* 1995), observed across the Galicia passive margin created by the separation of Iberia and North America. A similar variation has been observed for the whole crust, as well as for the lower and upper crusts, from seismic and gravity data along profile WAM (Louvel *et al.* 1992). Fig. 8 shows the continental and oceanic crusts derived from gravimetric models along profile WAM (Louvel *et al.* 1992), and the continental magnetic layer obtained by inversion with

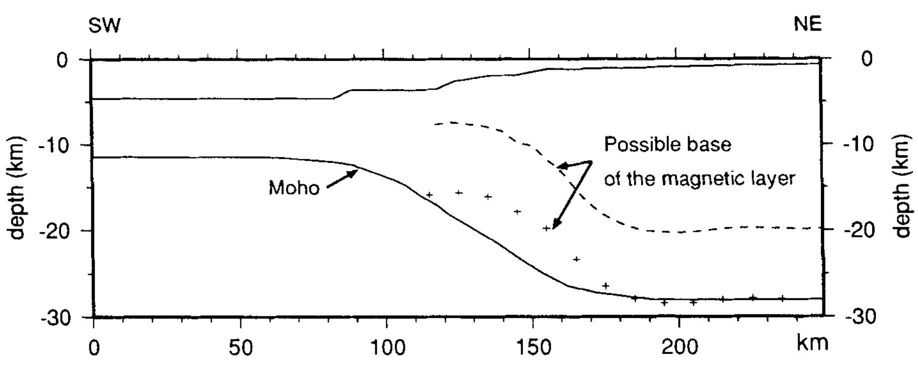

Figure 8. Comparison between the Moho obtained by seismic refraction data and gravity modelling (solid line), and the base of the magnetic layer computed by inversion with maximal thicknesses of 17 km (dashed line) and 28 km (crosses) along profile WAM.

various initial thicknesses. Comparison between thinnings of the magnetic layer and the continental crust along profile WAM shows that both are geographically linked, suggesting that both are related to the rifting and break-up of the Goban Spur–North Flemish Cap conjugate margins. This geographical relation is consistent with the pure shear mechanism suggested from the analysis of seismic and gravity data (Louvel *et al.* 1992). However, the thickness of the magnetic layer across the Goban Spur margin seems to decrease more than expected for continental crust thinned by pure shear. There may be, therefore, an additional loss of magnetization through a process related to the thinning of the continental crust, and possibly to the increasing number of upper crustal fractures and fissures with increasing thinning: sea-water circulation through the fracturation net alters the magnetic minerals and would therefore reduce the upper crustal magnetization with increasing thinning. Such a process would be superimposed on the lesser bulk magnetization of the continental crust caused by the thinning itself, and, under the assumption of a constant magnetization, yield an apparently thinner magnetic layer.

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