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The age and origin of the central Scotia Sea

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SUMMARY

Opening of the Drake Passage gateway between the Pacific and Atlantic oceans has been linked in various ways to Cenozoic climate changes. From the oceanic floor of Drake Passage, the largest of the remaining uncertainties in understanding this opening is in the timing and process of the opening of the central Scotia Sea. All but one of the available constraints on the age of the central Scotia Sea is diagnostic of, or consistent with, a Mesozoic age. Comparison of tectonic and magnetic features on the seafloor with plate kinematic models shows that it is likely to have accreted to a mid-ocean ridge between the South American and Antarctic plates following their separation in Jurassic times. Subsequent regional shallowing may be related to subduction-related processes that preceded backarc extension in the East Scotia Sea. The presence of a fragment of Jurassic–Cretaceous ocean floor in the gateway implies that deep water connections through the Scotia Sea basin complex may have been possible since Eocene times when the continental tips of South America and the Antarctic Peninsula first passed each other.

Key words: Tectonics and climate interactions; Kinematics of crustal and mantle deformation; Antarctica; South America.

1 INTRODUCTION

Drake Passage (Fig. 1) is the narrowest part of a Cenozoic circum-Antarctic seaway whose presence is necessary for the existence of the Antarctic Circumpolar Current. Today, this current sustains a set of oceanographic fronts that isolate Antarctica from warmer surface waters further north. This observation was the basis of an idea that the current's onset directly led to the buildup of perennial Antarctic ice at the Eocene–Oligocene boundary by restricting meridional heat transport (Kennett *et al.* 1975). Although computer model experiments have shown that this transport appears not to be particularly sensitive to gateway presence, some still show Antarctic cooling following the establishment of the Antarctic Circumpolar Current, especially under high atmospheric CO_2 conditions (Huber *et al.* 2004; Sijp & England 2004, 2009; Sijp *et al.* 2009). Despite this, there are stronger indications that the very sudden ice buildup at the boundary was triggered by orbital changes that favoured cool polar summers after a period of climatic preconditioning by declining atmospheric CO_2 concentrations (DeConto & Pollard 2003; Coxall *et al.* 2005).

Against this background, a desire to know the timing of the onset of Antarctic circumpolar flow long drove research into the tectonics of Drake Passage. The passage appeared during the growth of the Scotia Sea, which formed by the relative motions of small plates in the space between an east-migrating arc and trench in the east, and the South American and Antarctic plates in the west. For years the idea of local barriers to flow persisting into Oligo–Miocene times (Barker & Burrell 1982) was popular and, despite the subsequent weakening of evidence for a barrier at the Shackleton Fracture Zone (Livermore *et al.* 2004), the idea of complete or partial blockage at some other location remains influential because it can be seen as consistent with paleoceanographic and seismic evidence for the onset of, or changes in, the Antarctic Circumpolar Current as late as 20 Ma (Pfuhl & McCave 2005; Koenitz *et al.* 2008; Lagabrielle *et al.* 2009). Pfuhl & McCave (2005) noted that a ∼24 Ma onset Antarctic Circumpolar Current coincides with the Antarctic Mi-1 glaciation, and so may have caused it, and the accompanying global changes, by inhibiting meridional heat transport to Antarctica.

On the other hand, studies of small basins at the margins of the Scotia Sea have given rise to various suggestions of earlier deep and shallow water pathways, which are consistent with other lines of paleoceanographic evidence for Pacific–Atlantic connectivity (Lawver & Gahagan 1998; Eagles & Livermore 2002; Eagles *et al.* 2005, 2006; Livermore *et al.* 2005, 2007; Scher & Martin 2006; Ghiglione *et al.* 2008). Eagles *et al.* (2006) noted that because the onset of Eocene shallow connectivity at Drake Passage would have given rise to global changes in ocean circulation (Toggweiler & Bjornsson 2000) and coincided with changes in the timescales of carbon sequestration (Kurtz *et al.* 2003), that it may have contributed to the Cenozoic decline in atmospheric CO_2.

Hence, although the pattern of the Antarctic Circumpolar Current's onset and its role in Cenozoic climate change seem now to be less clear-cut than first hypothesized, the questions of its onset and of the timing of initial Pacific–Atlantic exchange remain very important. Knowledge of the tectonics of Drake Passage is crucial to answering these questions, with the plate kinematics of South Georgia and the central Scotia Sea key uncertainties in that knowledge. The South Georgia microcontinent is often depicted traversing

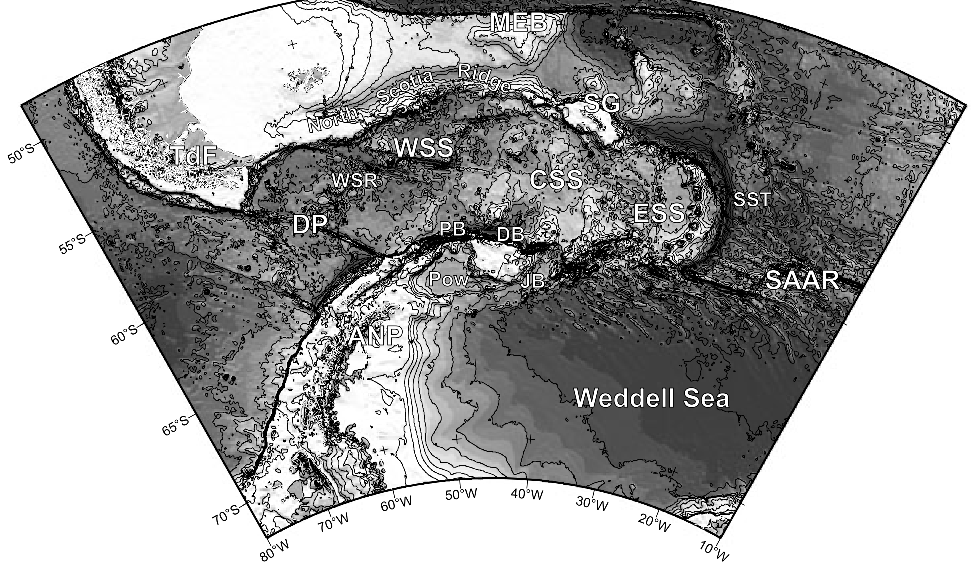

Figure 1. Predicted bathymetry of the southern Atlantic and Scotia and Weddell seas (Smith & Sandwell 1997). ANP, Antarctic Peninsula; CSS, central Scotia Sea; DP, Drake Passage; ESS, East Scotia Sea; MEB, Maurice Ewing Bank; PB, Protector Basin; Pow, Powell Basin; JB, Jane Basin; SAAR, South American–Antarctic Ridge; SG, South Georgia; SST, South Sandwich Trench; TdF, Tierra del Fuego; WSR, West Scotia Ridge; WSS, west Scotia Sea.

the Scotia Sea during Oligocene times from an original position near Cape Horn (e.g. Barker *et al.* 1991). At the same time, Hill & Barker (1980) proposed that north–south oriented spreading in the central Scotia Sea occurred in a Miocene backarc basin on whose northern margin South Georgia, drifting northwards, contrived along with arc volcanism to hinder deep water flow until as late as 22–17 Ma (Barker 2001; Lawver & Gahagan 2003; Maldonado *et al.* 2003). Alternatively, as there are difficulties accounting fully for South Georgia's dispersal from a position far to the west (Cunningham *et al.* 1995; Eagles 2010), and as it has been suggested that the central Scotia Sea may be much older than Miocene (DeWit 1977; Livermore *et al.* 1994; Eagles *et al.* 2005), there is a possibility that deep connectivity might have been achieved much earlier.

To look at these possibilities, I consider the geological and geophysical data available from the central Scotia Sea and its margins in detail. I go on to outline a basin evolution history that involves its formation in Mesozoic times and that is consistent with the majority of those data. I illustrate this interpretation using tectonic reconstructions and, before concluding, I outline and speculate on the causes of the only significant observation that it is inconsistent with.

2 AGE OF THE CENTRAL SCOTIA SEA

The central Scotia Sea is an area of $\sim$120 000 km^2 that lies at 4200–1800 m depth, with its shallowest points along a north-trending swell at 40°W. Away from this swell, the area is defined geophysically by a set of three corridors of linear, moderately coherent, high-amplitude, east-striking magnetic reversal anomalies (Barker 1970; Fig. 2). Discontinuities between the corridors coincide with linear trough-like free-air gravity anomalies (Fig. 2) whose trends suggest an NNE–SSW spreading azimuth. Currently, the favoured and most explicit model for the formation of the central Scotia Sea in the literature is as a product of north–south directed seafloor spreading that occurred in Miocene times. The genesis of this idea was with Barker (1970) who suggested that the magnetic anomalies formed after 40 Ma, based on their proximity to sequences of young anomalies immediately to the west and east. Following this, Hill & Barker (1980) presented models of Oligocene- and Miocene-aged spreading. Because the anomaly sequence is asymmetrical, each of Hill & Barker's (1980) models bears a good resemblance to the observed anomaly sequence near the assumed axis, but the resemblance degrades to the north. Maldonado *et al.* (2003) favoured a Miocene model, albeit on the basis of a single short magnetic profile that requires a number of spreading rate changes, some quite severe. Similarly, Hill and Barker tentatively favoured a Miocene model because it implies a sudden spreading rate change at chron C5C, and extinction at chron C3, that are contemporaneous with similar events at the neighbouring West Scotia Ridge (Barker & Burrell 1977; Eagles *et al.* 2005), as one might expect to see for two tectonically coupled spreading centres. Fig. 3 shows an updated Miocene model, using Gradstein *et al.*'s (2004) reversal timescale, from which it is evident that this conclusion is not inevitable. The updated model uses no sudden rate change at C5C, and extinction in it occurs at C4A ($\sim$9.2 Ma). From all this, it is appropriate to conclude that the idea of Miocene spreading in the central Scotia Sea can at best be supported, but not diagnosed, by magnetic data.

2.1 Thermal state of the lithosphere

The ages of most of the oceanic basins in the Scotia Sea have been estimated by comparing their bathymetry to depths predicted by thermal subsidence models (King & Barker 1988; Eagles *et al.* 2005, 2006). Using this technique, the isostatically corrected depth of the central Scotia Sea, which is covered by 800–1500 ms (two-way traveltime) of sediments (Barker & Lawver 2000; Maldonado *et al.* 2003; British Antarctic Survey, unpublished data), is consistent with formation during the period 30–10 Ma. Strikingly however, there is no tendency for steep bathymetric gradients and shallower seafloor to occur near Hill & Barker's (1980) young magnetically defined axis, as might be expected and as is seen in the west Scotia Sea

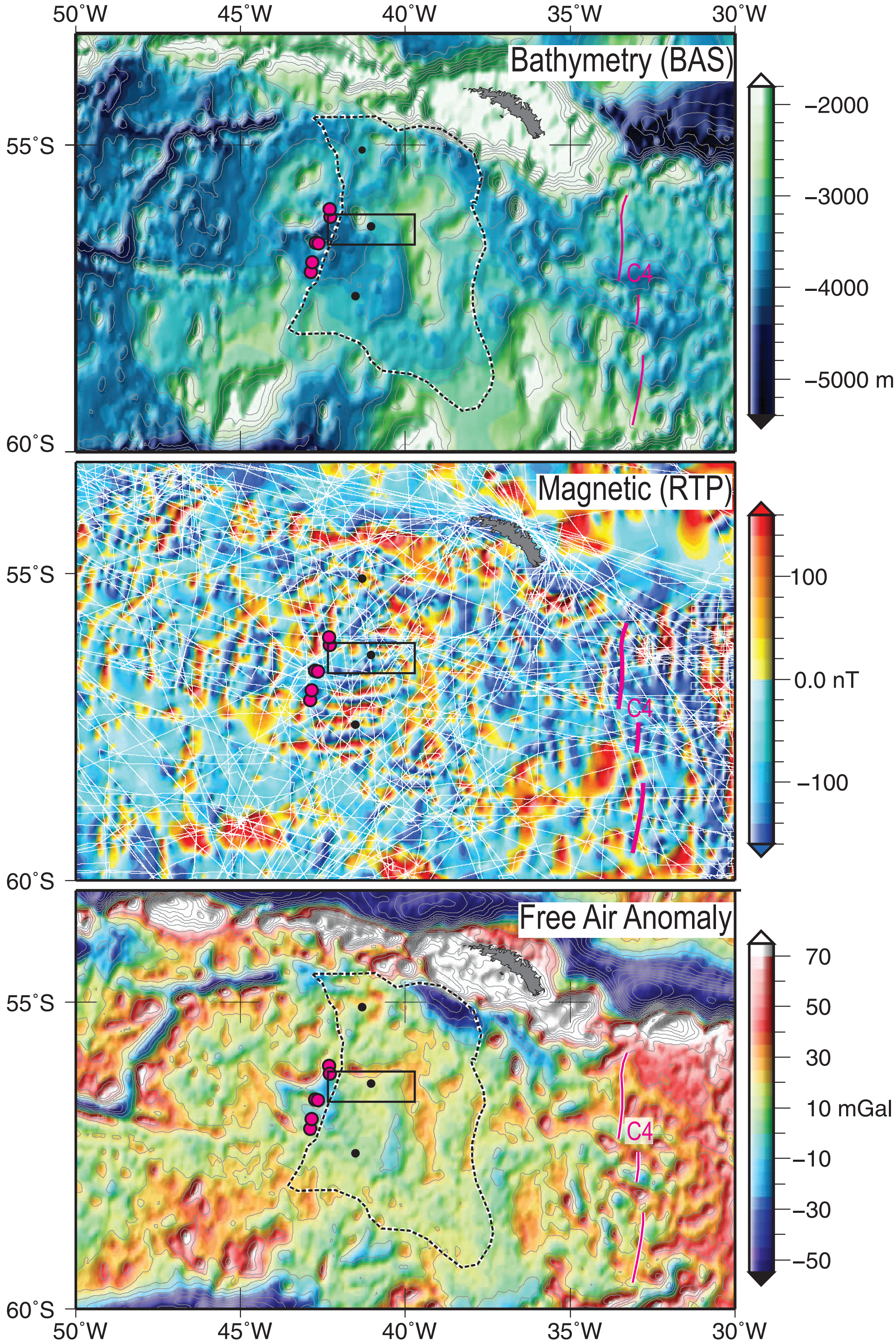

Figure 2. Top panel: bathymetry of the central Scotia Sea, gridded from ship-track data archived at the British Antarctic Survey. Middle panel: grid of magnetic anomalies (Eagles *et al.* 2005) after reduction to the pole. Grey lines: ship tracks contributing data to the grid. Bottom panel: free-air gravity anomalies (Sandwell & Smith 1997). Rectangular box: location of axis proposed by Hill & Barker (1980). Violet discs: C8 magnetic anomaly picks in west Scotia Sea (Eagles *et al.* 2005). Violet lines: magnetic anomaly C4 identification in the East Scotia Sea (Larter *et al.* 2003). Black discs: heat flow sites of Fig. 5 (from north to south, HF-01, -02 and -03). Dotted line: outline of province of E–W magnetic anomaly lineations.

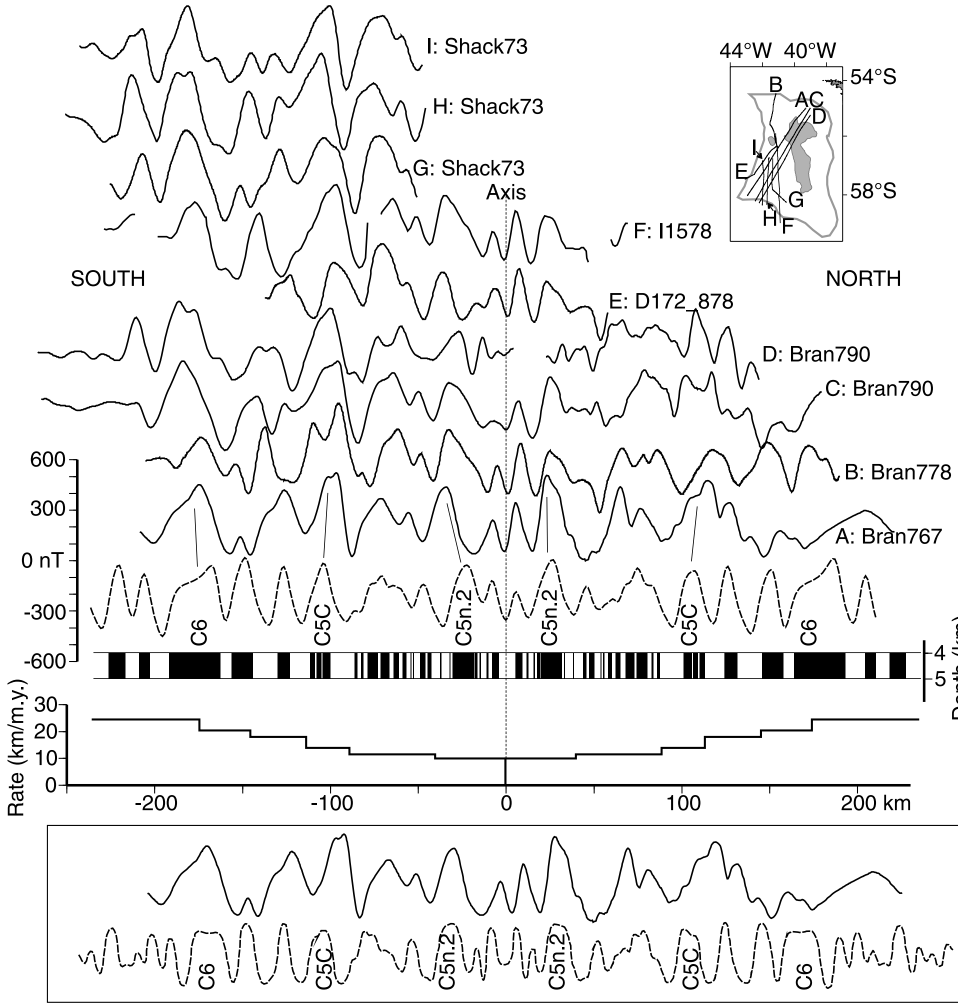

Figure 3. Miocene seafloor spreading model, generated using spreading rates and an axis similar to those of Hill & Barker (1980). The source is a flat 1-km-thick layer at 4 km depth. Dashed line: spreading model. The magnetic anomaly profiles sample the western corridor and parts of the bathymetric swell running north–south along 40°W (upper inset, with bathymetric highs in grey). Lower inset shows Hill & Barker's (1980) model, redrawn for comparison.

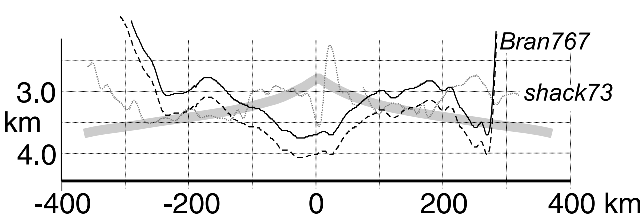

Figure 4. Bathymetry of the west Scotia Sea (profile shack73, WNW on left, dashed dark grey line) compared to that of the central Scotia Sea (profile Bran767, south on left, dashed line, dotted after isostatic correction for a constant-thickness sediment load of 1 km) and to a standard thermal subsidence curve appropriate to the two systems' spreading rates (thick light grey line).

(Figs 2 and 4). Instead, the characteristic relief of the central Scotia Sea is flat, and this flatness cannot be attributed to the sediment cover, which appears to be of relatively constant thickness. This regional flatness of the central Scotia Sea suggests rather that it does not host a recently extinct spreading centre, but at some point in its lifetime has reached a state of thermal equilibrium, with heat supply at the base of the oceanic plate equal to heat lost to the ocean above it. In spite of the regional shallowness, then, the relatively flat abyssal plain in the central Scotia Sea might be interpreted in terms of an age exceeding 80 Ma (Stein & Stein 1992) or a longer and more complicated kinematic history with respect to the underlying mantle and ridge crest at which it was produced (Adam & Vidal 2010) than Hill & Barker's (1980) scheme would allow.

Thermal subsidence models also reproduce the measurable decrease in heat flow as the oceanic lithosphere cools conductively, so that they can be applied to heat flow estimates to infer the age of the oceanic basement. In the central Scotia Sea, sediment temperature gradient measurements enable basement heat flow estimates at a total of six sites visited during two campaigns in March 1978 and November 1995, distributed over a north–south distance of 300 km (Zlotnicki *et al.* 1980; Barker & Lawver 2000). All of these sites returned heat flow estimates that are less than what might be expected over conductively cooling Miocene basement, five of them at 30 per cent or less of expected values. Despite this, Barker & Lawver (2000) asserted the Miocene age of the basement, and suggested that the much-reduced sedimentary temperature gradient was the result of two processes. The first and more important of these, according to those authors, is cooling in near seabed sediments by bottom water temperature changes prior to the two heat probe campaigns. They modelled the effect of a 15-year-long bottom water temperature reduction of 0.7°C, and showed that, a year and a half after returning the bottom water to 0°C, the sediment temperature profile would resemble that in sediments cooling solely by conduction over Mesozoic basement. However, there is no direct evidence for either the magnitude or the timing of this proposed bottom water temperature change. Instead, current-meter moorings and repeat CTD profiles reveal subannual and interannual bottom-water temperature fluctuations of less than 0.2°C, and furthermore that the interannual trend prior to 1995 was one of bottom water warming, rather than cooling (Meredith *et al.* 2008). Using these figures as a guide, and assuming the underlying basement rocks to be of Mesozoic age, Fig. 5 shows that the measured sediment temperature gradients can be modelled convincingly as the signatures of a Mesozoic oceanic geotherm that is responding to known seasonal and interannual bottom-water temperature changes in the central Scotia Sea.

The second process suggested to affect the heat probe records in the central Scotia Sea is circulation of cool seawater through volcanic basement rocks under the sediment pile (*cf*. Davis *et al.* 1999). In a Miocene central Scotia Sea, this process would need to be to particularly important on a regional scale to mitigate the effect of what is now known to be the more modest bottom water temperature variability than Barker & Lawver (2000) invoked; a 0.2°C temperature increase would reduce estimated basement heat flow to just 70 per cent of Miocene geothermal values after 6 months, instead of the change to 40 per cent that their larger but unsupported change achieves. Basement flow requires basement outcrops to act as intakes for the circulating seawater, as well as sites for it to be vented elsewhere, in the process transferring heat from the basement into the overlying ocean. Basement does crop out in small circular regions at distances 20–35 km from the sites of heat flow measurements in the central Scotia Sea (Barker & Lawver 2000), but it is questionable whether such outcrops are close enough or large enough for lateral fluid flow to achieve heat flow reductions like those claimed. For example, even vigorous lateral seawater flow via intake at a circular basement outcrop on 3.5-Myr-old seafloor near the Juan da Fuca Ridge appears only to depress heat flow out to distances of 0.5 km from the outcrop (Fisher *et al.* 2003). Similarly, it would seem that basement pore water flow is of negligible importance in the neighbouring Powell and Jane basins, which share similar basement outcrop and sediment cover characteristics with the central Scotia Sea, and yet have returned heat-flow estimates that are modelled well by conductive cooling from a well-constrained Miocene basement (Lawver *et al.* 1991, 1994). In summary, the heat-probe data from the central Scotia Sea are most easily and simply modelled by a conductive cooling model with underlying lithosphere of Mesozoic age.

2.2 Mechanical state of the lithosphere

A prominent free-air gravity anomaly low of over 80 mGal occurs at the northern edge of the central Scotia Sea, immediately south of the South Georgia continental block (Figs 1, 2 and 5). Fig. 2 shows relatively subdued bathymetry in the area; the gravity anomaly is the consequence of a pile of sediments filling a basin on the northern edge of the central Scotia Sea. The basin has been the site of two large recorded earthquakes whose focal mechanisms suggest the South Georgia microcontinent is being thrust onto the central Scotia Sea floor, and may have been doing so since ∼6 Ma when spreading ceased at the West Scotia Ridge (Thomas *et al.* 2003; Eagles *et al.* 2005).

If, as this setting suggests, the basin is a flexural feature caused by thrust loading at the northern edge of the oceanic Scotia plate, it is possible to calculate the plate's flexural rigidity and, hence, elastic

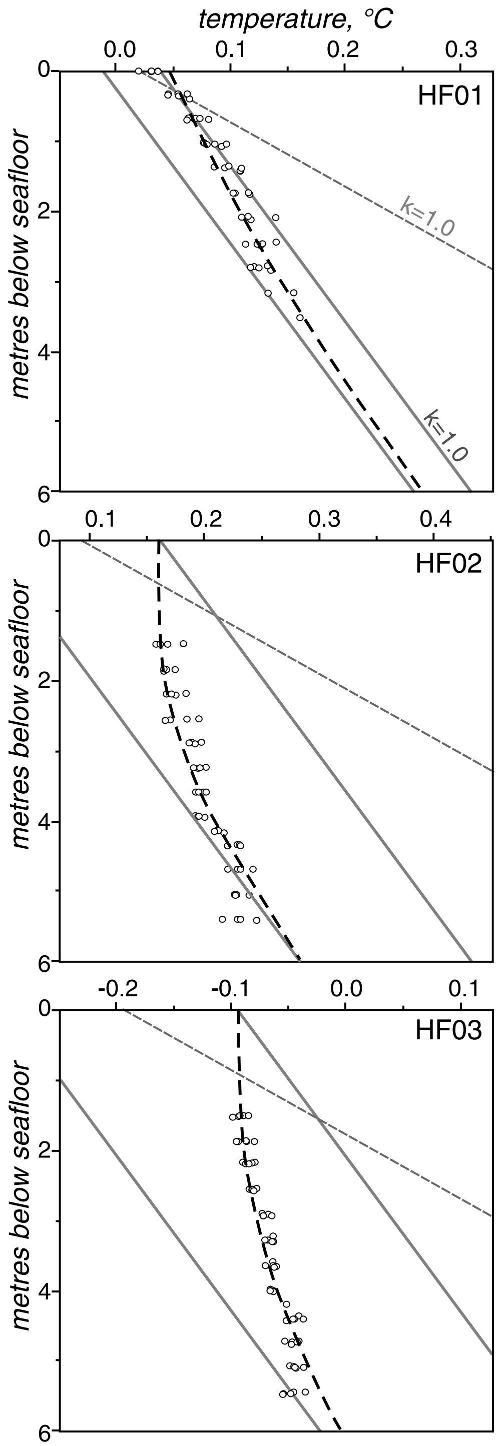

Figure 5. Measured (diamonds; Barker & Lawver 2000) and modelled central Scotia Sea sediment temperature gradients. The variable seabed temperature offsets are estimated based on Barker & Lawver's (2000) fig. 2, and are consistent with short-timescale bottom water temperature variations that they attribute to mesoscale eddy activity. Grey lines: expected equilibrated gradients above basement with 'Mesozoic' heat flow of 45 mW m^{-2}, for sediment thermal conductivity (k) of 1.0 before and after instantaneous bottom water temperature changes of 0.05°C (top), 0.15°C (middle) and 0.2°C (bottom). Dashed black lines: sediment temperature gradients at 6 months (top and middle) and 1 year (bottom) after each of these increases. Grey dashed lines: expected equilibrated gradient in sediments above basement with Miocene heat flow of 110 mW m^{-2}, for sediment thermal conductivity of 1.0.

thickness by starting with the width of the basin (e.g. Beaumont 1981). This elastic thickness can then be interpreted in terms of the thermal structure of the plate, and hence its age. A large source of uncertainty in this kind of analysis is that involved in estimating the width of the basin from the gravity anomaly. The anomaly south of South Georgia varies in width over a range of 70–90 km. Taking typical values for Young's modulus of 80 GPa, Poisson's ratio of 0.25, and basin fill density of 2100 kg m^{-3}, basins with this range of widths would form on elastic plates with thicknesses in the range 12–16 km, which is comparable to many of Bry & White's (2007) standard deviations for model misfits to multiple gravity profiles of similar length to those used here. Plotting this thickness range in with the array of estimates published by Calmant *et al.* (1990) might suggest the central Scotia Plate was anything between 40 and 85 Myr old at the time of its loading, but modern elastic thickness data sets show far more scatter and so caution against such precision in interpretations of lithospheric ages (e.g. Watts 2001; Watts *et al.* 2006; Bry & White 2007). Nonetheless, Bry & White (2007) report a tendency for elastic thickness estimates from short free-air gravity anomaly profiles at trenches to be less than 10 km in oceanic lithosphere aged 20 Ma or younger, and for them to increase towards 15 km as the age of the lithosphere increases towards 150 Ma, suggesting that it may be appropriate at least to view the central Scotia Sea as having formed before the Miocene.

Fig. 6 illustrates an alternative basis for this interpretation by comparing the central Scotia gravity anomaly profile to two nearby flexural profiles for oceanic lithosphere of known ages. The profile crossing the South Sandwich Trench near 55°S illustrates the bending of 75-Myr-old oceanic lithosphere caused by the overriding Sandwich Plate. The other profile illustrates bending of 23-Myr-old oceanic lithosphere in the west Scotia Sea as a result of its loading by Burdwood Bank, a part of the same ridge that causes flexure of the plate in the central Scotia Sea. The wavelengths of gravity anomalies related to the basin and the forebulge in a flexural profile depend on the lithospheric rigidity and, hence, its thickness and age. The wavelengths of the central Scotia and Sandwich Trench flexural profiles are comparable, suggesting that the two plates have similar elastic thicknesses and, thus, similar loading ages. The dissimilarity of the central and west Scotia Sea profiles speaks, on the other hand, against an interpretation in which the two are of similar age. Although the basis of this analysis ensures its conclusion is blunt, that conclusion does nonetheless reinforce the suggestions from other observations that the seafloor in the central Scotia Sea did not form in Miocene times, but instead is likely to be considerably older.

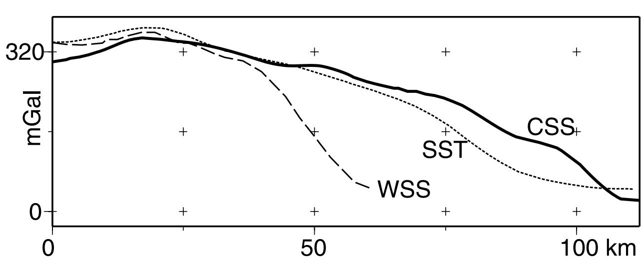

Figure 6. Free-air gravity profiles (sampled from a recent version of the data of Sandwell & Smith (1997)) over flexural basins in the northern central Scotia Sea (CSS), northern west Scotia Sea (WSS) and over the northern South Sandwich Trench (SST). The central and west Scotia profiles are exaggerated vertically to ease comparison of the wavelengths.

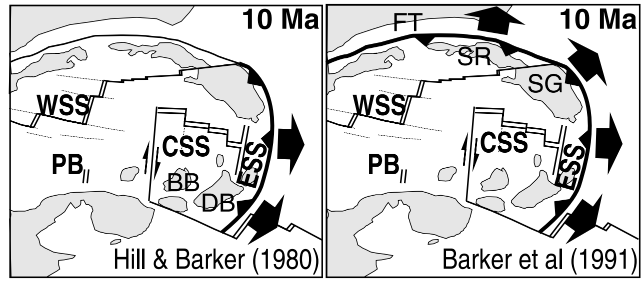

Figure 7. Two conceptions of the central Scotia Sea as a Miocene backarc basin. Thick black arrows denote direction of trench rollback. BB, Bruce Bank; CSS, central Scotia Sea; DB, Discovery Bank; ESS, East Scotia Sea; FT, Falkland Trough; SG, South Georgia microcontinent; SR, Shag Rocks; WSS, west Scotia Sea. Modified from Hill & Barker (1980) and Barker *et al.* (1991).

2.3 Miocene backarc basin kinematics and dynamics

Hill & Barker (1980) proposed that the north–south directed seafloor spreading in the central Scotia Sea occurred in a backarc basin that opened to separate South Georgia from Bruce Bank (Fig. 7). Such a process would be an unusual feature in the post-Eocene plate kinematic setting of the Scotia Sea, which is framed by WNW–ESE relative plate motions between South America and Antarctica (Barker 1995; Livermore *et al.* 2005; Eagles *et al.* 2005, 2006). To account for this, Hill & Barker (1980) related the basin to a southerly component of trench rollback at a northwest-dipping subduction zone at Discovery Bank (Fig. 7), which is now inactive. The presence and action of this subduction zone was well established by that time, but a small backarc basin (Jane Basin; Barker *et al.* 1984; Lawver *et al.* 1994) was subsequently identified as having opened behind it.

The central Scotia Sea later came to be pictured as coupled to rollback of a buried trench formed at a south-directed subduction zone consuming South American oceanic lithosphere in the eastern Falkland Trough (e.g. Fig. 7; Barker *et al.* 1991; Livermore *et al.* 2007). Surface wave tomographic models are of limited resolution in the region, but along with seismic refraction and reflection data they show no sign of a subducted slab beneath the eastern parts of the trough (Ludwig & Rabinowitz 1982; Vuan *et al.* 2000). Nearer the surface, the only evidence for convergence comes in the form of minor reverse faults and folds seen in seismic reflection and sidescan sonar data north of Shag Rocks, and there is no such evidence at all for Cenozoic convergence north of South Georgia (Ludwig & Rabinowitz 1982; Cunningham *et al.* 1998). Furthermore, there is no outcrop evidence on either South Georgia or Shag Rocks for Miocene volcanism related to this proposed subduction. Although it is faintly possible that glaciation has removed or concealed Miocene-aged calc-alkaline volcanic rocks from South Georgia, and that subduction-related features remain to be discovered offshore, the simplest conclusion to draw from present data is that there was no Miocene subduction zone in the eastern Falkland Trough, and thus no reason for there to have been a Miocene backarc basin in the central Scotia Sea.

2.4 Plate boundaries and triple junctions

A Miocene spreading centre in the central Scotia Sea ought to have left interpretable evidence for its action and connections to neighbouring plate boundaries. But it is very difficult to make such interpretations in any convincing way from available data. First, assuming from the lack of evidence of subduction of Scotia Sea floor beneath the north or south Scotia ridges that the fossil spreading system is intact (in spite of the lack of any evidence for a fossil ridge in the basin), magnetic anomalies at similar distances north and south of the proposed fossil spreading centre ought to have similar lengths and complementary strike rotations with respect to the fossil ridge. Instead of this, anomalies maintain an easterly strike and lengthen southwards of the proposed fossil ridge, but shorten and rotate clockwise by 30° northwards of it (Fig. 2).

Secondly, Livermore *et al.* (1994) noted that the slow spreading rate implies the presence of an extinct median valley between prominent rift shoulders. Like the median valley of the neighbouring extinct slow-spreading West Scotia Ridge, such a feature should be readily identifiable in bathymetric and gravity data, but there is no evidence for one either at or away from Hill & Barker's (1980) fossil axis (Figs 2 and 4).

Finally, the east-striking magnetic lineations terminate close to anomaly C8 in the west Scotia Sea (Fig. 2). This means that the boundary between simultaneously active west and central Scotia seas must have been a north–south trending transform fault connecting the 'central Scotia' spreading centre to a triple junction on the south or north Scotia ridge (e.g. Barker & Hill 1981; Fig. 7). Gravity, bathymetric and seismic reflection data (BAS, unpublished data) show neither this fault nor any of the kinds of features, such as offset basins or transpressional ridges, that it might be expected to have given rise to. Barker & Hill (1981) portrayed the eastern boundary of the central Scotia Sea as having formed by the action of a ridge–ridge–ridge triple junction between the central Scotia and East Scotia spreading centres (Fig. 7). Such a triple junction need not have left any particularly strong or characteristic gravimetric or bathymetric signal, but it should be evident in the form of increasingly larger magnetic bights for successively younger magnetic anomalies. In a Miocene central Scotia Sea, the largest bight should have formed for the youngest magnetic anomaly, C4, and be easiest to identify. Fig. 2 shows that although C4 is well defined in the East Scotia Sea, it does not connect with the proposed C4 identifications in the central Scotia Sea to form a magnetic bight.

3 A MESOZOIC CENTRAL SCOTIA SEA

There are no data to support the existence of an extinct ridge crest or triple junction traces that would be associated with a two-sided Miocene spreading system in the central Scotia Sea. Instead, the available data almost unanimously invite the conclusion that the central Scotia Sea is at least 80 Myr old and probably hosts part of just one flank of an oceanic spreading system. In the following, I re-examine the region by taking this conclusion as a starting point and propose a scenario for its development as a consequence of processes acting since the onset of Gondwana breakup.

3.1 Sheared and extended margins and M-series anomalies of the Weddell and central Scotia seas

Just how old might the seafloor in the central Scotia Sea actually be, and where is the other half of the ocean that it formed part of? We have seen that the majority of the marine geophysical constraints favour an age in excess of 80 Ma. Outcrop geology at the margins of the central Scotia Sea, albeit restricted to the islands of South Georgia, may on the other hand be more prescriptive, because the islands bear a sequence of Late Jurassic and Cretaceous turbidites (Storey & Macdonald 1984). At first glance, it seems straightforward to interpret these rocks as having accumulated on the continental margin of the central Scotia Sea, which can therefore be assigned a maximum Late Jurassic age. However, it should be noted that an influential alternative interpretation states the island formed part of the floor of a subduction-related marginal basin near Cape Horn, far to the west of the central Scotia Sea (Dalziel & Elliot 1971; Dalziel *et al.* 1975; Storey & Macdonald 1984; Barker *et al.* 1991). After a detailed examination of that interpretation, Eagles (2010) shows nonetheless that the more economical interpretation of the tectonostratigraphy of South Georgia is one in which it really did occupy part of the oceanwards, transitional, portion of an extended continental margin along the Falkland Plateau, and its neighbouring ocean, that formed as Gondwana broke up in Jurassic times.

In Gondwana reconstructions, the conjugate elements to this system in South Georgia and the central Scotia Sea lie in the southern Weddell Sea. The Weddell Sea is blanketed by up to 4 km of sediments in the southwest, obscuring its magnetic record, and by around 1–1.5 km further east, of which a fairly constant 0.5–1.0 km accumulated following the onset of Antarctic glaciation (Rogenhagen *et al.* 2004). Although the southern margins of the Weddell Sea bear no rock outcrops for comparison with those of South Georgia, seismic data do allow the interpretation of a similar mixed history of shearing and stretching to that recorded on the islands (Kristoffersen & Haugland 1986). Offshore, Jokat *et al.* (2003) identified a sequence of magnetic anomalies that starts with chron M19 (146 Ma) in the eastern Weddell Sea, and showed that the sediment-obscured sequence further west should start with even older anomalies. Eagles & Vaughan (2009) showed that these magnetic isochrons and the attendant fracture zones in the Weddell Sea sequence are closely comparable to those predicted independently for the separation of two plates, South America and Antarctica, by the construction of a three plate circuit with data from the SW Indian and south Atlantic oceans. A simple and testable conclusion to draw from all this is that the magnetic anomaly sequence in the central Scotia Sea formed following the breakup of Gondwana by the separation of South America from Antarctica.

The test can be undertaken by comparing the central Scotia magnetic anomaly sequence to a model made using the slow (10–18 km Myr^{-1}) spreading rates constrained by Eagles & Vaughan's (2009) plate circuit rotations. Affirmatively, Fig. 8 shows a set of correlations for southward-younging anomalies M1n to M25 (125–155 Ma) to the profiles across the western corridor, and suggests the shorter profiles across the central corridor might be offset by a small amount from it. As with the Neogene models of Hill & Barker (1980), the resemblance between the data and the model is reasonable, but the correlations are by no means unique; an alternative sequence starting around M10 is also shown. The magnetic data are thus consistent with, but not diagnostic of, a Mesozoic age. However, together with the very low heat flow values, the ~15 km elastic thickness, the flat abyssal plain, the apparently one-sided or asymmetrical nature of the magnetic anomaly sequence, and the absence of evidence for active post-Eocene plate boundaries in the central Scotia Sea, the magnetic model forms a set of indications that are very convincing evidence for the existence of a region of Mesozoic seafloor in the central Scotia Sea.

3.2 Formation and regional context of Mesozoic seafloor in the central Scotia Sea

Can such a sequence of magnetic anomalies really have appeared by the accretion of oceanic crust to the South American Plate at a South American–Antarctic ridge in the Weddell Sea in Jurassic

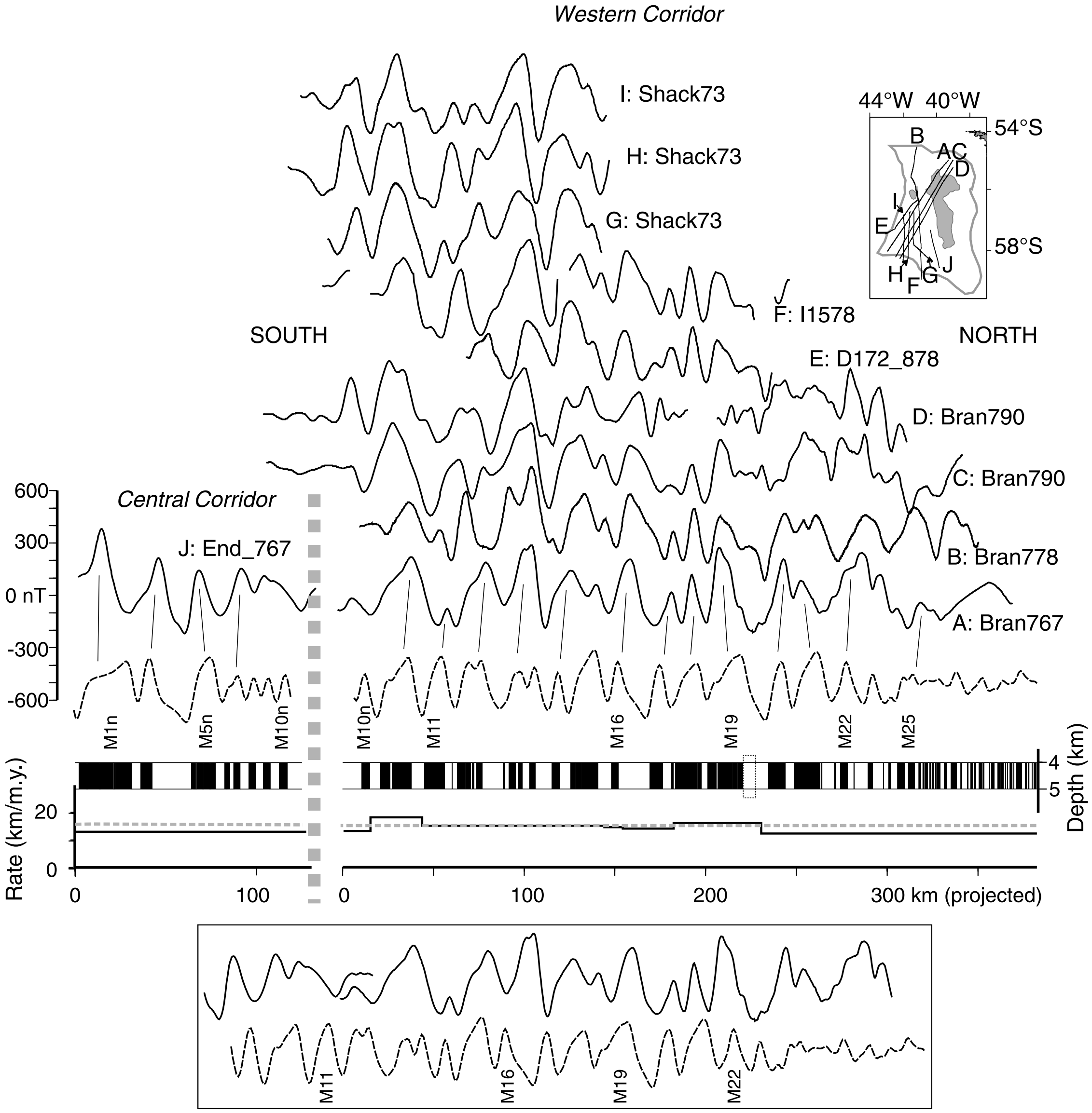

Figure 8. Jurassic–Cretaceous models of spreading anomalies in the central Scotia Sea. The source is a flat 1-km-thick layer at 4 km depth. Dashed lines: spreading model. The magnetic anomaly profiles sample the whole western corridor and part of the central corridor. A dotted line shows a constant spreading rate of 15 km Myr^{-1}, as suggested by calculations of South America–Antarctica relative motions. Inset, bottom, alternative correlations to the same model.

and early Cretaceous times? To illustrate that it can have, I use reconstructions of major plate motions (Figs 9 and 10). Although many of the details of the region's history remain the subject of speculation, in part because of the subduction of seafloor that might have recorded them, and in part because of its proximity to the complexities of the long-lived active Pacific margin of Gondwana, the central idea here can be tested in this way given the recent increase in confidence in knowledge of the motions of the major plates. Fig. 9 shows magnetic and gravity data of the Weddell and central Scotia seas reconstructed to chron C34y, along with a set of synthetic flowlines and isochrons, made using Eagles & Vaughan's (2009) rotations. In addition to fitting the Weddell Sea data well, the flowlines also predict fracture zone azimuths and the spacing of M-series anomalies accurately in the central Scotia Sea, with misfits of <20 km to the set of correlations beginning at anomaly C34o in Fig. 8, suggesting that it might be favoured over the set that begins with M10.

The FIT reconstruction in Fig. 10(a) illustrates an extensional or transtensional plate boundary zone crossing southwestern Gondwana towards its active margin with the paleo-Pacific ocean. By analogy with the still-active analogue through Burma and the Gulf of Thailand (Hall 2009), this boundary may have been broad and unstable, and early in its history may have formed basins floored by transitional and/or oceanic crust over the Falkland Plateau and Weddell Sea Embayment. In and around South Georgia, the earliest phase of relative movements between East Gondwana, bearing East Antarctica, and West Gondwana, bearing South America, in this zone is recorded in the Drygalski Fjord Complex. These parts of the margin went on to form the southernmost Weddell Sea and, I suggest, the northernmost central Scotia Sea, which may partly therefore be situated on transitional or highly extended continental, rather than oceanic, crust (Fig. 10b). Shortly after this, at around chron M19 (146 Ma; Fig. 9), a change in magnetic anomaly orientations in the central Scotia Sea shows that transcurrent motions

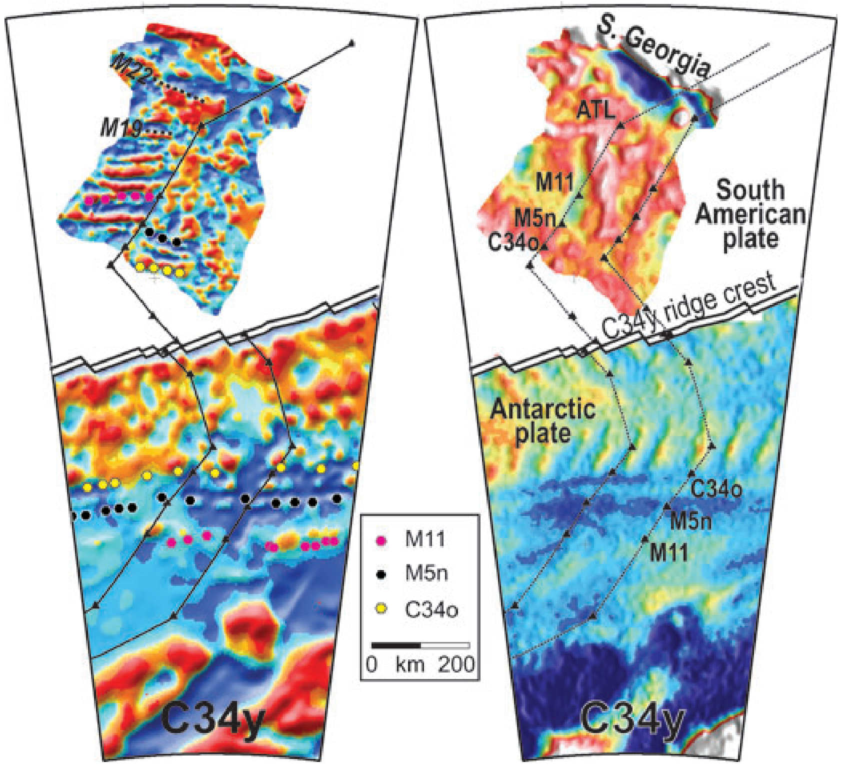

Figure 9. Left panel: reconstruction of magnetic anomalies (Golynsky *et al.* 2001; Eagles *et al.* 2005) for C34y, in East Antarctica reference frame. The central Scotia data have been upward continued by 1.0 km in view of the greater depth to source in the Weddell Sea. Black lines are synthetic flowlines that describe movements until chron C34y (Eagles & Vaughan 2009). Model isochrons on the flowlines are shown by triangles, for comparison with the positions of magnetic isochrons identified in the Weddell and central Scotia seas (hexagons; König & Jokat 2006). Some other isochrons are shown with black dotted lines. ATL: the rotation dating from the time of first divergence in the South Atlantic. Right panel: same reconstruction with gravity data, showing the fit to fracture zones in the Weddell Sea and central Scotia Sea. Before rotation as part of the South American Plate, the central Scotia Sea data are restored as part of a central Scotia plate to the position they would have occupied prior to spreading in the west Scotia Sea (Eagles *et al.* 2005).

gave way to divergence. This event has been shown to be a result of the onset of continental extension elsewhere in the plate circuit between the South American and African plates (Eagles 2007; Eagles & Vaughan 2009). In South Georgia, this period saw a northward jump or refocussing of the plate boundary to near the southern margin of the island, and is recorded in the thermal history of the Drygalski Fjord Complex (Tanner & Rex 1979). Away from South Georgia, the onset of divergence added to the ocean floor already existing in the central Scotia and Weddell seas, and possibly also saw the onset of spreading in the Rocas Verdes Basin of Tierra del Fuego (Calderón *et al.* 2007).

Accretion of the floor of what was to become the central Scotia Sea was finished by chron C34o (∼125 Ma; Figs 8 and 10c). This is the beginning of a period that is the least well understood for major plate motions and regional tectonics, because of the absence of dateable magnetic anomalies in the Cretaceous normal polarity superchron, which lasted until 84 Ma. We know that the time leading up to the beginning of the superchron also saw an abrupt drop in the seafloor spreading rates between South America and Antarctica, giving rise to a new pattern of ridge segmentation in the Weddell Sea (Livermore & Hunter 1996), and possibly the cessation of spreading in the Rocas Verdes Basin (Stern & DeWit 2003; Calderón *et al.* 2007). We also know that the middle of the superchron saw the compression of the Rocas Verdes basin first becoming evident in or around Cenomanian times (Dalziel 1981), and the cessation of relative movements between the Antarctic Peninsula and East Antarctica in the 107–103 Ma Palmer Land event (DiVenere *et al.* 1996; Vaughan *et al.* 2002). Like the Palmer Land event, the evolution of the Rocas Verdes Basin is often linked to terrane collisions and translations occurring at the Pacific margin of Gondwana, but the spreading rate change and the strong curvature of fracture zones in the Weddell Sea shows that large changes were occurring in the plate kinematics of Gondwana breakup too, and these may have played a role in the mid Cretaceous development of southernmost South America and the Antarctic Peninsula.

3.3 Subduction of South American seafloor but survival of part of it in the central Scotia Sea

By the present day, much of the seafloor that accreted to the South American Plate at South American–Antarctic Ridge has been destroyed by subduction at the South Sandwich Trench. The central Scotia Sea, if it did form at an ancestor of that ridge in the Weddell Sea, must have escaped this subduction somehow to become preserved behind the trench. To understand how this might have occurred, it is necessary to explain the onset of subduction and track the kinematic history of the trench and plate boundaries that connected to it. The post-C34o reconstructions in Figs 10(d)–(f) constitute the framework for doing this; together with those images the following text explains how the inception of subduction to the south of the central Scotia Sea, and the subsequent northward propagation of the trench through seafloor to the east of it, combined to isolate a fragment of Mesozoic seafloor within the body of the

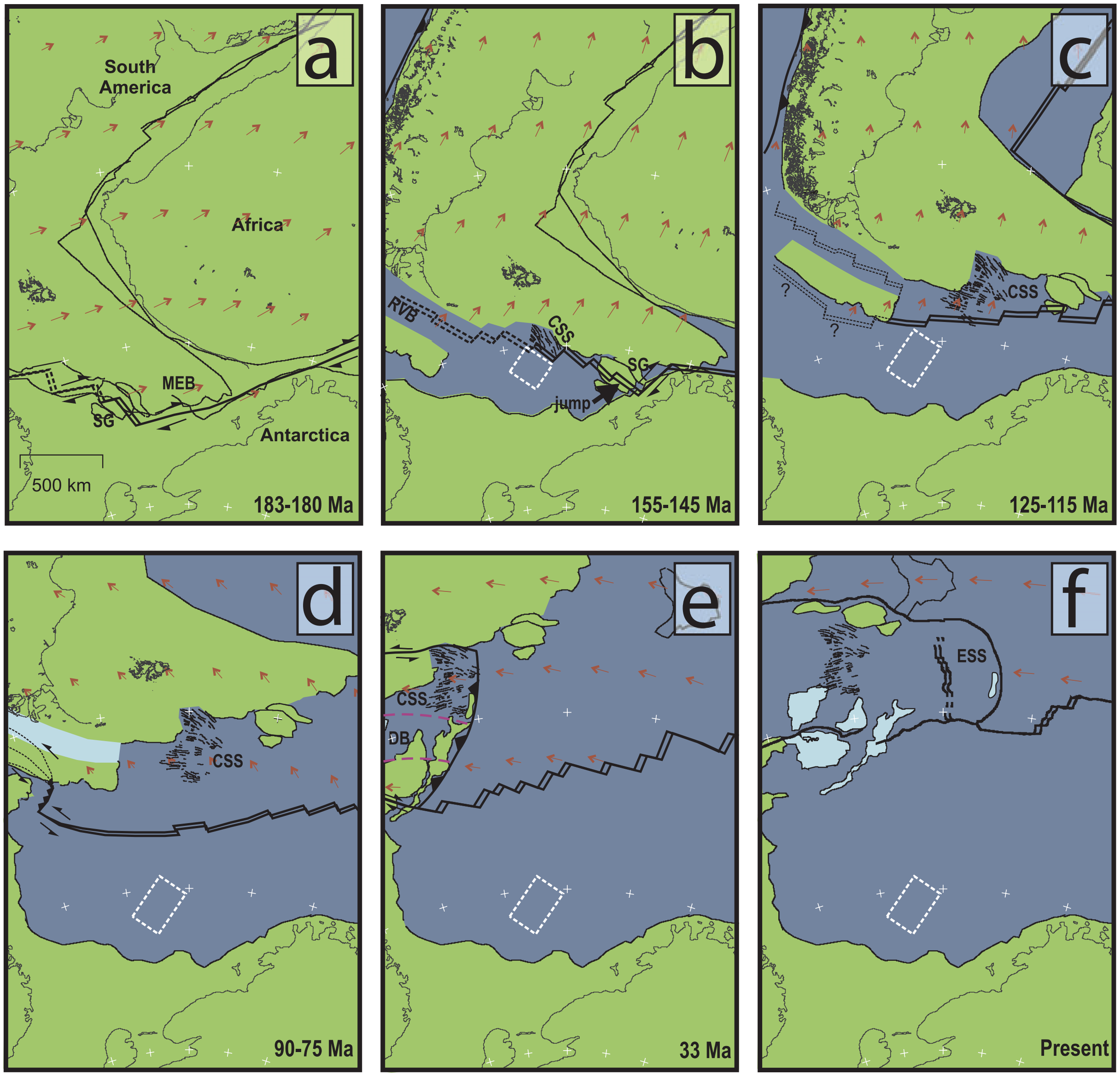

Figure 10. Reconstruction maps, made in a fixed East Antarctica reference frame. Red arrows: Instantaneous motion azimuths on West Gondwana (before South Atlantic opening) and South America (afterwards). Double lines: Divergent plate boundaries, based on magnetic lineations. Lines with barbs: Collision or subduction zones. Single lines: transform zones. Dashed purple lines: abandoned transfer zones. Green: emergent or shallow marine. Light blue: intermediate or deeper water on continental or arc basement. White box: approximate position of Weddell Sea conjugates to proposed Mesozoic anomalies of the central Scotia Sea. CSS, central Scotia Sea; DB, Dove Basin; ESS, East Scotia Sea; RVB, Rocas Verdes Basin. The central Scotia Sea, South Georgia and the North Scotia Ridge have been restored to their pre-west Scotia spreading positions, and the width of the Rocas Verdes Basin shown is based on the associated shortening between the Magallanes and South America plates (Eagles *et al.* 2005). This width is thus a minimum estimate from plate kinematic studies, as no account has been made for any cross-basinal component of mid-Cretaceous shortening that occurred in the oblique collision between South America and the Antarctic Peninsula.

Scotia Plate. Although the details of some of these processes are only sketchily understood, the basic framework is based on strong enough evidence to make the overall scenario a plausible one.

The Cretaceous magnetic quiet zone in the Weddell Sea (Golynsky *et al.* 2001) narrows towards the eastern margin of the Antarctic Peninsula, showing that a mid ocean ridge would have abutted that margin somewhere near 65°S at 100 Ma. With the peninsula fixed with respect to East Antarctica at this time, the margin north of that point must have hosted a branch of the South America–Antarctica plate boundary connecting the mid ocean ridge to the oblique collisional plate boundary zone in southern Tierra del Fuego (Kohn *et al.* 1993). In its earliest and most simple form, a transform fault zone at the eastern margin of the peninsula would have trended NE, parallel both to the direction of relative plate motion at the time and the margin itself, and would have been active in an overall left-lateral sense. Del Valle & Miller (2001) give structural evidence for mid Cretaceous faults consistent with such a setting in northern Graham Land. Over the next 15 Myr, the South America–Antarctica relative motion vector rotated from NE–SW, via N–S, to NW–SE, requiring the transform zone at the eastern edge of the peninsula to evolve into a continent–ocean collision zone (Fig. 10d). This evolution may be recorded in the rapid uplift of the northern parts of the peninsula around 87 Ma (Hervé *et al.* 1996). With subsequent ongoing NW–SE directed relative motion, South American oceanic lithosphere at this collision zone could eventually have been thrust down to such depths that west-directed subduction initiated. This subduction is well known from marine geophysical evidence south and east of the South Orkney Microcontinent (e.g. Barker *et al.* 1984); its initiation may be dated to early Eocene times on the basis of dredge-haul evidence from Powell Basin (Barber *et al.* 1991), or to Maastrichtian–Paleogene times (∼68–60 Ma) based on indications of calc-alkaline volcanism to the east of the James Ross Basin (Elliot 1997; Lomas & Dingle 1999).

Today, west-directed subduction in the region continues at the South Sandwich Trench, far to the northeast of the site of its inception. The trench appears to have reached this location by eastwards retreat into the interior of the South American Plate in combination with northward propagation or migration of both its northern and southern tips. Northward migration of the southern tip is well understood to have occurred in response to collisions of successively more northeasterly segments of the South American–Antarctic Ridge in the Weddell Sea (e.g. Barker *et al.* 1984); with the Antarctic Plate both overriding at the trench and trailing behind the ridge, each ridge-crest—trench collision meant subduction ceased along the collided portion of the trench. The reason for movement of the trench's northern tip is harder to understand, but can be speculated on given the knowledge that its current position is determined by tearing of the South American lithosphere along the east-trending southern edge of the Northeast Georgia Rise, a region of thickened and presumably more buoyant oceanic crust (Forsyth 1975). After the initiation of subduction, similar tearing would have occurred along the southern margin of continental Tierra del Fuego and its offshore continuation through to Bruce Bank (Eagles *et al.* 2006; Fig. 10d). Upon reaching the eastern end of this constraint, it is possible that tearing then proceeded along the next-best available crustal asperity: the thinned crust of a NNE-trending Mesozoic fracture zone on the South American plate. Whatever the cause of its migration, dredge samples from the present South Sandwich fore-arc suggest that the northern end of the trench was present at least as far north as 57.5°S by 35 Ma (Livermore *et al.* 1994; Fig. 10e).

Today, the northern and southern ends of the trench are connected to the global plate circuit, at the Pacific margins of Antarctica and South America, via east-trending transfer zones along the North Scotia Ridge—Magallanes–Fagnano fault zone and the South Scotia Ridge and Shackleton Fracture Zone (Fig. 10f). Given the history of northward trench migration, it is necessary to envisage earlier equivalents to the southern transfer zone through the Powell Basin, and to the northern one through the South Scotia Ridge, then later the northern margin of Dove Basin, and eventually the North Scotia Ridge (Fig. 10e). Fig. 10 suggests how these early northern transfer zones would have lengthened into the region south of the central Scotia Sea, excising it by parts from the subducting part of the South America Plate by chron C13. This, together with northwards migration of the South Sandwich Trench into the region to the east of the central Scotia, would eventually have served to isolate the Mesozoic seafloor fragment in the central Scotia Sea as part of the Scotia Plate.

4 ANOMALOUSLY SHALLOW BATHYMETRY

We have seen that the regional seafloor depth (but not its flatness) is inconsistent with a Jurassic to early Cretaceous age for the central Scotia Sea (Figs 2 and 4). Fig. 11 quantifies this inconsistency as deviations of the regional bathymetry (Smith & Sandwell 1997) from a surface predicted using the seafloor age as gridded from the magnetic anomaly identifications given here and by Eagles *et al.* (2005) and Stein & Stein's (1992) GDH1 subsidence model. Based on the stated accuracy of Smith & Sandwell's (1997) predicted bathymetry, and on scatter in the data to which the GDH1 model is compared, values exceeding ±500 m are likely to represent significant deviations from normal thermal subsidence, and hence are shaded in Fig. 11. The bathymetric swell along 40°W is shallower than any normal oceanic depths; here, spreading anomalies are not as clear as further west, perhaps having been overprinted by the signature of volcanism ancestral to that at the South Sandwich arc.

Sediment distribution over the central Scotia Sea is not known in detail, but available seismic profiles show typical thicknesses

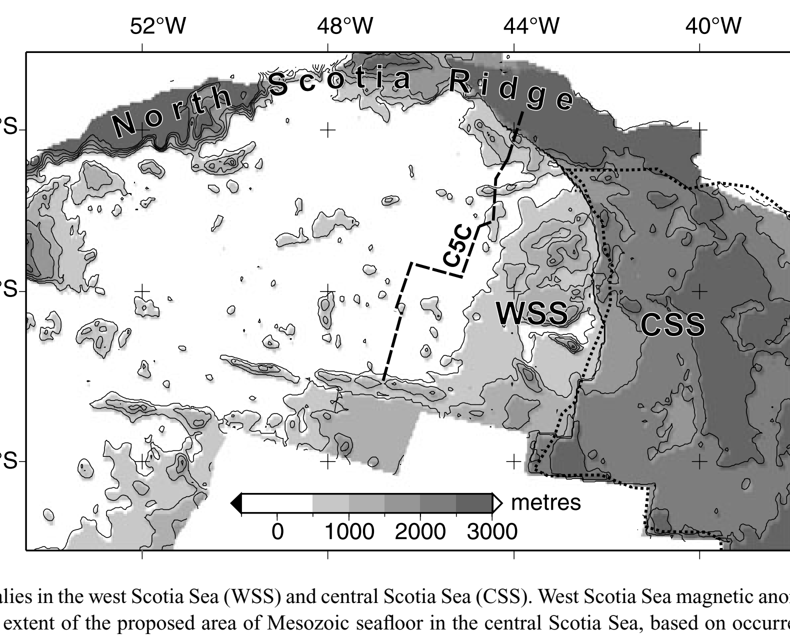

Figure 11. Residual depth anomalies in the west Scotia Sea (WSS) and central Scotia Sea (CSS). West Scotia Sea magnetic anomaly data are from Eagles *et al.* (2005). Dotted outline shows the extent of the proposed area of Mesozoic seafloor in the central Scotia Sea, based on occurrences of east-striking magnetic anomalies. Contour interval is 500 m.

of 800–1500 m covering wide areas, for which a simple isostatic correction would reduce the anomalies of Fig. 11 by 480–900 m (Crough 1983). Once this has been taken into account, areas with strong east–west magnetic lineations can be ascribed basement depth anomalies of 0.6–2.0 km. Similar anomalous basement depths occur in neighbouring parts of the west Scotia Sea, where sediment cover is much thinner and patchier, and the age of the lithosphere is well known. The basement depth anomalies of the central Scotia Sea thus appear to be real, and not an artefact of failing to acknowledge a Miocene age. This conclusion can be supported further by analogy to the Vitus Arch in the western Bering Sea. Like the central Scotia Sea, this buried basement arch's relief averages 1–2 km around its surroundings and peaks at around 4 km (Cooper *et al.* 1992). What is more, the Vitus Arch coincides with a ~200 km wide province of possible Mesozoic seafloor spreading anomalies that appears to be trapped in a post-Eocene backarc region (Cooper *et al.* 1976).

The boundary between the central and west Scotia seas is characterized by subdued gravity anomalies, suggesting it has relatively smooth basement topography and that therefore the central Scotia Sea was already anomalously shallow when seafloor spreading started in the west Scotia Sea. Furthermore, the basement depth anomaly in the west Scotia Sea is confined to areas older than chron C5C, near which the normal thermal subsidence trend reverses (Eagles *et al.* 2005; Figs 4 and 11). This time saw the onset of backarc spreading in the East Scotia Sea (Barker 1995; Larter *et al.* 2003), making it possible to interpret the process responsible for the basement depth anomalies there and in the central Scotia Sea as having been related to subduction in the absence of backarc extension. Further study of this enigmatic process must await new information about the deeper crust and mantle beneath the central Scotia Sea, which for the moment is limited to the possibility of a slightly slower mantle seismic velocity than that beneath the central, normally subsided, parts of the west Scotia Sea (Ewing *et al.* 1971).

5 PALEOCEANOGRAPHICAL IMPLICATIONS

Considering the central Scotia Sea as a fragment of Mesozoic ocean trapped within the Scotia Arc removes the largest areal source of uncertainty currently to affect ideas about the timing of establishment of a conduit for the exchange of water between the Pacific and Atlantic oceans. Without backarc basin opening in the central Scotia Sea, the interpretation of the South Georgia microcontinent as having originated on the eastern end of the North Scotia Ridge (Eagles 2010) is strengthened, in the light of which the possibility of it having played a role in blocking the Drake Passage gateway until Miocene times can be disregarded.

With this model, the timing uncertainty in the opening of Drake Passage, and the development of deep water in it, is much more tightly constrained to

(i) uncertainty in the age of small basins in the southeastern Scotia Sea, which have only been dated indirectly from basement depth and with much shorter magnetic anomaly profiles even than those in the central Scotia Sea, to an earliest age of ~42 Ma (Eagles *et al.* 2006), but which may be younger (Galindo-Zaldívar *et al.* 2006),

(ii) the age and nature of the earliest extension at the margins of the west Scotia Sea, which may have begun as early as Paleogene times and formed continental rift basins in Tierra del Fuego (Ghiglione *et al.* 2008),

(iii) the area and degree of Miocene uplift of transverse ridges at the Shackleton Fracture Zone (Livermore *et al.* 2004),

(iv) the timing and extent of possible volcanic arc development along the 40°W swell in the central Scotia Sea before backarc spreading in the East Scotia Sea started around 17 Ma, and

(v) the location and nature of transform-dominated plate boundaries connecting the intra-Scotia plates to the Pacific margin via the South American and/or Antarctic plates at times prior to 26.5 Ma. Possibilities exist to interpret very early narrow conduits through transtensional basins at such boundaries, or alternatively to interpret local barriers to northward-turning currents at transpressional ridges.

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