Deciphering tectonic phases of the Amundsen Sea Embayment shelf, West Antarctica, from a magnetic anomaly grid

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A R T I C L E    I N F O

Article history:
Received 7 September 2011
Received in revised form 15 May 2012
Accepted 20 June 2012
Available online 2 July 2012

Keywords:
Aeromagnetics
Data processing
Pine Island Bay
Goodwana breakup
Bellingshausen Plate
West Antarctic Rift System

A B S T R A C T

The Amundsen Sea Embayment (ASE), with Pine Island Bay (PIB) in the eastern embayment, is a key location to understanding tectonic processes of the Pacific margin of West Antarctica. PIB has for a long time been suggested to contain the crustal boundary between the Thurston Island block and the Marie Byrd Land block. Plate tectonic reconstructions have shown that the initial rifting and breakup of New Zealand from West Antarctica occurred between Chatham Rise and the eastern Marie Byrd Land at the ASE. Recent concepts have discussed the possibility of PIB being the site of one of the eastern branches of the West Antarctic Rift System (WARS). About 30,000 km of aeromagnetic data – collected opportunistically by ship-based helicopter flights – and tracks of ship-borne magnetics were recorded over the ASE shelf during two RV Polarstern expeditions in 2006 and 2010. Grid processing, Euler deconvolution and 2D modelling were applied for the analysis of magnetic anomaly patterns, identification of structural lineaments and characterisation of magnetic source bodies. The grid clearly outlines the boundary zone between the inner shelf with outcropping basement rocks and the sedimentary basins of the middle to outer shelf. Distinct zones of anomaly patterns and lineaments can be associated with at least three tectonic phases from (1) magmatic emplacement zones of Cretaceous rifting and breakup (100–85 Ma), to (2) a southern distributed plate boundary zone of the Bellingshausen Plate (80–61 Ma) and (3) activities of the WARS indicated by NNESSW trending lineaments (55–30 Ma?). The analysis and interpretation are also used for constraining the directions of some of the flow paths of past grounded ice streams across the shelf.

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1. Introduction

The Amundsen Sea Embayment contains one of the largest continental shelves of the Pacific margin of West Antarctica. Pine Island Bay, in the eastern part of the embayment, has for a long time been suggested to contain the crustal boundary between the Thurston Island block and the Marie Byrd Land block (e.g. Dalziel and Elliot, 1982; Grunow et al., 1991; Storey, 1991). Plate tectonic reconstructions suggest that this region was a key area for the initiation of continental breakup, that it was the location of a possible plate boundary, and that it may have been, or still is, an active branch of the West Antarctic Rift System (Dalziel, 2006; Eagles et al., 2004a; Gohl et al., 2007; Jordan et al., 2010; Larter et al., 2002; Wobbe et al., 2012). In spite of all this, little is known about the tectonic evolution and architecture of the embayment from direct study there.

Tectonically induced displacements of the crust are the underlying processes controlling the development of landscapes upon which climate processes play out. This context is of particular importance for reconstructing continental ice sheet evolution. In the Amundsen Sea Embayment, the Pine Island, Thwaites, Smith and Kohler glacier systems are thinning at rapid rates, and some of them have also started to flow at dramatically increased rates (e.g. Rignot et al., 2008; Pritchard et al., 2009). If these glaciers were to drain their catchment area, the volume of ice lost to the ocean could potentially lead to 1.5 m of sea-level rise (Vaughan et al., 2006). Modelling results (Pollard and DeConto, 2009) suggest that the ice sheet in the Amundsen Sea Embayment may have retreated with similar dynamics several times since the Pliocene. Identifying tectonic lineaments and understanding the tectonic architecture of the shelf of the Amundsen Sea Embayment may thus not only help explaining the geodynamic and kinematic processes of continental rifting in this West Antarctic realm, but also provide valuable constraints on flow paths and subglacial substrate of basement for palaeo-ice sheet modellers.

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An extensive ship-borne and helicopter-borne magnetic dataset was collected for the first time in the Amundsen Sea Embayment and Pine Island Bay by the Alfred Wegener Institute during RV Polarstern expeditions ANT-XXIII/4 in 2006 and ANT-XXVI/3 in 2010. The distribution and spacing of the survey tracks are suitable to allow spatial gridding and 3D field analysis for delineating crustal and basement features. In this paper, we describe the workflow from magnetic data acquisition to processing and modelling and put forward a model for the tectonic architecture of the offshore Amundsen Sea Embayment.

2. Geological and geophysical background

Fig. 2 summarizes the main stages in the tectonic development of the Pacific margin of West Antarctica. At least three distinct phases since Late Cretaceous have been discussed in recent literature.

The extension, and subsequent separation, of New Zealand and West Antarctica dominate the tectonic signature of the Amundsen Sea Embayment. This early divergence of the Pacific and Antarctic plates led first to rifting and crustal extension between Chatham Rise and the Amundsen Sea Embayment off easternmost Marie Byrd Land as early as 90 Ma (Eagles et al., 2004a; Larter et al., 2002; Wobbe et al., 2012). Rifting possibly continued within the Great South Basin between the Campbell Plateau and the South Island of New Zealand until its abandonment in favour of a new extensional locus to the south, forming the earliest oceanic crust between Campbell Plateau and Marie Byrd Land by 84–83 Ma.

From about 79 Ma, the Bellingshausen Plate moved independently of the Pacific and Antarctic Ridge on the southern flank of the Pacific–Antarctic Ridge until about 61 Ma, when a major plate reorganisation occurred in the South Pacific (e.g. Eagles et al., 2004a; Larter et al., 2002). This small plate’s western boundary passed through the region of the Marie Byrd Seamounts, north of the Amundsen Sea Embayment. Its eastern transpressional boundary lies along the Bellingshausen Gravity Anomaly lineament in the western Bellingshausen Sea (Eagles et al., 2004a; Gohl et al., 1997). Although a discrete southern plate boundary has been depicted running from the seamounts onto the shelf and mainland, its true nature is poorly known (Eagles et al., 2004a,b) and it may be a more distributed feature.

The location of Pine Island Bay has led several researchers to suggest that it hosts a major crustal boundary between the Marie Byrd Land block to the west and the Thurston Island/Ellsworth Land blocks to the east. These blocks are suggested to have moved with respect to each other during the Late Cretaceous–Antarctic separation and perhaps also in early Mesozoic or Palaeozoic times (e.g. Dalziel and Elliot, 1982; Grunow et al., 1991; Storey, 1991). However, direct evidence of the presence of such a boundary is still missing. Conceptual models also suggest that Pine Island Bay and the eastern Amundsen Sea Embayment hosted basins of the West Antarctic Rift System. Jordan et al. (2010) invert airborne gravity data to reveal an extremely thin crust and low lithospheric rigidity beneath the onshore Pine Island Rift. Müller et al. (2007) and Eagles et al. (2009) considered how, at times between chron 21 and 8 (48–26 Ma), the West Antarctic Rift System east of the Ross Sea operated in either dextral strike-slip or extensional motion through the region to the south and east of the Amundsen Sea Embayment connecting eventually to a Pacific–Phoenix–East Antarctic triple junction via the Byrd Subglacial Basin and the Bentley Subglacial Trench. There are indications for an early West Antarctic Rift System extension in western Marie Byrd Land in the mid-Cretaceous (e.g. McFadden et al., 2010), but its eastern continuation is less well understood. It is possible that in the north–south striking zone of thinned crust in Pine Island Bay was an eastern arm of this early manifestation of the West Antarctic Rift System (Dalziel, 2006; Ferraccioli et al., 2007; Gohl et al., 2007; Jordan et al., 2010).

3. Magnetic surveys and data processing

The Amundsen Sea was the target area for geoscientific, oceanographic and biological studies during the RV Polarstern expeditions ANT-XXIII/4 in 2006 and ANT-XXVI/3 in 2010. Ship-borne magnetic data were continuously recorded with two 3-component fluxgate magnetometer sensors, which are permanently installed on the crow’s nest. One of the two BO-105 helicopters on board was equipped with a caesium-vapour magnetometer sensor towed by a...
30 m long cable. Helicopter-borne magnetic data were collected opportunistically along track lines as far as possible within a pre-determined survey pattern, but subject to the constraints of the vessel’s primary marine geoscientific and oceanographic tasks. Each survey flight covered 350–450 km in 2 to 2.5 h at a nominal 100 m flight height with a sampling interval of 3–5 m. Up to 3 or 4 flights were conducted on days with good flight conditions and no other transport tasks. With this survey approach, the Amundsen Sea Embayment shelf was effectively covered with patches of survey lines that total about 30,000 km of data (Fig. 3). The survey line spacing

![Fig. 2. Plate tectonic reconstruction of the tectonic development in the Amundsen Sea area from 90 to 61 Ma. The plates are rotated according to rotation parameters compiled and derived by Eagles et al. (2004a). The outline of the West Antarctic Rift System (WARS) faults or boundaries is an approximate estimation, because an accurate geometry of the rift system has not been identified yet. Abbreviations are: AP Antarctic Peninsula, ASE Amundsen Sea Embayment, BP Bellingshausen Plate, BT Bounty Trough, BS Bollons Seamount, CP Campbell Plateau, CR Chatham Rise, GSB Great South Basin, MBL Marie Byrd Land, PAC Pacific plate, PHO Phoenix Plate, PIB Pine Island Bay, SNS South Island New Zealand, and WA West Antarctica. Modified from Gohl (2011).](image1)

![Fig. 3. Map with combined helicopter-borne (red lines) and ship-borne (blue lines) magnetic tracks of both RV Polarstern expeditions in 2006 and 2010, which are included in the preparation of the magnetic anomaly grids. The irregular line geometry is due to opportunistic surveying.](image2)
records appeared to be less noisy than the ship-borne data. In general, the helicopter-magnetic data, the 3-component ship-borne magnetic data had to be pre-processed to compensate for interactions between the ship’s ferrous body and the variable geomagnetic field. To facilitate this, a number of figure-eight shaped magnetic compensation loops were completed during the cruises from which compensation coefficients for all headings were calculated. Corrections were applied using these coefficients according to the methods and descriptions of König (2006) and Wobbe et al. (2012). The main contribution to an erroneous shift of the measured magnetic data comes from the ship’s heading (Nogi and Kaminuma, 1999). For instance, a one degree change of the ship’s heading causes about 300 nT of variation in the horizontal components. The vertical component is less affected by change of the ship’s heading but shows noise due to roll and pitch. In order to avoid noise in the total intensity field, the data acquired during and immediately after substantial heading changes were excluded from further processing.

Before merging into the processing stream with the helicopter-magnetic data, the 3-component ship-borne magnetic data had to be pre-processed to compensate for interactions between the ship’s ferrous body and the variable geomagnetic field. To facilitate this, a number of figure-eight shaped magnetic compensation loops were completed during the cruises from which compensation coefficients for all headings were calculated. Corrections were applied using these coefficients according to the methods and descriptions of König (2006) and Wobbe et al. (2012). The main contribution to an erroneous shift of the measured magnetic data comes from the ship’s heading (Nogi and Kaminuma, 1999). For instance, a one degree change of the ship’s heading causes about 300 nT of variation in the horizontal components. The vertical component is less affected by change of the ship’s heading but shows noise due to roll and pitch. In order to avoid noise in the total intensity field, the data acquired during and immediately after substantial heading changes were excluded from further processing.

The total magnetic field data of helicopter-borne caesium-vapour sensor turned out to be rather unaffected by the helicopter’s heading, except for major turns at flight line change. Such data were also removed from further processing. In general, the helicopter-magnetic records appeared to be less noisy than the ship-borne data.

After this initial pre-processing, both ship-borne and helicopter-borne data went through the same principal processing steps (Table 1) using Geosoft Oasis montaj™ software for most applications:

**Removal of International Geomagnetic Reference Field (IGRF):** We removed the IGRFs of 2005 and 2010 from the data of 2006 and 2010, respectively. We used the mean date for each survey.

**Editing and filtering:** Visual editing of obvious erroneous data improved the overall data quality of the datasets of both expeditions. For instance, data affected by tilts in loops of the helicopter’s flight tracks were removed. Large numbers of data were removed in particular from the ship-borne records, owing to noise induced during ship manoeuvres, heading changes and crane movements. A band-pass filter removed spikes and other short-wavelength noise from all datasets. As it was not possible to set up a magnetic base station in the coastal area of the Amundsen Sea Embayment for the duration of both cruises, the data may still be affected by diurnal variations of the geomagnetic field.

**Levelling:** This process is required to the data in order to minimise the differences at cross-points before any gridding is applied. The plane tolerance of crossing survey line points was chosen to be 0.0005° (~50 m). As the helicopter data have shown to provide more reliable results than the ship-borne data, the two datasets were treated separately. The differences at cross-points within the dataset were minimised using an iterative approach to their mean values. We divided the intersection points of the helicopter lines into reliable intersection points (differences vary less then ±10 nT) and less reliable intersections points (all other intersection points). In the first step, these lines were levelled separately: Lines with reliable intersection points were levelled using an ‘iterative levelling’ algorithm. Lines with less reliable intersections were levelled using ‘full levelling’ algorithm. ‘Iterative levelling’ calculates a least-squares trend line through the error data to derive a linear trend error curve, which is then added to the data to be levelled. These steps are repeated for crossing lines until a convergence is reached. ‘Full levelling’ adjusts the values of an input channel by adding a correction defined in an error channel. In a second step, an iterative levelling algorithm was again applied, this time simultaneously to all profiles from both platforms. In

### Table 1

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<th>Magnetic data processing and analysis flow.</th>
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varies between 5 and 25 km and – although a regular flight grid could not be maintained – allows spatial gridding and 3-D field analysis for delineating regional crustal and basement features.

![Cross Point Analysis (Helicopter borne)](image1)

![Cross Point Analysis (Ship borne)](image2)

Fig. 4. Results of the cross-point analysis for the helicopter-borne and ship-borne survey lines before and after levelling.
Fig. 5. Processed magnetic anomaly grids with grid cell sizes of (A) 5 km, (B) 2 km, and (C) 2 km with a 30-km cut-off high-pass filter. Fig. 5c also shows the boundary (white line) between the outcropping basement of the inner shelf in the south and the sediment basin in the north, and an estimated 3-km-depth marker (hashed black line) to which the top of basement dips northward from this boundary.
this step, we levelled the ship-borne data into the levelled framework of helicopter data after splitting the ship-borne data into straight segments and generating an intersection table with points of ship-borne data crossing helicopter survey lines. Intersection points within the ship-borne dataset were ignored. The cross-point histograms and tables (Fig. 4) show the large reduction in intersection misfits brought about by the levelling. It is noteworthy that the levelling process removes the large errors (~350 nT) at ship-line cross-points in south-eastern Pine Island Bay, which are a long distance from the nearest compensation loop.

**Gridding and micro-levelling:** Particular care was taken in gridding the data because of the irregular line spacing and orientation. We applied a minimum curvature approach between less sampled areas and those areas with good coverage. Initially, a coarse grid with a large grid cell size is used. Nodes represent either the real data, which is found within a specific radius, or an average of all data points of the grid. Iteratively, the grid is adjusted to the real data nearest to the grid nodes. If an acceptable misfit between real data and coarse grid node is achieved, the cell size is divided by two and the procedure starts again. The gridded data still show some noise resulting from residual levelling and gridding errors. Although these errors are small in comparison to the measured anomalies, they account considerably for noise in further processing. Removing these errors by the application of a low-pass filter to smooth derivative grids is a process known as micro-levelling (Ferraccioli et al., 1998; Green, 1983). As the screening of the line data indicated, the shelf is partitioned into two magnetic anomaly domains characterised by different wavelengths. We found that micro-levelled grids of 5 and 2 km cell size best represent both domains with a minimum of artefacts (Fig. 5A,B). A further decrease of the cell size to 1 km in combination with a 30-km high-pass filter (Fig. 5C) shows clearly the wave-field distinction between the inner shelf domain and the middle to outer shelf but also creates numerous artificial anomalies.

**Further analytical field analysis:** We applied a wide spectrum of standard wave-field processing methods to the grid in order to constrain directional trends and the shapes and edges of magnetic anomaly units. These methods included the calculation and application of (1) horizontal, vertical and tilt derivatives, (2) upward field continuations, (3) high-pass filters, (4) pseudo gravity and (5) analytical signal. At moderate wavelengths, these methods serve to highlight the magnetic bodies and tectonic lineation trends that can be coarsely interpreted from the regular anomaly grids. High-pass filtering, in particular, much more readily delineates the boundary zone between outcropping basement of the inner shelf and the sedimentary cover of the middle and outer shelf (Fig. 5C).

## 4. Depth estimates

In addition to standard power spectra analyses for estimating the depths of magnetic anomaly sources, we applied the Euler deconvolution method to the grid in an attempt to better delineate the spatial distribution and extent of the tops and edges of source bodies. We used the approach of Thomson (1982) and Reid et al. (1990) in which the depth solution is related to a chosen structural index $S_I$ (0 for a planar contact, 1 for a dike or sill, 2 for a vertical pipe, 3 for a sphere). A square window is defined within the grid, which contains at least $3 \times 3$ grid cells. For each grid cell, the deconvolution is calculated. Hence, a $3 \times 3$ cell window results in 9 equations, which are solved via a least-squares inversion. The
window is then shifted across the grid and the same procedure is applied. It is obvious that the solution is dependent on the chosen window size. If the window size is so large that neighbouring anomalies intrude into it, the inversion can fail and its solutions are rejected. This is the case for distinct anomalies on the inner shelf in the south (Fig. 6A). On the other hand, if a window size is too small, then anomalies from deep-seated sources may not be adequately represented in the inversion, interfering with the solutions for shallower sources and resulting in artificially shallow source solutions. Fig. 6B shows that smaller window sizes generate more solutions and that long-wavelength anomalies yield shallower depth solutions than when a larger window is used.

Our two examples of Euler deconvolution solutions use different window sizes of 14 and 20 km but the same structural index (Fig. 6). The solutions provide estimates of a dense distribution of source tops at depths below 7 km over the middle to outer shelf, while further south, below the inner shelf, source tops are typically located at depths shallower than 7 km.

5. Tectonic lineaments and 2D modelling

One of the most striking observations in the magnetic anomaly grid is the domination of the inner shelf by relatively short-wavelength anomalies and the middle and outer shelf by middle to long wavelength anomalies (Fig. 5A,B). The boundary between the two domains is abrupt and clearly shown in the high-pass filtered wavefield (Fig. 5C). This boundary is the northern limit of a coastal and inner shelf domain in which basement crops out or exists in shallow subcrop, as suggested by the rugged topography (Fig. 1). It can be assumed that most of this basement is composed of crystalline rocks similar in type to the granitoids and porphyritic dykes observed from coastal and island outcrops of the embayment (e.g. Kipf et al., 2012; Pankhurst et al., 1993). Conversely, the longer wavelength signals over the middle and outer shelf are consistent with the presence of a thick drape of sediments overlying the basement (Lowe and Anderson, 2002; Weigelt et al., 2009). Depth estimates from Euler deconvolution help identify a boundary zone.
across which the basement drops downwards to the north to about 3 km depth (Fig. 5C), which is consistent with observations from seismic data across this boundary (Graham et al., 2009; Lowe and Anderson, 2002; Uenzelmann-Neben et al., 2007; Weigelt et al., 2009).

One of the objectives of the magnetic surveys was to identify crustal features for delineating tectonic events in the history of the crustal formation in the Amundsen Sea Embayment (Fig. 7). The middle and outer shelf show fundamentally different preferential anomaly strikes. While a major set of long-wavelength anomalies strike WSW–ENE along the middle shelf, the outer shelf west of 107° W exhibits mainly NW–SE striking anomalies which may continue on the inner shelf of the eastern embayment. A third directional trend is observed for positive anomalies on the inner and middle shelf striking NNE–SSW. These are less pronounced but seem to border a broad magnetic zone of low amplitude negative anomalies on the middle shelf at 106–109° W.

We selected two 2D modelling transects crossing the Amundsen Sea Embayment shelf from north to south and NW to SE across the major observed magnetic anomaly lineaments and proposed source bodies (Fig. 7). The data along the transects were taken from the magnetic anomaly grid at 5 km cell size. Encom ModelVision™ was used as modelling software. The depth estimates from Euler deconvolution solutions provided some constraints for the parameterisation of these magnetic susceptibility models. Other constraints on crustal properties, such as crustal thickness, sediment thickness and the approximate locations of mafic (high density, high susceptibility) bodies, were implemented from seismic reflection data (Weigelt et al., 2009), a deep crustal seismic refraction model and a gravity anomaly inversion of the Amundsen Sea shelf region with crustal thicknesses of 20–25 km (Gohl et al., 2007). Water depths were taken from the bathymetry compilation of Nitsche et al. (2007).

We chose a modelling procedure in which we parameterized the model starting with thick layers and large bodies from the lower crust for long-wavelength signals and moving on to small-size layers and bodies of the upper crust to generate short-wavelength signals. The responses were iteratively calculated for a range of susceptibilities and body sizes such that the root-mean-square (rms) misfit between observed and calculated total intensities could be minimised. Due to the different wave-field characteristics of transects ASE-West A, ASE-West B and ASE-East, each model was treated independently of the others. Although various studies show that the ranges of susceptibilities of individual rock types can span several orders of magnitude, it is possible to distinguish domains of sedimentary, felsic and mafic composition on the basis of susceptibilities. Most published susceptibilities of intrusive felsic rocks range from 0.0001 to 0.02 SI unit (average 0.0012 SI), while those of intrusive rocks of mafic composition have values of 0.0001 to 0.13 SI (average 0.012 SI) (e.g. Sanger and Glen, 2003). We applied a constant susceptibility of 0.000005 SI for sediments and of 0.001 SI for the bulk of the continental crust in all models and started modelling with the average values of intrusive bodies by Sanger and Glen (2003). By adjusting the susceptibilities of these bodies, we stayed within the above given ranges.

The model transect ASE-West A (Fig. 8A) covers the continental slope and the outer shelf and includes two high-susceptibility bodies embedded into the upper crust of the continental rise. Much of the upper surface of the large high-susceptibility body 3 beneath the
The modelled properties of the dominant ENE–WSW trending magnetic anomalies A, B, C and D (Fig. 7) indicate that they are caused by major elongated and deep-seated features of predominantly mafic composition. They parallel the initial spreading centre’s azimuth between Chatham Rise and West Antarctica and can thus be related to rift processes occurring before breakup as early as 100 Ma and during breakup between 90 and 85 Ma (Eagles et al., 2004a; Larter et al., 2002; Wobbe et al., 2012). With their seaward-dipping outer edges, bodies of anomalies A and B might be interpreted as thick wedges of basaltic lava and mafic intrusions of the kind known from volcanic extended continental margins such as the Voring margin or Afar Rift (Planken and Eldholm, 1994; Woldenden et al., 2005). The body modelled for anomaly A reaches near the seafloor at its southern limit which correlates with the steep northern flank of the inner Dotson–Getz Trough (Graham et al., 2009; Larter et al., 2009). The trough coming from the Getz B Ice Shelf strikes nearly in east–west direction on the inner shelf before merging with the outflow troughs originating from the Getz A and Dotson Ice Shelves and turning northwards on the middle shelf and to the northwest on the outer shelf (Fig. 1). We suggest that this intrusive body is more resistant to pre-glacial and glacial erosion and caused an east to northeastward deviation of grounded ice streams from the eastern Getz Glacier.

The NW–SE trending set of magnetic anomalies (including E and F of Fig. 7) is parallel to the so-called Peacock Gravity Anomaly northwest of Peacock Sound (Eagles et al., 2004a; Larter et al., 2002) which has been modelled with an underlying high-density body (Gohl et al., 2007). It is interpreted as a magmatic zone, which interferes with the ENE–WSW trending rift structure (causing anomalies A to D). Modelling along transect ASE-West A indicates that the magnetic anomalies in this zone are caused by elongated bodies of similar depth, dimension and susceptibility as those causing the ENE–WSW trending anomalies. Consistent with the plate tectonic process described for the Bellingshausen Plate (Eagles et al., 2004a; Wobbe et al., 2012) we infer that these features represent mafic intrusions emplaced when the southern Bellingshausen Plate boundary was active between 80 and 61 Ma, giving rise to a distributed pattern of compressional, extensional and translational movements at various stages of the activity of this boundary. The location of these magmatic units over a long time period in a very wide continent suggests the presence of a sedimentary sub-basin within the middle shelf. The margins of this low are asymmetrical; its southeastern margin is characterised by greater amplitudes and shorter wavelength anomalies than its northwestern margin. The basin, therefore, might be interpreted as having a half-graben geometry, with a bounding basement-involved normal fault zone on its southeastern edge, and a downthrown floor characterised by broad flexure northward away from the fault zone. The Palaeogene plate tectonic model of Müller et al. (2007) predicts that a crustal boundary with this orientation would have responded to West Antarctica–East Antarctica relative motion along the WARS in right-lateral strike-slip motion. They suggest that this boundary extended from the Bentley Subglacial Trough and Byrd Subglacial Basin across western Ellsworth Land and the Thurston Island block, where it continues into the area of the dominant north-trending gravity anomaly lineaments of the Bellingshausen Sea. More consistent with the interpretation of a NNE-trending basin, however, the alternative plate tectonic model of Cande et al. (2000) predicts a strongly oblique right-lateral extension. In any case, the orientation of the identified sub-basin on the eastern ASE shelf (annotated as ‘WARS rift basin’ in Fig. 7) infers that it may be related to eastern WARS motion, possibly active as early as 55 Ma and lasting until at least 30 Ma before the zone of translatational deformation migrated or jumped eastward to join the triple junction at the spreading ridge colliding with the convergent margin of the southern Antarctic Peninsula as postulated by Müller et al. (2007). Somewhat more speculative, but not unrealistic, is a correlation of the sub-basin with the dextral transtensional strain in western Marie Byrd Land observed by Siddoway (2008), who interpreted this strain as an expression of an early WARS activity in the Late Cretaceous.

The shallow, small-sized bodies of very high susceptibilities between 0.02 and 0.09 SI unit on the inner shelf (Fig. 8b,c), are interpreted to be caused by a mafic dyke field related to the Cenozoic volcanic province of eastern Marie Byrd Land and western Ellsworth Land (e.g. LeMasurier, 1990; Paulsen and Wilson, 2010). Due to the presence of the nearby volcanic Hudson Mountains, one might expect volcanic, ~5 km radius cone-like features on the shelf as well, but the limited spacing of the survey does not allow for the unequivocal identification or clear resolution of such features.

7. Conclusions

Newly acquired magnetic datasets from helicopter-borne and ship-borne surveys in 2006 and 2010 were used to generate levelling anomaly grids from the shelf of the Amundsen Sea Embayment. We applied grid processing, Euler deconvolution and 2D modelling for the analysis of magnetic anomaly patterns, identification of structural lineaments, and characterisation of magnetic source bodies.

The magnetic anomaly grid clearly outlines the boundary zone between the inner shelf where basement rocks crop out and a more thinly-sedimented middle to outer shelf. This sedimentation can be related to a long history of distributed extension inferred from the regional plate tectonic setting and from the presence of distinct zones of anomaly patterns and lineaments that can be associated with at least three tectonic phases in that history.

The first of these phases was the establishment of magmatic emplacement zones during Cretaceous New Zealand–Antarctic rifting and breakup (100–85 Ma). The second phase involved further magmatism in a distributed plate boundary zone between the Bellingshausen and Antarctic plates (80–61 Ma). Finally, the West Antarctic Rift System may have caused further extension in the region at later times (55–30 Ma?), focussed within a possible half-graben basin system bounded by NNE–SSW trending tectonic lineaments. Mafic dykes cross the inner shelf of the embayment and Pine Island Bay and may be associated with Cenozoic volcanism observed on land. They possibly relate to activities of one of the eastern branches of the West Antarctic Rift System.

Our magnetic data and the analysis show for the first time an identification of at least three sets of superposed tectonic lineaments in the Amundsen Sea Embayment shelf. The correlation of this observation to plate tectonic motion and associated magmatic emplacement in the West Antarctic realm suggests that the wide shelf of the Amundsen Sea Embayment may be the product of a long-lived zone...
of distributed crustal deformation, and that it may host multiple failed rifts and rift accommodation zones. By an example of a modelled intrusive body, this study also shows that such emplacement of resistive material acted as an obstacle for past ice flows across the inner shelf.

Acknowledgements

Many thanks to the ship crews, the helicopter pilots of Heli-Service GmbH and the geophysical teams for their support of acquiring the magnetic data during RV Polarstern expeditions ANT-XXIII/4 (2006) and ANT-XXVI/3 (2010). We, in particular, thank Volker Leinweber, Sonja Sukcro and Christina Mayr for their engagement in the helicopter-magnetic team of 2006. The collaboration with Prof. B. Tetzkan of the University of Cologne in co-supervising A. Denk’s Diploma/MSc thesis, of which parts contributed to this paper, is gratefully acknowledged. This project is entirely supported by institutional funds of the Alfred Wegener Institute.

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