

Crustal structure and tectonic provinces of the Riiser-Larsen Sea area (East Antarctica): results of geophysical studies

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Abstract About 16,000 km of multichannel seismic (MCS), gravity and magnetic data and 28 sonobuoys were acquired in the Riiser-Larsen Sea Basin and across the Gunnerus and Astrid Ridges, to study their crustal structure. The study area has contrasting basement morphologies and crustal thicknesses. The crust ranges in thickness from about 35 km under the Riiser-Larsen Sea shelf, 26–28 km under the Gunnerus Ridge, 12–17 km under the Astrid Ridge, and 9.5–10 km under the deep-water basin. A 50-km-wide block with increased density and magnetization is modeled from potential field data in the upper crust of the inshore zone and is interpreted as associated with emplacement of mafic intrusions into the continental margin of the southern Riiser-Larsen Sea. In addition to previously mapped seafloor spreading magnetic anomalies in the western Riiser-Larsen Sea, a linear succession from M2 to M16 is identified in the eastern Riiser-Larsen Sea. In the southwestern Riiser-Larsen Sea, a symmetric succession from M24B to 24n with the

central anomaly M23 is recognized. This succession is obliquely truncated by younger lineation M22–M22n. It is proposed that seafloor spreading stopped at about M23 time and reoriented to the M22 opening direction. The seismic stratigraphy model of the Riiser-Larsen Sea includes five reflecting horizons that bound six seismic units. Ages of seismic units are determined from onlap geometry to magnetically dated oceanic basement and from tracing horizons to other parts of the southern Indian Ocean. The seaward edge of stretched and attenuated continental crust in the southern Riiser-Larsen Sea and the landward edge of unequivocal oceanic crust are mapped based on structural and geophysical characteristics. In the eastern Riiser-Larsen Sea the boundary between oceanic and stretched continental crust is better defined and is interpreted as a strike-slip fault lying along a sheared margin.

Keywords Antarctica · Continental margin · Riiser-Larsen Sea · Geophysical data · Seismic stratigraphy · Crust

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Introduction

The Riiser-Larsen Sea is the westernmost of several marginal seas that lie off the lengthy East Antarctica coastline facing the Indian Ocean. The Riiser-Larsen Sea is bounded by prominent bathymetric ridges, the Astrid Ridge in the west and the Gunnerus Ridge with the adjoining Kainanmaru Bank in the east. All existing reconstructions of adjoining portions of Africa, Madagascar, India and Antarctica demonstrate land-mass overlap (see e.g. Martin and Hartnady 1986; Lawver et al. 1992, 1999), apparently

mainly due to lack of information on the amount of pre-break-up crustal extension that affected marginal portions of the Gondwana supercontinent. In earlier works devoted to Gondwana reconstructions, many authors (e.g. Norton 1982; Lawver et al. 1992; Roeser et al. 1996) used the 2,000 m isobath as the first approximation to the pre-breakup plate (continent-ocean) boundary. In later studies (e.g. Lawver et al. 1999; König and Jokat 2006) the continent-ocean boundary was assumed to coincide with the coastal free-air gravity anomaly derived from satellite gravity database. Both these approaches are a reasonable simplification to define the fit of continents but do not provide rigorous solutions. This paper summarizes for the first time the results of extensive marine surveys performed in the Riiser-Larsen Sea area aboard the Russian ship RV “Akademik Alexander Karpinsky” by the Polar Marine Geosurvey Expedition in cooperation with the Norwegian Petroleum Directorate (NPD) and the University of Bergen and presents interpretation of collected data in the light of prior studies.

Previous work

Previous geological investigations in the Riiser-Larsen Sea were done principally by countries with permanent stations or seasonal field bases located in Dronning Maud Land. The first marine geophysical data in the Riiser-Larsen Sea area were obtained in 1975–1981 by the South African National Antarctic Expedition (SANAE). Reconnaissance bathymetric, magnetic and single-channel seismic observations were carried out mainly in the western Riiser-Larsen Sea. In the Astrid Ridge area these early surveys considerably improved the bathymetry, and mapped the shallow basement morphology and sediment distribution. In deeper parts of the Riiser-Larsen Sea they provided the first insight into early basin evolution from identification of Mesozoic magnetic anomalies M1–M10N (Bergh 1987). The SANAE survey also outlined an elongated basement high trending northeast from the northern part of the Astrid Ridge. Bergh (1987) cites this basement high as the Astrid Fracture Zone, which indicated the Mesozoic direction of relative motion between Africa and Antarctica.

During the 1981/82 austral season, the Indian Antarctic Expedition acquired 450 km of high-resolution seismic reflection data using a multi-electrode 8–24 kJ sparker source. The survey lines were oriented in an E-W direction and concentrated in the axial zone of the southern Astrid Ridge. They detected a linear sub-longitudinal graben about 20 km wide and 80–90 km long. Deformed sediments and offsets at the seabed within the graben suggested relatively recent stresses, which were related to changes in relative motion of the African, Indian and Antarctic plates (Rao et al. 1992).

Implementation of advanced seismic technologies (involving multi-channel seismic reflection (MCS) profiling and refraction observations) in combination with ship-based and airborne potential field studies provided better insight into the geological structures and geodynamic history of the area. Such surveys were accomplished by German, Japanese, Russian and joint Russian-Norwegian expeditions, and the acquired data are synthesized in this paper. The location of MCS survey lines is shown in Fig. 1.

In 1978, the Federal Institute for Geosciences and Natural Resources of Germany (BGR) carried out MCS, magnetic, gravity and sonobuoy studies off Dronning Maud Land between 30°W and 20°E. A 48-channel, 2668-m-long analogue streamer and airgun array of 23.4 l total volume were used for seismic profiling (Hinz and Krause 1982). About 1,500 km of MCS lines and six sonobuoys were acquired in the southern Astrid Ridge area and the adjacent south-western part of the Riiser-Larsen Sea. In the southern Astrid Ridge, Hinz and Krause (1982) identified a prominent reflector overlying the higher-velocity (>4 km/s) layered unit whose nature was not clearly defined.

During the 1985/86 season, the Japan National Oil Corporation conducted research in the eastern part of the area and collected about 1,000 km of MCS data, more than 1,500 km of magnetic and gravity profiles and five sonobuoys, as well as three dredges of bottom sediments (Saki et al. 1987). Seismic data were acquired using a 24-channel, 600-m-long streamer and two 6 l airguns. Two regional unconformities were mapped in the sedimentary cover. An acoustic interface with velocities 5.4–5.8 km/s within the Gunnerus Ridge and 4.5 km/s below the middle continental rise of the eastern Riiser-Larsen Sea were identified and interpreted as basement of “continental” and “oceanic” nature, respectively. Dredged granitoid and gneissic rocks were interpreted as autochthonous debris reflecting the composition of the Gunnerus Ridge and the Kainanmaru Bank basement.

In 1990, BGR and Alfred Wegener Institute for Polar and Marine Research (AWI) jointly performed a reconnaissance geophysical survey in the deep-water Riiser-Larsen Sea and across the Gunnerus Ridge. About 2,100 km of MCS, magnetic and gravity data were acquired during this survey. The seismic data were recorded with a 48-channel analog streamer and airgun array of 25.6 l in volume. The seismic and potential field data allowed general characterization of the basement structure and reconstruction of the depositional history of the Riiser-Larsen Sea basin. The seafloor spreading model for the western Riiser-Larsen Sea based on earlier magnetic data (Bergh 1987) and the new data were improved

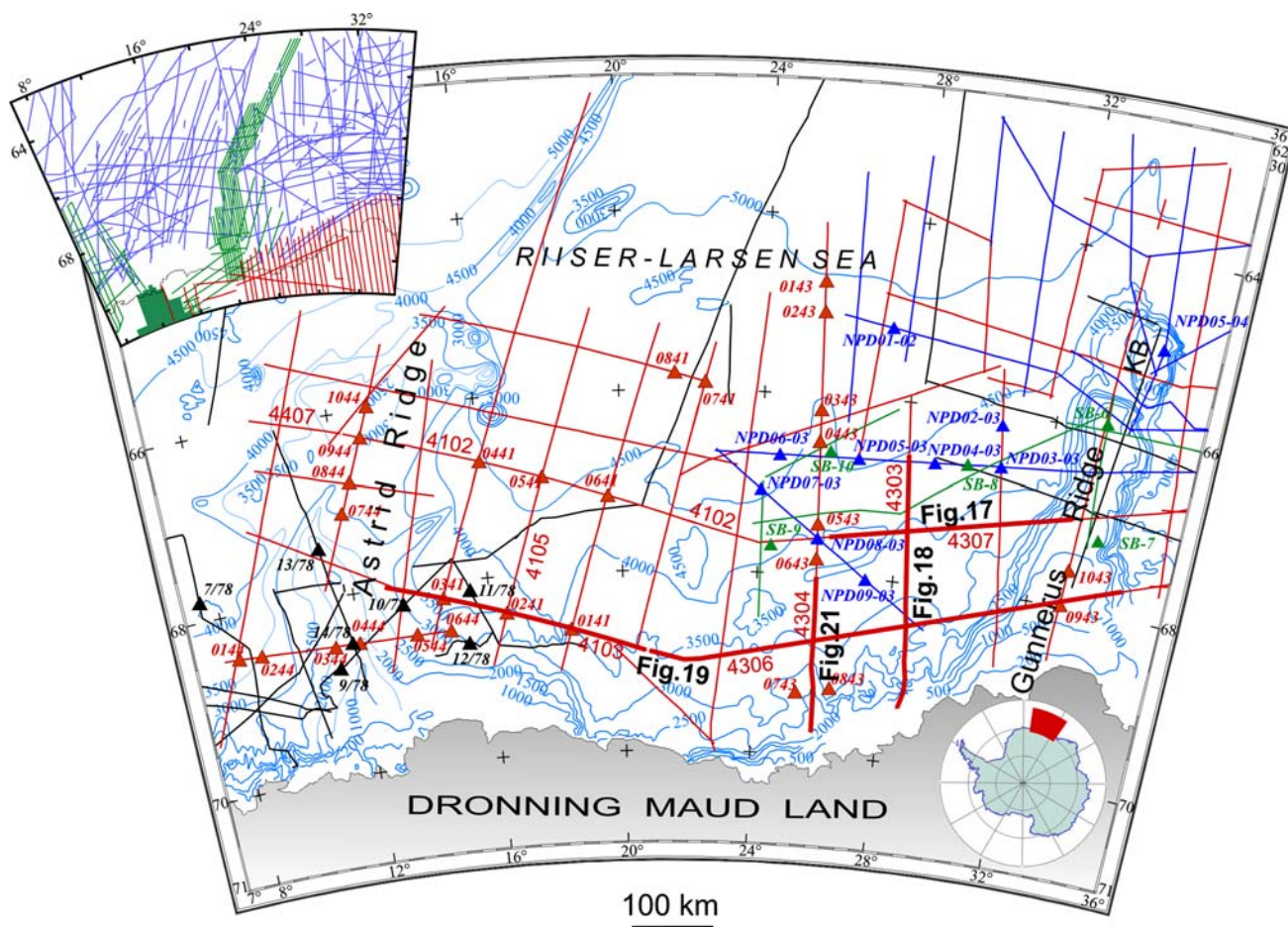


Fig. 1 Bathymetry of the Riiser-Larsen Sea area (modified from GEBCO 1997) and location of geophysical data. Names of multi-channel seismic profiles (lines) and sonobuoy sites (triangles with sonobuoy names inscription) correspond to original field documentation by national expeditions (red—Russian, black—German,

green—Japanese, blue—joint Norwegian-Russian cruises). Inset shows aerogeophysical flight lines (red—Russian, green—German, green box—German detailed survey with 4.4 km line spacing) and all marine magnetic observations (blue). Thicker red lines are fragments of MCS profiles illustrated in other Figs. KB—Kainanmaru Bank

with identification of anomalies back to Chron M24 (Roeser et al. 1996), thus advancing ideas on East Gondwana early break-up (separation of Africa, Madagascar, India and Antarctica). Additional MCS lines (414 km) were acquired by BGR in 1996 in the southern Astrid Ridge and the immediate Lasarev Sea neighborhood with a 120-channel, 3000-m-long streamer and 53.5 l airgun array (Hinz et al. 2004).

Airborne remote sensing technologies were implemented in the Riiser-Larsen Sea area during three field seasons (see inset in Fig. 1). An aerogeophysical survey (magnetic, gravity and radio-echo sounding observations) was carried out in 1990/1991 season by the Polar Marine Geosurvey Expedition (PMGE) under the auspices of the Russian Antarctic expedition. The lines spaced at 20 km were flown at 2,500 m altitude above sea level (a.s.l.) over the inshore area of East Antarctica between 22°E and 45°E (Golynsky et al. 2002). The survey mapped a

prominent high-amplitude (>500 nT) linear magnetic anomaly of up to 80–100 km in width extending over the Riiser-Larsen Sea shelf and continental slope. Part of the same anomaly had also been detected between 7°E and 16°E on the shelf south of the Astrid Ridge by a more detailed aeromagnetic survey conducted in 1996/1997 by BGR at 2,800 m a.s.l. flying altitude, with 4.4 km line spacing (Damaske et al. 2005; Fig. 1, inset). AWI made an important test of the linearity and strike of the marine magnetic anomalies by a magnetic survey carried out by a ship-based helicopter during the 2000/2001 season within a narrow swath (Jokat et al. 2003; Fig. 1, inset). The N-S and NE-SW oriented lines were spaced at about 9 km and flown at 150 m a.s.l. altitude. The new data generally supported the earlier seafloor spreading model of Roeser et al. (1996). They also enabled new insights into the early Riiser-Larsen Sea opening, such as resolving the difference in spreading

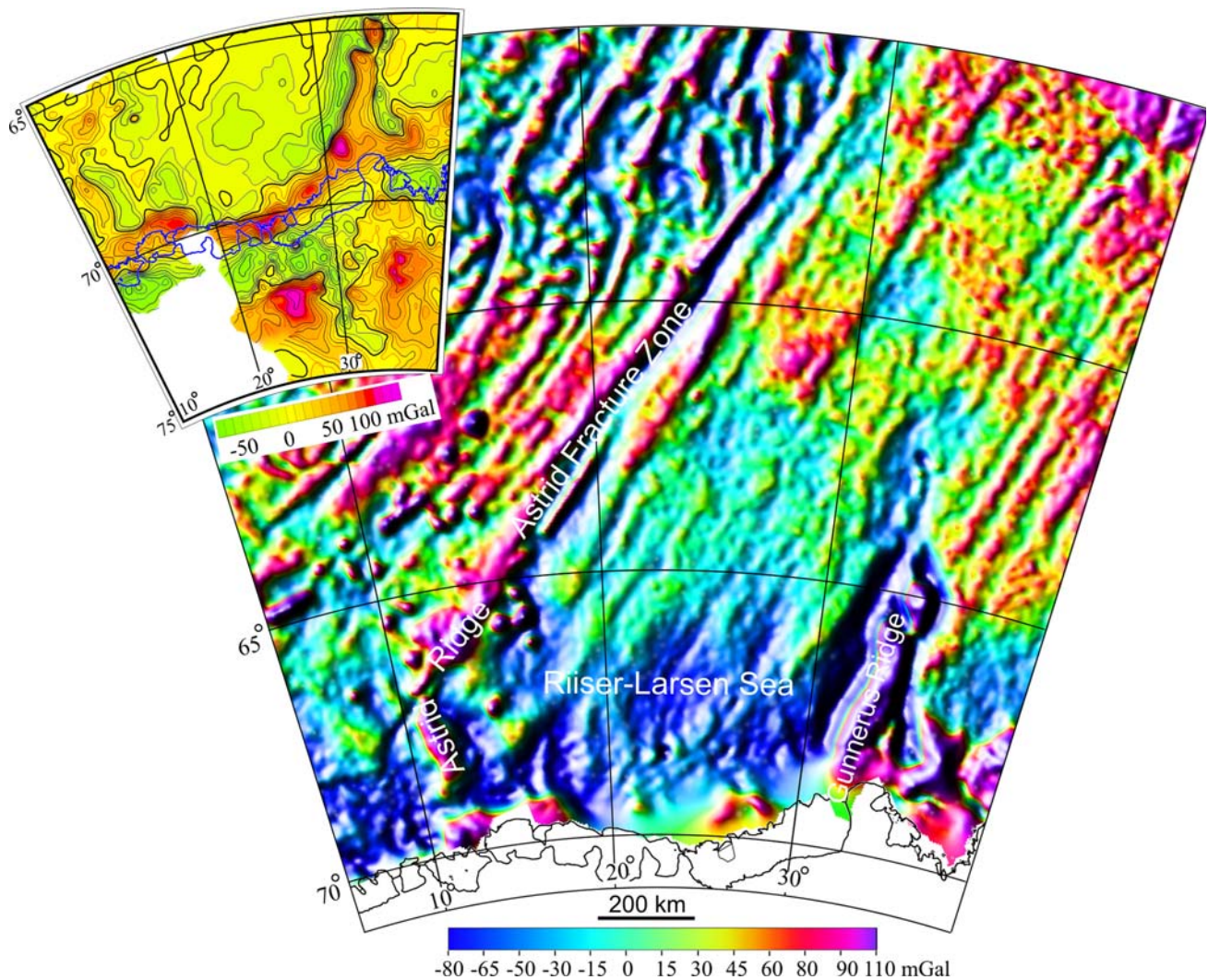


Fig. 2 GEOSAT/ERS-1-derived free-air gravity anomaly map of the southernmost Indian/Atlantic boundary region (from Sandwell and Smith 1997). Inset shows free-air gravity anomaly map of eastern

Dronning Maud Land (from Russian airborne gravity surveys) and the Riiser-Larsen Sea (from marine surveys) compiled in this study using all available information

directions before and after Chron M23 (Jokat et al. 2003). The results of these airborne surveys were partly incorporated into an overview magnetic anomaly map of Antarctica compiled in the framework of Antarctic Digital Magnetic Anomaly Map (ADMAP) project (Golynsky et al. 2007).

Geosat and ERS-1 satellite altimeter data collected over the Southern Ocean and recomputed into gravity field (Sandwell and Smith 1997) provided additional information relevant to this study (Fig. 2). These data sets illustrate the intricate pattern of ocean ridges, fracture zones and other basement and seafloor features significant for geodynamic interpretations of the Riiser-Larsen Sea area.

A major contribution to marine investigations in the study area was made by PMGE operating RV “Akademik Alexander Karpinsky”. The vessel conducted six

successive cruises, the first three (1996, 1998, and 1999) as part of the Russian national Antarctic program, and the subsequent ones (2002, 2003 and 2004) under a cooperative agreement between PMGE and the Norwegian Petroleum Directorate (NPD). Altogether 16,140 km of MCS lines with gravity and magnetic measurements and 28 sonobuoys were obtained in a systematic grid that covered the entire Riiser-Larsen Sea, Astrid Ridge and Gunnerus Ridge (Fig. 1). The data collected by PMGE during the first (1996) cruise in the Astrid Ridge area were included in Hinz et al. (2004) and referred mainly to comparison of geophysical characteristics in the Lazarev Sea and the western Riiser-Larsen Sea. The acquisition and processing technologies applied during the “Akademik Alexander Karpinsky” cruises and for subsequent laboratory studies are described below.

Table 1 Equipment and recording parameters of MCS surveys

	PMGE (1996)	PMGE (1998)	PMGE (1999)	NPD/PMGE (2002)	NPD/PMGE (2003)	NPD/PMGE (2004)
Source	6 PI-20					SG-2B
Number of airguns	2					9-18
Net volume	40 l					47 l
Pressure (kg/cm ²)	120–130					140
Gun depth (m)	6–13			4–6		5
Shot interval (m)	100	50				50
Streamer	GECO-PRACLA				MSX-6000	
Length (m)	2400	200	1800	1800	3000	4400
Near offset (m)	605	150	115	162	225	150
Group interval (m)	50	25			12.5	12.5
Depth controller	No				Yes	Yes
Number of channels	48	96	48	72	240	352
Recording	GAK (Russia)				MSX (USA)	MSX (USA)
Sample interval (ms)	4	2				2
Recording Filter (Hz)	4–125	4–125		4–125	2–206	2–206
Coverage (fold)	12	24			30	44

Data acquisition and processing

Acquisition of MCS data during PMGE and joint PMGE/NPD surveys aboard RV “Akademik Alexander Karpinsky” was accomplished using equipment and recording parameters as listed in Table 1.

Sonobuoys used to record refraction and wide-angle reflection data and to compute crustal velocity parameters were engineered at PMGE as recoverable equipment (i.e., raised onboard after data acquisition) and supplied with a floatation system to weaken the influence of ocean waves and minimize ambient noise. Sonobuoy measurements were conducted with 80 l airgun source for PMGE surveys and 40 l for PMGE/NPD surveys. The shot intervals were 100 m and 50 m, respectively. The maximum offsets with reasonable signal-to-noise ratios ranged from 40 to 55 km, and in many cases the refraction profiles were reversed.

The MCS data collected during PMGE and PMGE/NPD cruises were processed using IBM RISK SYSTEM-6000 with ProMAX software (ADVANCE Geophysical Corporation, USA). Processing of the data usually followed a conventional sequence which included demultiplexing, predictive deconvolution, band pass filter, velocity analysis (every 5–7 km), normal moveout correction (stacking), automatic gain control and display scaling.

Minimum phase deconvolution, automatic gain control and filtering were applied to process the sonobuoy field records before well-correlated first and secondary arrivals could be digitized for subsequent crustal modeling. For modeling of refraction data (mostly first arrivals), we used WINZOND software developed by PMGE. This software

is capable of computing one-dimensional depth-velocity solutions under the assumption of a horizontally bedded medium with constant layer velocities. Refraction (head wave) velocities at the tops of layers were applied as layer velocities and in most cases stacking velocities derived from the processing of MCS data were used for upper layers. To minimize the influence of the inclination of layer boundaries, depth and velocity values were determined as the averages between one-dimensional solutions of two reversed sonobuoy records. Figure 3 shows one-dimensional velocity-depth crustal sections derived from PMGE and PMGE/NPD sonobuoy data and extracted from Hinz and Krauze (1982) and Saki et al. (1987).

The major unconformities identified by seismic stratigraphic analyses (including the basement surface) were correlated with reflection and associated refraction arrivals recorded by sonobuoys. Where a good correspondence of respective horizons was established, seismic velocities in the underlying layer were estimated and considered sufficiently reliable in view of the very small dips (normally less than one degree) of crustal boundaries. All velocities were then integrated into a regional velocity model that is based mainly on the results of PMGE and PMGE/NPD surveys, although sonobuoy data obtained by German and Japanese expeditions were also incorporated. The velocity model was applied to make time-to-depth conversions of the MCS sections, which were subsequently used to compile structural maps (contoured in km; presented in next section) and to characterize the properties of the sedimentary cover and basement across the entire study area.

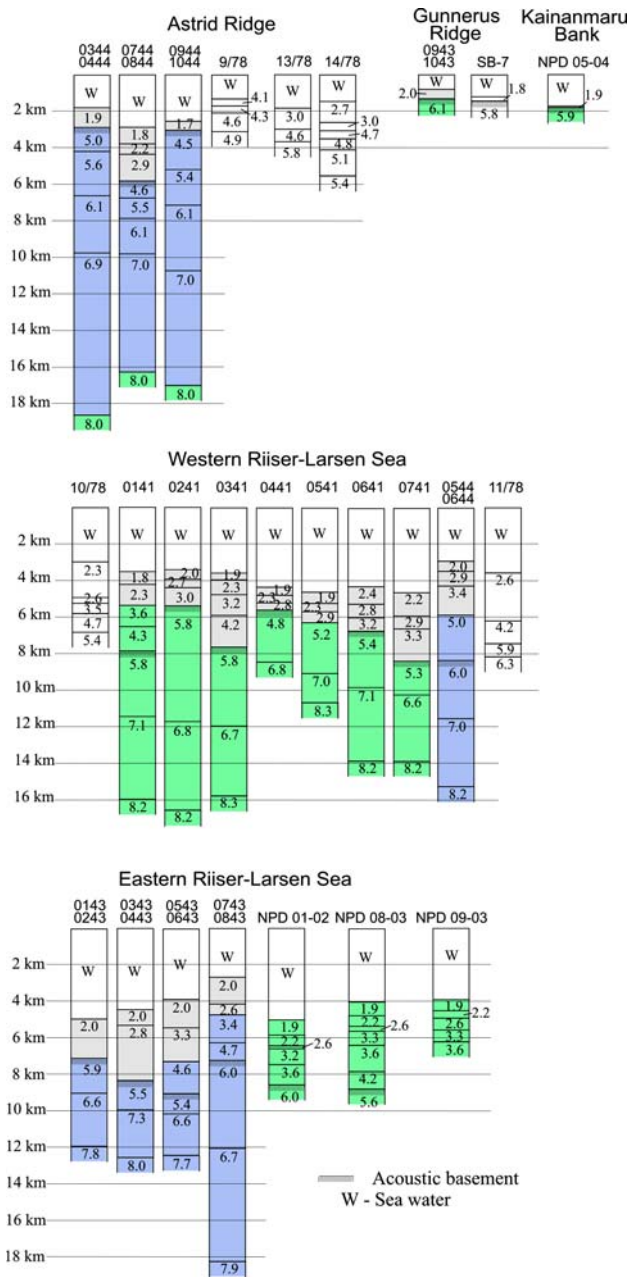


Fig. 3 Velocity (in km/sec) versus depth sections for the Riiser-Larsen Sea crust derived from sonobuoy data. Sections calculated from PMGE and PMGE/NPD sonobuoy data are colored to show how velocities were derived. Velocities of grey-colored crustal layers are stacking velocities derived from MCS data processing. Velocities of blue-colored crustal layers are averaged refraction velocities derived from a pair of reversed sonobuoys. Velocities of green-colored crustal layers are refraction velocities derived from one of reversed sonobuoys or from single sonobuoy. See Fig. 1 for sonobuoy locations

Seismic surveys in PMGE and PMGE/NPD cruises in the Riiser-Larsen Sea included potential field studies. Magnetic measurements were carried out with a magnetic gradiometer. Two magnetometer sensors (accuracy of about 0.1 nT) towed at the distances of 175 m and 350 m

recorded the total magnetic intensity. A “CHETA-AGG” marine gravimeter and a pendulum AMP-1 gravimeter system installed near the center of the ship were used for measurements of relative increments and absolute values of the gravity field, respectively. Reference values for absolute gravity measurements were obtained in Capetown before and after each field season. Global Positioning System (GPS) with the MX-4400 receiver and standard ship equipment were used for navigation. All geophysical observations were accompanied by water depth recording with a 12 kHz echo-sounder. Onboard processing of magnetic and gravity data was accomplished using software developed at PMGE. The total intensity of the magnetic field free of daily variations was derived by integration of magnetic gradients along the profiles. Magnetic anomaly values were calculated with the International Reference Field (IGRF 2000). The RMS error of magnetic measurements determined by crossover analysis ranged between 10 and 20 nT. Free-air gravity anomalies were calculated using the 1985 Gravity Formula. The RMS of gravity measurements was less than 1 mGal.

Structure of Riiser-Larsen Sea from hydrographic and geophysical data

Bathymetry

The bathymetry of the Gunnerus Ridge and the Astrid Ridge is relatively well documented, while that for the adjacent ice-covered and largely inaccessible Riiser-Larsen Sea shelf is not well known (Fig. 1). The Gunnerus Ridge and the Kainanmaru Bank average 100–130 km in width and stretch from the coastline in a north-northeast direction for almost 450 km. The tops of the ridges form a smooth, flat surface that deepens northward from 500 m to 1,500–2,000 m and is bordered by relatively straight, steep slope escarpments with gradients of about 3° on the west, and up to 5° on the eastern and northern sides. The base of the slopes lies in water depths of 4,000–4,500 meters. The Kainanmaru Bank is separated from the Gunnerus Ridge by a 50 km wide and 500–1,000 m deep sea valley.

The Astrid Ridge is delineated by 3,500–4,000 m isobath (Fig. 1). The southern Astrid Ridge is nearly linear and has a well defined crest with relatively gentle western and eastern slopes. North of 67°S the Astrid Ridge bends to the northeast and its relatively flat sea bottom is locally disrupted by steep escarpments. North of 65°S, a narrow long spur of the Astrid Fracture zone delineated by the 4,000–5,000 m isobaths extends to the northeast.

The sea floor between the Astrid Ridge and the Gunnerus Ridge is relatively flat with gradients 2–3. Numerous canyons with walls up to 1,000 m high cut into

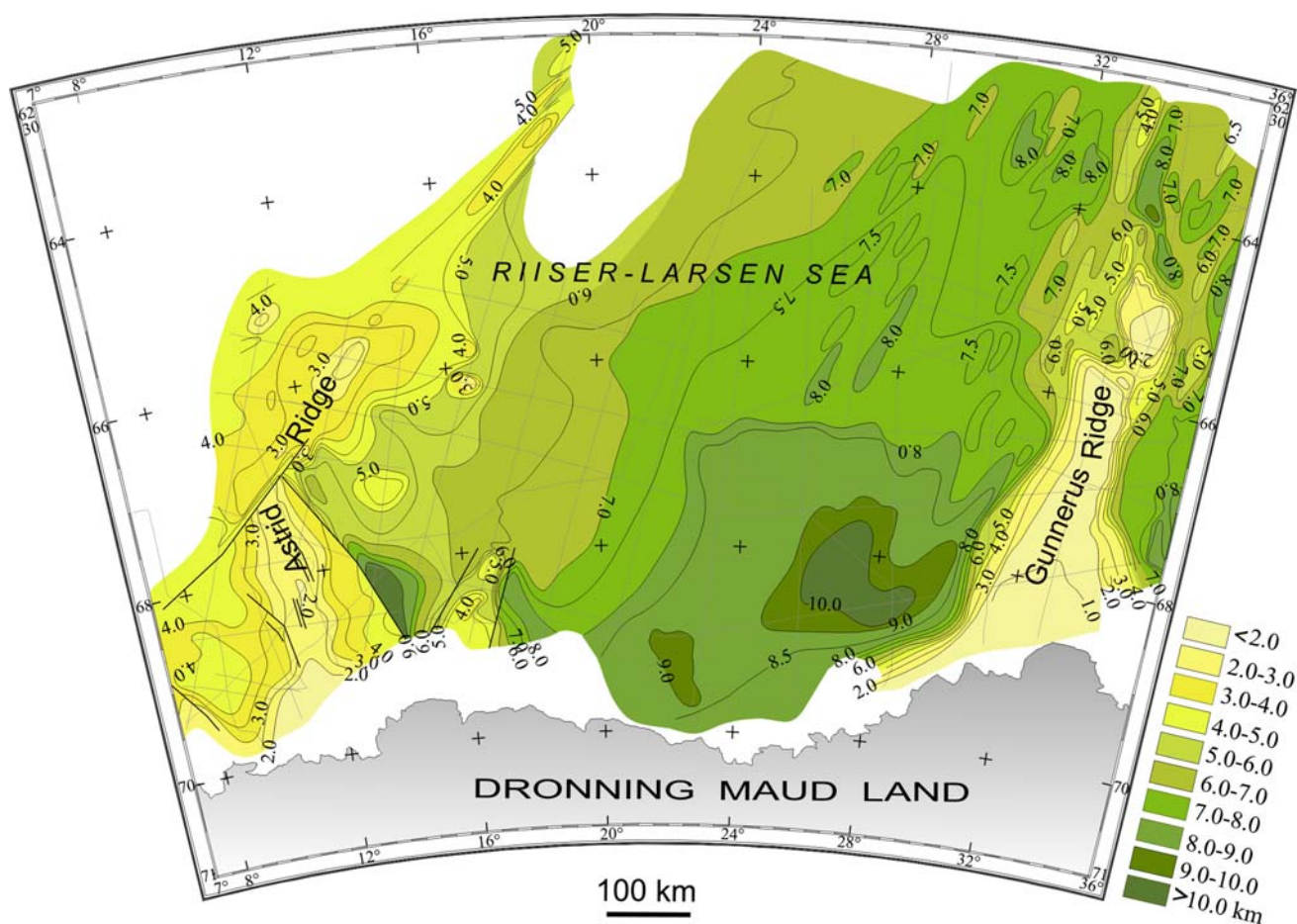


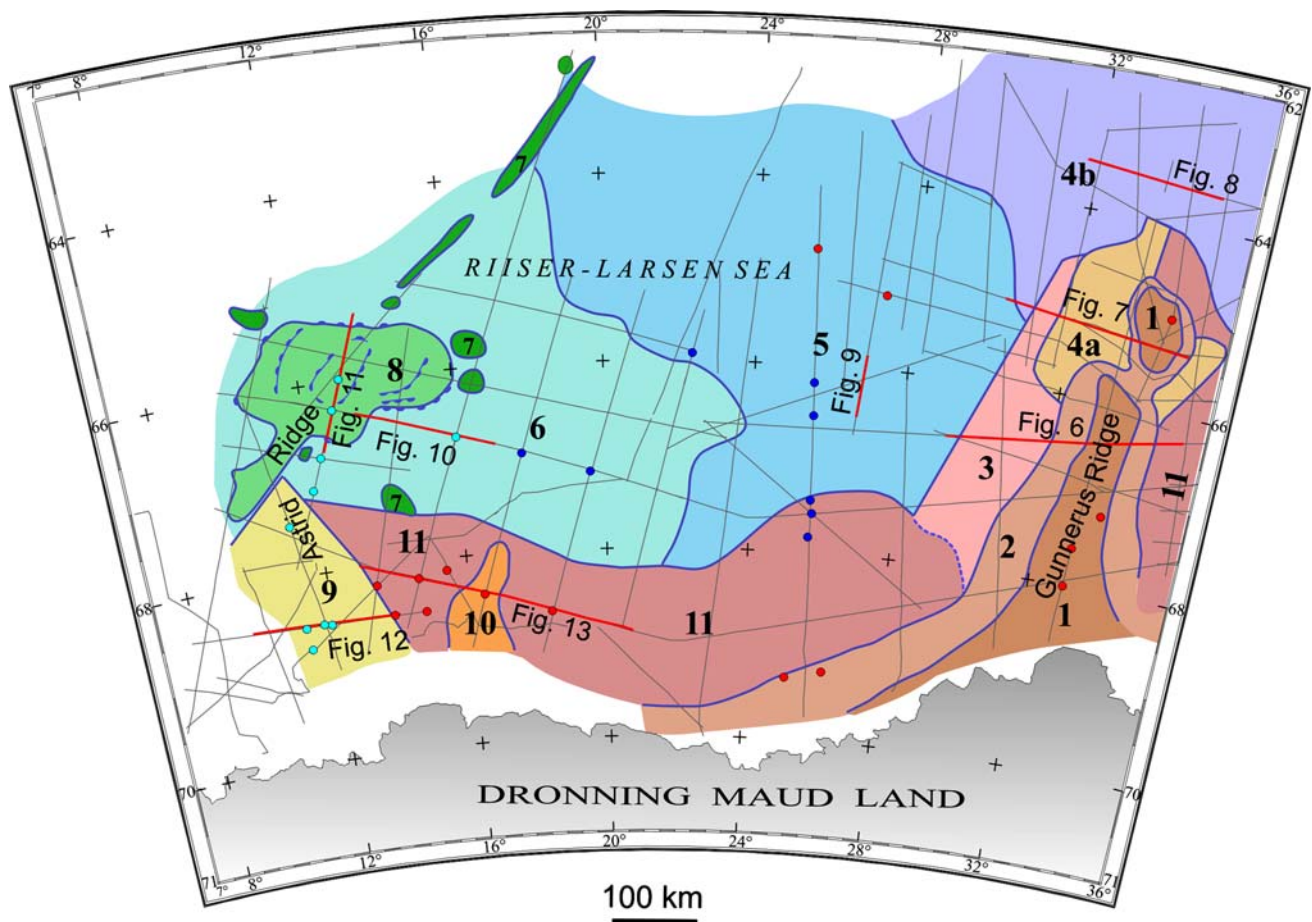
Fig. 4 Total depth to basement in the Riiser-Larsen Sea area

the lower continental slope. The relief of canyons diminishes across the continental rise and disappears at depths exceeding 4,000 m. The outer limit of the continental rise is poorly defined and probably lies north of the 4,500–5,000 m isobath.

Basement morphology

The acoustic basement can be traced almost continuously throughout the Riiser-Larsen Sea area. It is identified in seismic records as a strong reflector separating the clearly stratified sedimentary succession from the underlying unit, which is either acoustically opaque, or contains weak and irregular but distinguishable reflectors attributed to intra-basement layering. Based on this identification a depth-to-basement map with a 0.5–1.0 km contour interval has been manually compiled (Fig. 4) using the seismic data and supplementary information, such as satellite altimetry, aeromagnetic data, the apparent influence of shallow basement morphology on sea floor topography, and visual inter-line correlation of the basement features expressed in seismic images.

Throughout the study area, the average depth, declivity, reflectivity and morphology of the basement surface varies considerably, as do the basement velocities. Using these characteristics we identified 11 types of basement, which are schematically presented in Fig. 5. Examples of basement morphology in the Gunnerus Ridge and the Kainanmaru Bank are illustrated in Figs. 6 and 7. The main basement high of the Gunnerus Ridge (basement type 1) has a slightly convex top surface 60–75 km wide; it dips northward from about 0.5 km b.s.l. near the shelf to 2.0 km b.s.l. in the north, with a gradient of approximately 8–10 m/km, or about 0.5° . The surface is smooth between the coast and 67°S , but further north it displays relief up to 100–200 m. The slopes of the Gunnerus Ridge (basement type 2) are somewhat asymmetric and steep ($7\text{--}8^\circ$ to 15° ; Fig. 6). At the northern end of the ridge a NW-SE-striking extensional graben cuts off a slightly elongated basement block about 75×100 km in size underlying the Kainanmaru Bank (basement type 1). The basement slopes around the Kinanamaru Bank (basement type 2) are steeper than those of the Gunnerus Ridge sloping as



Basement type	Roughness (relief range, km)	Depth (km)	Notes
1	Smooth to slightly rugged	1–2	
2	Downfaulted	1–9	Sharply descending
3	Moderately rugged (1–1.5)	7–8	Moderate diffractions
4	a Highly rugged (1–2)	5–7	
	b Highly rugged (1–3)	6–9	
5	Slightly to moderately rugged (1–1.5)	7–8	Numerous diffractions
6	Smooth	5–7	Underlying reflectors
7	Smooth (up to 3)		Isolated basement highs
8	Smooth, terraced	3–5	Underlying reflectors
9	Smooth	2–4	Underlying reflectors
10	Highly rugged (1–2)	4–7	
11	Slightly to moderately rugged (1–1.5)	8–10	

— - scarps ●● - refraction velocities at the basement surface (red - 5.8–6.1 km/s; blue - 5.0–5.6 km/s; light-green - 4.4–5.2 km/s)

Fig. 5 Distribution of morphologically and seismically distinct types of basement surface in the Riiser-Larsen Sea area. Red lines indicate the fragments of seismic profiles illustrated in Figs. 6–13

much as 15–25°. West of the Gunnerus Ridge the basement lies at 7–8 km depths and has NNE-SSW-striking block-faulted structures (basement type 3; Figs. 4, 6).

Basinward of the Kainamaru Bank, basement deepens to 5–7 km b.s.l. and its surface is rough having amplitudes 1.5–2.0 km (basement type 4a; Figs. 4, 7). A similar basement relief is observed at average depths of

Fig. 6 Seismic section showing basement morphology of the northern Gunnerus Ridge and the adjacent eastern Riiser-Larsen Sea (basement types 1, 2 and 3). See Fig. 5 for location

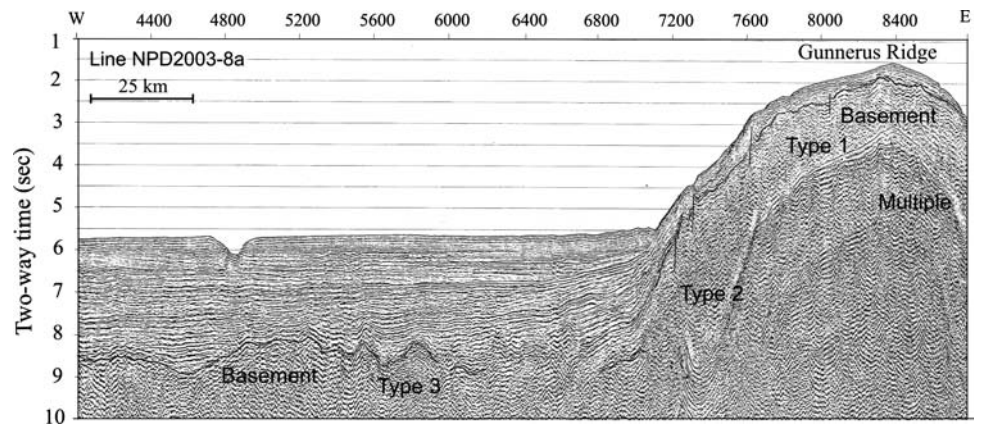


Fig. 7 Seismic section showing basement morphology of the Kainanmaru Bank and the adjacent eastern Riiser-Larsen Sea (basement types 1, 2, 3, 4a and 4b). See Fig. 5 for location

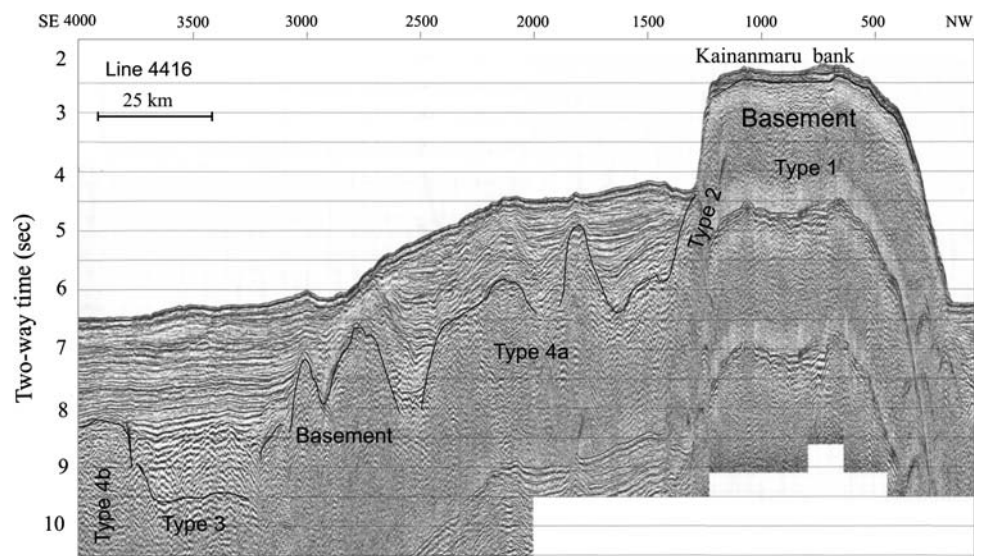
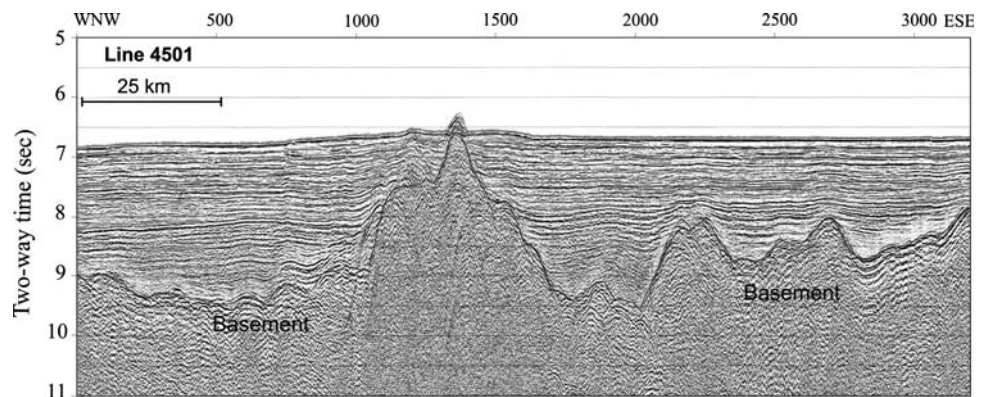


Fig. 8 Seismic section showing basement morphology of the northeastern Riiser-Larsen Sea (basement type 4b). See Fig. 5 for location

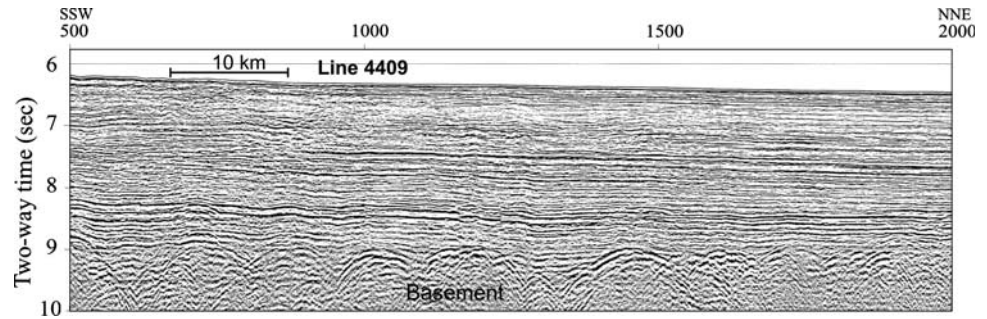


approximately 7–8 km b.s.l. further north and northwest (basement type 4b; Fig. 8). Here, the basement relief is locally up to 4–5 km with some ridges piercing the seafloor.

The east central Riser-Larsen Sea is characterized by small to moderately rough basement at 7–8 km b.s.l. with prominent diffractions (basement type 5; Figs. 4, 9). In the west central basin, basement is smooth and has occasional

internal reflectors (basement type 6). Nearby, at the base of the northern Astrid Ridge, the basement is distinctly stratified (Fig. 10) and is complicated by isolated isometric rises (up to 2.5 km relief) crossed by few seismic lines and documented spatially by satellite-derived gravity anomalies (basement type 7; Figs. 2, 4, 5). The basement surface lies at 2.5–5.0 km b.s.l. below the northern Astrid Ridge and has subhorizontal internal layering over an area of

Fig. 9 Seismic section showing basement morphology of the central Riiser-Larsen Sea (basement type 5). See Fig. 5 for location



about 30,000 km² (basement type 8; Figs. 4, 10, 11). A series of step-like scarps with relief of up to several hundred meters successively elevate the basement to its shallowest position in the center of the ridge. Hinze et al. (2004) interpret these scarps as the edges of individual and superimposed basaltic flows. The concentric configuration of the scarps (Fig. 5) that we mapped supports their interpretation.

The shallow crest of the southern Astrid Ridge has a flat basement surface that is incised by narrow (10–20 km wide) grabens (basement type 9; Figs. 4, 12). The northern graben is reflected in seabed topography, whereas the southern one is completely covered by sediments. The western slope of the southern Astrid Ridge is locally faulted and passes westward into the flat-lying basement of the Lazarev Sea, whereas the eastern slope is smooth but at a distance of 50–100 km from the ridge axis drops abruptly toward the Riiser-Larsen Sea basin (Figs. 4, 5, 12). The basement of the southern Astrid Ridge (especially the crest and the eastern slope) is characterized by internal reflectors conformable with its surface (Fig. 12).

From the steep base of the southern Astrid Ridge the basement shallows to the east and becomes a 40-km-wide and more than 120-km-long horst (basement type 10), characterized by a rough surface (Figs. 4, 5, 13). Eastward, the basement deepens to 7–9 km b.s.l. (Fig. 4), and beneath the middle continental rise is marked by a strong continuous irregular reflector (basement type 11; Fig. 13, eastern part). Landward, beneath the upper continental rise, only segments of the basement reflector can be seen due to strong multiple reflections.

Magnetic anomalies

General description

All marine and airborne magnetic surveys performed in the Riiser-Larsen Sea area by expeditions of different countries (see inset in Fig. 1) were merged into a composite magnetic map (Fig. 14). The map was constructed using the geophysical image processing and visualization package of

the Geological Survey of Canada (Verhoef et al. 1996); adjustment, leveling and gridding procedures are described by Golynsky et al. (2002).

Magnetic anomaly pattern mostly corresponds with the types of acoustic basement identified in the Riiser-Larsen Sea area (Fig. 14). A compact group of broad high-amplitude (400–600 nT) positive anomalies is associated with the western foot of the southern and northern Astrid Ridge, and a large area farther to the north and northwest. The southern Astrid Ridge has relatively shallow basement and is mostly non-magnetic, whereas the western slope and the northern part of the Astrid Ridge has deeper basement with intense magnetic anomalies. The magnetic pattern in the western Riiser-Larsen Sea (area of basement types 5 and 6) displays distinct linearity, which gradually fades away toward the east and southeast where the anomaly amplitudes increase (Fig. 14). The eastern and southern Riiser-Larsen Sea (basement types 3, 4a and 11) is generally characterized by a quiet magnetic field with only a few local positive anomalies.

The shelf and continental slope of the Riiser-Larsen Sea between 10°E and 32°E is marked by notably high-amplitude positive long-wavelength magnetic anomalies, which connect with the short-wavelength magnetic anomalies of the southern Gunnerus Ridge (Fig. 14). The anomaly is locally over 100 km in width and up to 500–600 nT in amplitude and represents a part of a regionally extensive (almost 2,000 km long) magnetic feature that continues uninterrupted (except for a breach at the eastern base of the Gunnerus Ridge) to the Cosmonaut Sea shelf east of the study area (Golynsky et al. 2002, 2007). Between 16°E and 18°E the anomaly has a horn-like northward bifurcation that broadly corresponds to the basement horst (basement type 10; Fig. 14). The northern Gunnerus Ridge and the Kainanmaru Bank appear nonmagnetic.

Seafloor spreading anomalies

Earlier marine and airborne studies showed distinct magnetic lineations in the western Riiser-Larsen Sea (Bergh 1977, 1987; Roeser et al. 1966; Joket et al. 2003). The

Fig. 10 Seismic section showing basement morphology of the western Riiser-Larsen Sea and the northern Astrid Ridge (basement types 6 and 8)

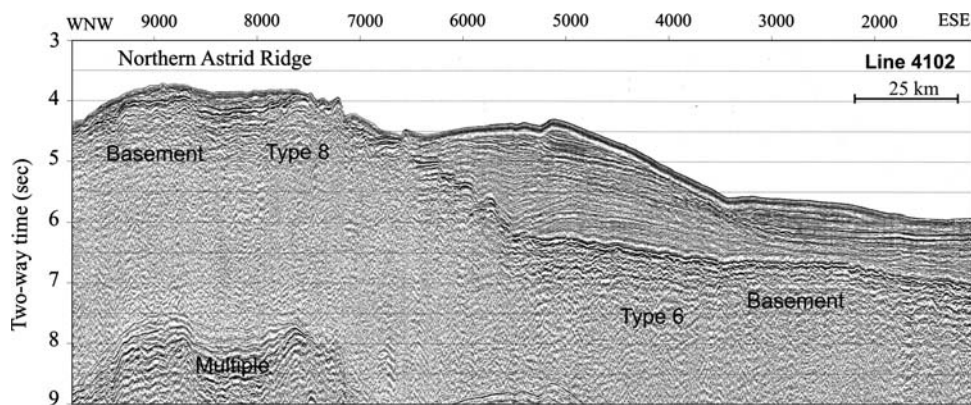


Fig. 11 Seismic section showing basement morphology of the northern Astrid Ridge (basement type 6 and 8). See Fig. 5 for location

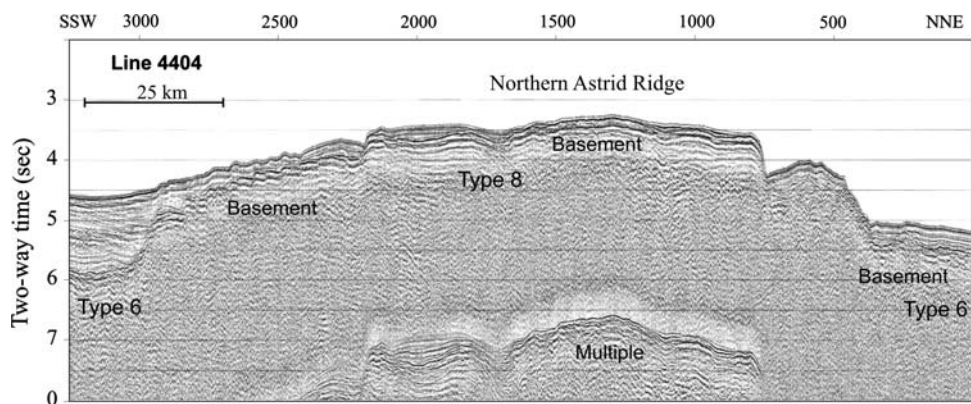
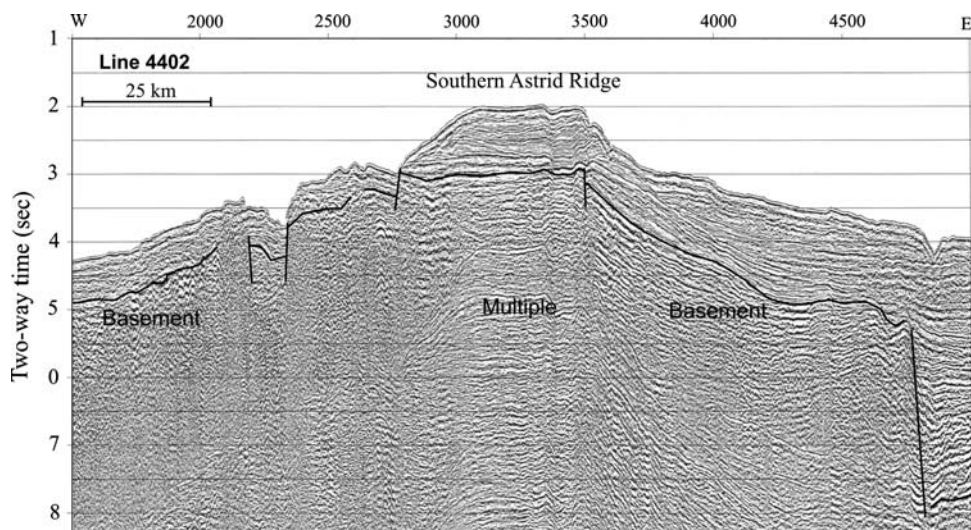


Fig. 12 Seismic section showing basement morphology of the southern Astrid Ridge (basement type 9). See Fig. 5 for location



relatively dense PMGE and PMGE/NPD line coverage in the eastern Riiser-Larsen Sea between 24° and 28°E and north of 66°S now allows identification of a succession of anomalies from M2 to M16 with offsets of 60–70 km along the transform faults (Fig. 15a). Half spreading rates that we calculate for this succession range from 1.2 to 2.2 cm/a (Fig. 15a) and generally agree with those defined by Roeser et al. (1996) and Jokat et al. (2003) for the western Riiser-Larsen Sea. The magnetic anomaly pattern in the

northeastern part of the Riiser-Larsen Sea is complicated and our identification of magnetic anomalies is tentative (Fig. 15a).

The few magnetic lines acquired by PMGE in the southwestern part of the Riiser-Larsen Sea (between 66°–69°S and 16°–24°E) allow us to speculate about the oldest magnetic lineation and seafloor spreading history in this area. Here, Jokat et al. (2003) identify the oldest anomaly M24n and note that M24n is parallel to the present-

Fig. 13 Seismic section showing basement morphology of the southern Astrid Ridge and the southern Riiser-Larsen Sea (basement types 10 and 11). See Fig. 5 for location

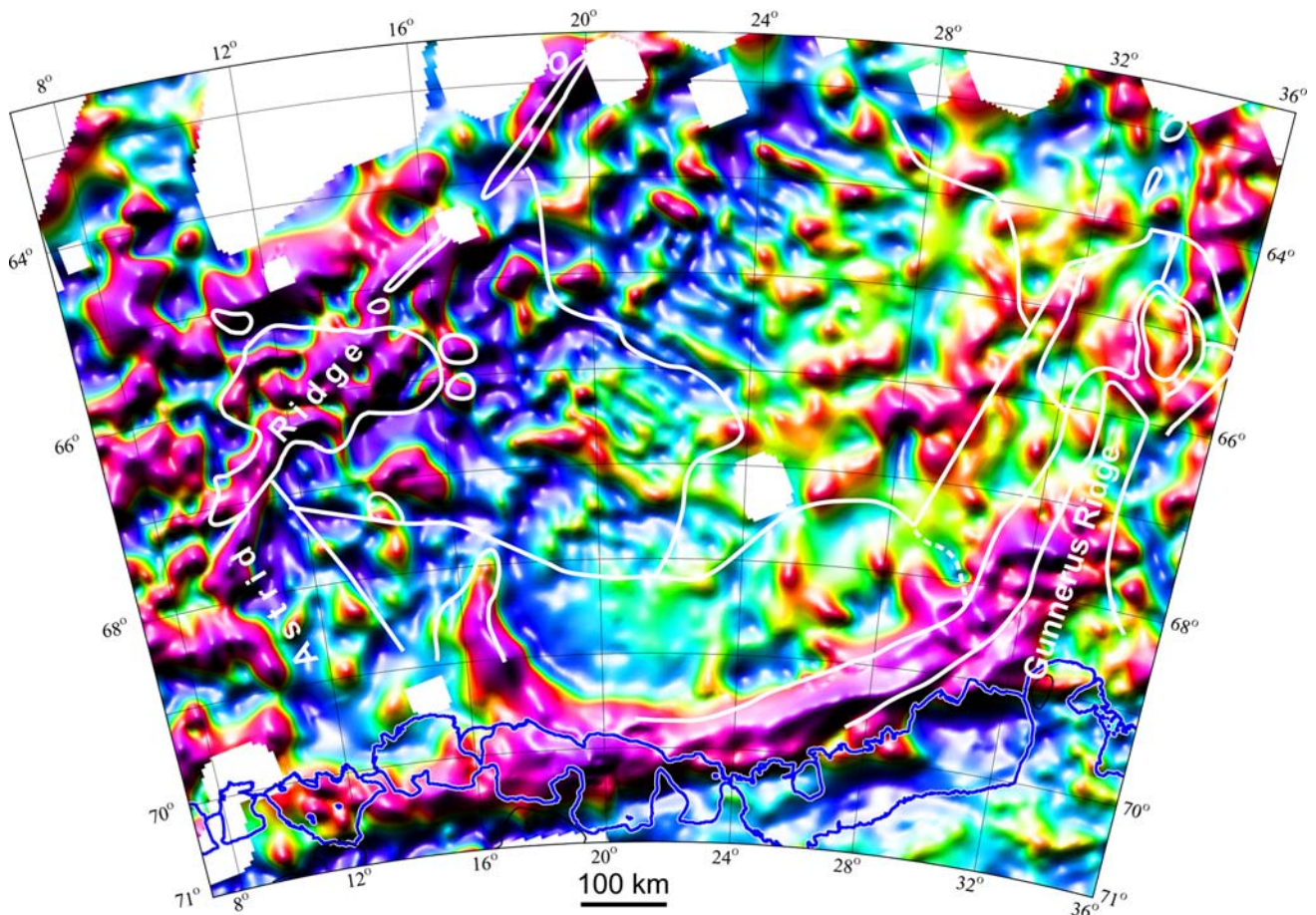
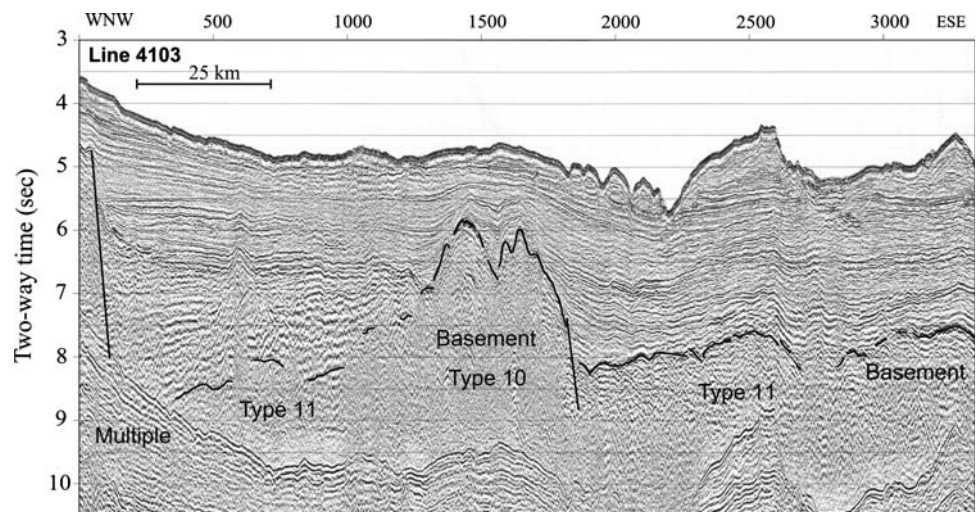


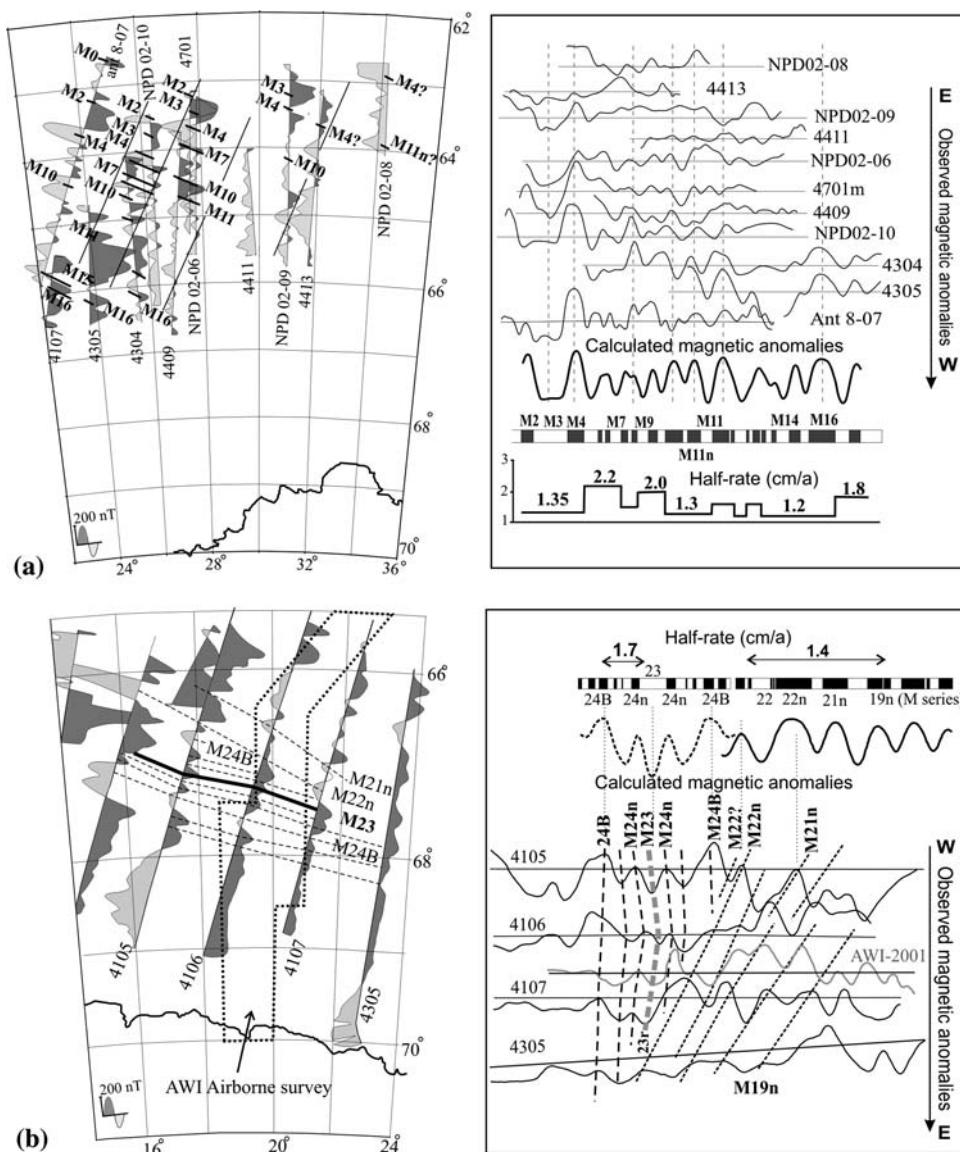
Fig. 14 Magnetic anomaly map of the Riiser-Larsen Sea region. The map has been compiled based on all available magnetic data contributed into “Antarctic Digital Magnetic Anomaly Map”

(ADMAR) Project (Golynsky et al. 2007). White contours show borders of the basement types presented in Fig. 5

day coastline, whereas younger anomalies lie along a different azimuth. PMGE magnetic profile 4105 located west of the AWI survey shows a distinct symmetry of variable-polarity magnetic anomalies relative to a negative central

anomaly (Fig. 15b). The southern part of the symmetric succession can be correlated eastward, whereas the northern part of the succession cannot because it is truncated by the younger lineation M22–M22n (Fig. 15b). Jokát et al.

Fig. 15 Magnetic anomalies and sea-floor spreading models in the eastern (a) and southwestern (b) Riiser-Larsen Sea. On a selected fracture zones (interpreted from satellite-derived gravity anomalies) are shown. The line AWI-2001 on b is the center line of the AWI survey extracted from Jokat et al. (2003). The spreading models are based on a 1 km thick magnetized source layer at 7 km depth, using the reversal timescale of Cande and Kent (1995) and present-day magnetization directions



(2003) model the oldest anomaly of the observed obliquely oriented lineation as Chron M22. Using their identification we further note from the new data that the symmetric succession is best correlated with Chrons M24B (~154 Ma) to Chron 24n (~152.2 Ma) and a central anomaly M23 (~151.5 Ma; Fig. 15b). Hence, we interpret the axial negative anomaly (i.e., Chron M23) as that from an extinct spreading ridge. The new interpretation suggests that sea-floor spreading in the southwestern Riiser-Larsen Sea shifted to a new position at about M23 time (~151.5 Ma; Fig. 15b). The abandoned spreading center corresponds to a shallow (about 100 m depth) depression (barely detectable in the smooth basement surface), and is also marked by gravity low of up to 5 mGal, which is a characteristic feature of the extinct mid-ocean ridges (e.g., Osler and Louden 1995).

Crust thickness and basement velocities

Two representative crustal models derived from seismic and potential field data are shown in Fig. 16. The interpreted MCS lines and sonobuoy refraction velocities were used to estimate crustal geometries and densities for two-dimensional gravity modeling. Only the high-amplitude inshore magnetic anomaly was modeled (Fig. 16a). The models include three crustal layers (sedimentary, upper crust and lower crust) and the upper mantle. The upper and lower crust together is referred in this study to as the crystalline crust. Density estimates were derived by conversion of seismic velocities using the equation of Ludwig et al. (1971).

The south-north model across the eastern Riiser-Larsen Sea (Fig. 16a) shows a ~35 km thick crust beneath the

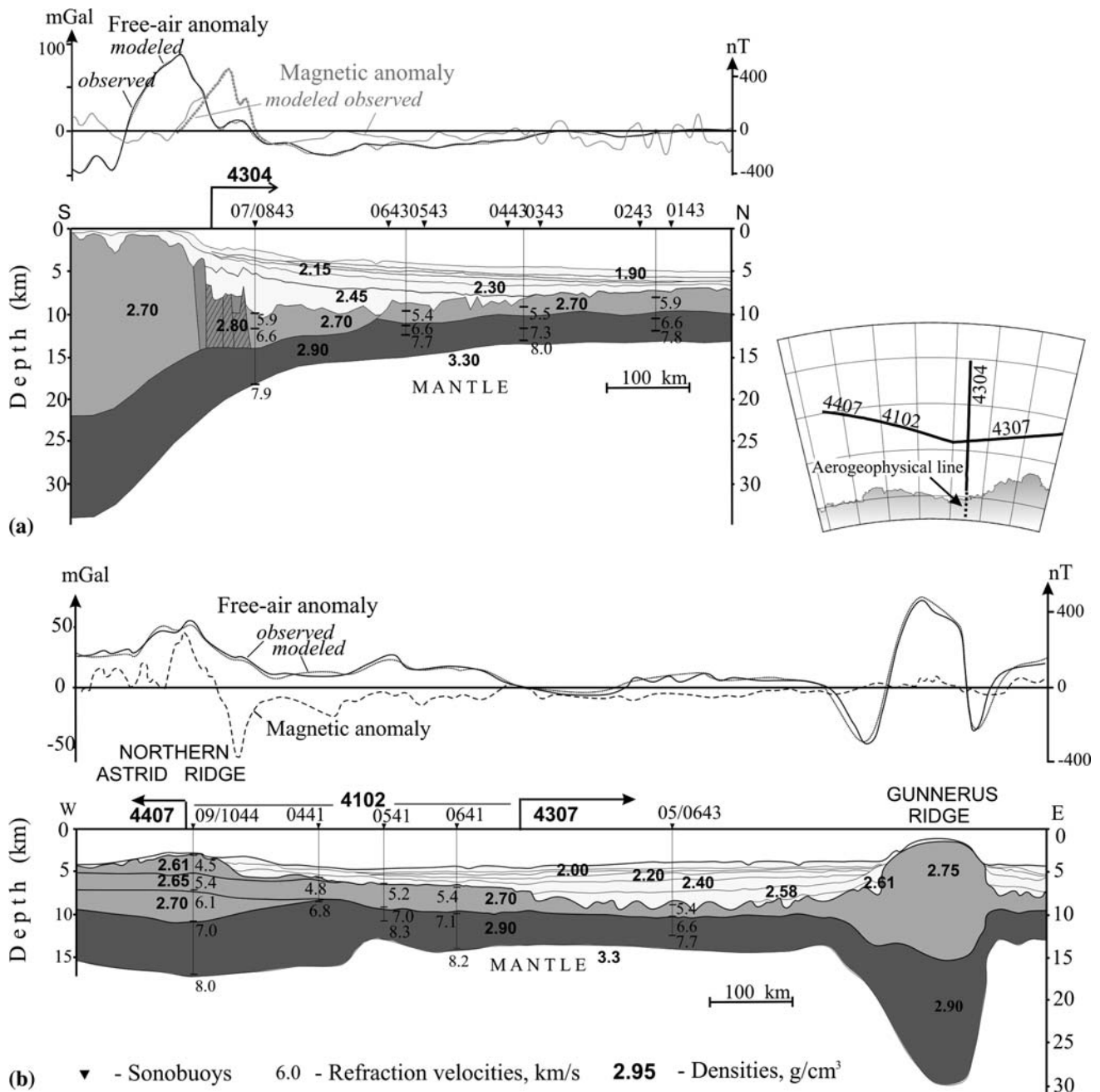


Fig. 16 Crustal models along MCS Line 4304 (a) and MCS Lines 4407-4102-4307 (b). Marine magnetic and gravity observations along Line 4304 were supplemented by airborne potential field data, acquired by PMGE in 1990/1991. Hatched blocks in the southern part

of Line 4304 shows magnetic bodies with induced magnetizations ranging from 0.35 to 0.7 A/m. IGRF-1990 magnetic inclination (62°) and declination (-38°) were used for the modeling

inner continental shelf and a rapid decrease in crustal thickness oceanward to 13–14 km. A 50-km-wide upper crustal block with increased density (2.8 g/cm^3) and magnetization (0.35–0.7 A/m) is needed to model the observed gravity and magnetic anomalies across the continental margin. Farther north a thin crystalline crust is traced under the middle and lower continental rise and into the abyssal plain (areas with basement types 11 and 5; Fig. 5), and the

total crust thickness diminishes to about 8 km mainly due to gradual thinning of the sedimentary layer. The crustal thickness along an east-west transect (Fig. 16b) decreases from 12–14 km in the Astrid Ridge area to about 9.5–10 km in the central Riiser-Larsen Sea and increases to 26–28 km below the Gunnerus Ridge.

Basement velocities along the strike of the Gunnerus Ridge are in the range of 5.8–6.1 km/s; no reliable velocity

data were obtained for the lower crust (Fig. 3). In contrast, the northern Astrid Ridge displays distinct acoustic layering documented by both reflection and refraction seismic data. Seismic velocities increase down-section from 4.5–4.8 km/s at the basement surface to 5.4 km/s at 2.0–2.5 km depth, 6.1 km/s at 4.0 km depth, and 7.0 km/s at 8.0 km depth below the top of the basement (Figs. 3, 16b). The crystalline crust thins towards the east to 12 km, with greatest thinning in the upper crust. Local shallowing of the upper mantle is modeled to the east of the northern Astrid Ridge based on both sonobuoy (site 0541) and gravity data (Fig. 16b). The shallowing coincides with a fracture zone readily seen in the satellite-derived gravity anomaly map (Fig. 2).

From depth-velocity solutions (Fig. 3; sonobuoys 0344 and 0444), the crust at the shallowest part of the southern Astrid Ridge is estimated to reach 17 km thickness, of which only about 1 km is sedimentary cover. The upper crystalline crust is 7 km thick and characterized by the presence of two layers of roughly equal thickness but with different velocity parameters of 5.0–5.6 km/s in the upper layer, and more than 6.0 km/s in the lower one (Fig. 3).

Seismic stratigraphy of the sedimentary section

The seismic stratigraphy of the Riiser-Larsen Sea was not clearly-known from prior investigations due to limited MCS seismic coverage. Hinz and Krause (1982) recognize four seismic sequences (named down-section from WS-1 to WS-4) and two sub-sequence (WS-3A and WS-3B) over a large region from the eastern Weddell Sea to the western Riiser-Larsen Sea. They interpret interfaces bounding these sequences as major erosional unconformities and use the reflection patterns and morphologies of seismic sequences to correlate: sequence WS-4—with a volcanic wedge; the boundary between WS-3A and WS-3B—with major climatic threshold at the Eocene-Oligocene boundary; the base of WS-2—with development of Circum Antarctic Current in the Late Oligocene; and the base of WS-1—with the Middle Eocene Antarctic ice cap formation. Saki et al. (1987) identify two key horizons (shallower Q-1 and deeper Q-2) bounding laminated seismic sequences (I, II and III) and the top of acoustic basement, but do not discuss ages or the origins of the sequences.

Kuvaas et al. (2004a) consider the detailed structure of the upper sedimentary cover in the eastern Riiser-Larsen Sea. They identify a prominent horizon RLS-4 that marks the up-section transition from parallel and subparallel reflectors to a more varied reflection pattern. Kuvaas et al. (2004a) interpret this horizon as the base of glaciomarine deposits. Number “4” for the horizon was designated to show its tentative correlation with previously recognized glacial-onset surface W4 in the Weddell Sea (Kuvaas and

Kristoffersen 1991); however, subsequently Kuvaas et al. (2004b) proposed that glaciomarine deposits in the Riiser-Larsen Sea formed at a later stage than in the Weddell Sea. Solli et al. (2007a) identify additional local horizons RLS-5 and RLS-6 above RLS-4 in the eastern Riiser-Larsen Sea, and note the change of down-slope deposition across them.

In this paper we focus on the preglacial strata (i.e., below horizon RLS-4 of Kuvaas et al. 2004a). We recognize five horizons that separate the Riiser-Larsen Sea basin fill into six seismic units (Figs. 17, 18, 19). The horizons are numbered up-section as 1, 1a, 2, 3 and 4 with letter prefix RLS (Riiser-Larsen Sea) to denote their geographical position. Horizons 2, 3 and 4 can be traced continuously throughout the Indian Ocean sector of the Antarctic and correlated directly with similarly numbered regional unconformities outside the study area (Leitchenkov et al. 2007). Horizons 1 and 1a are detected only in the Riiser-Larsen Sea.

The horizons successively onlap each other and/or onlap the basement surface near areas of high-standing basement, such as the Gunnerus Ridge, the Astrid Ridge and intra-basin highs. Similar onlapping is inferred from seismic data across the continental slope and upper continental rise. In deep-water areas (>4,000 m) the horizons are usually sub-parallel correlative conformities and they onlap or downlap basement at low angles.

The oldest horizon, RLS1, is identified only locally in the sedimentary section beneath the continental slope and the upper continental rise of the southern Riiser-Larsen Sea (Figs. 18, 19, inset). The reflectivity of RLS1 is variable and in the southeastern Riiser-Larsen Sea it generally decreases basinward. The unit between horizon RLS1 and the basement is up to 3.0 km thick and is characterized by undulating, subparallel, semi-continuous low-frequency internal reflectors sloping basinward subparallel to the basement surface (Figs. 18, 19). Only thin accumulations of this unit are inferred in half-grabens along the foot of the Gunnerus Ridge and the Kainanmaru Bank. In the southwestern part of the study area, horizon RLS1 tops the 3.5-km-thick unit (Fig. 19) that Hinz et al. (2004) interpret as a thick pile of volcanic rocks. We attribute this unit to sedimentary or sedimentary-volcanic strata because the unit has distinct stratification and a nonmagnetic nature. The average velocity of the unit between RLS1 and the basement is 4.6 km/s, as calculated from sonobuoy data.

RLS1a onlaps basement at the southwestern and southern boundaries of the basin, along the Gunnerus Ridge and 100–150 km seaward of the RLS1/basement contact (Fig. 19). The unit between horizons RLS1 and RLS1a is characterized by seismic velocities of 4.0–4.2 km/s. It has a maximum thickness of up to 2.5–3.0 km beneath the upper continental rise, and the internal structure displays irregular, low-frequency, discontinuous reflectors.

Fig. 17 Seismic stratigraphy and reflection pattern of sedimentary cover in the eastern Riiser-Larsen Sea (Line 4307). Note limited extent and thickness of the lower unit (i.e. unit below RLS1) as compared to its much greater distribution near the coast (see Fig. 19). See Fig. 1 for line location

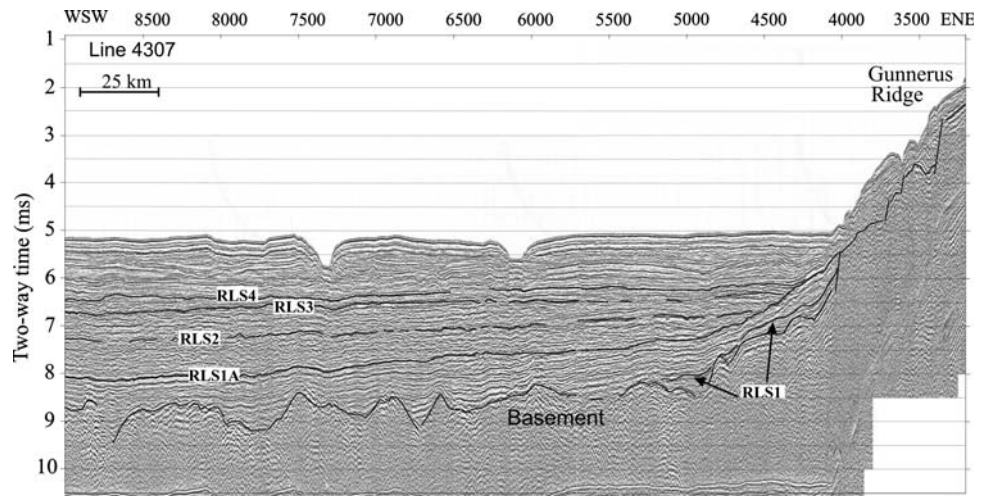
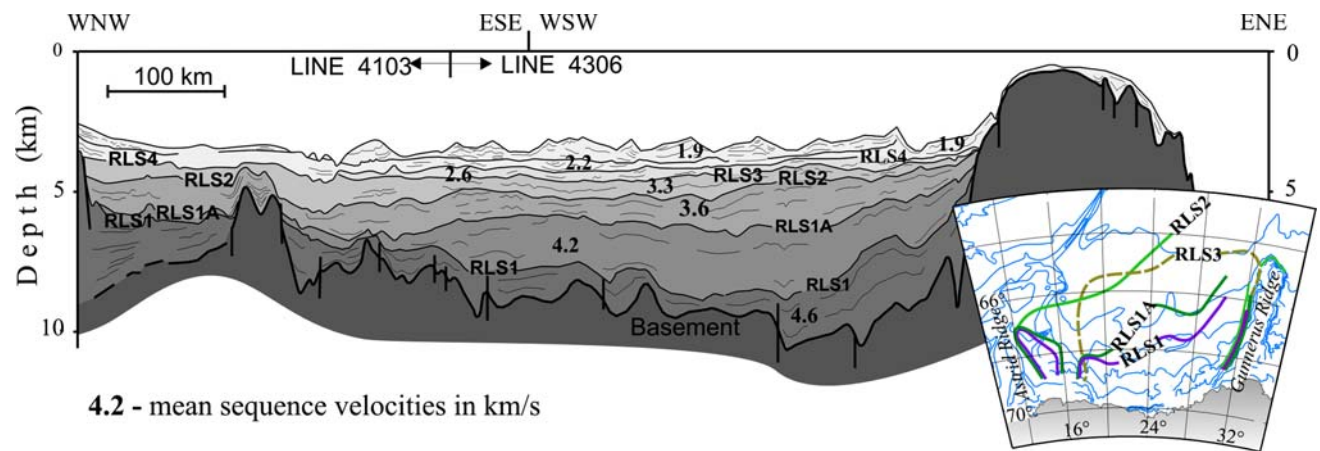
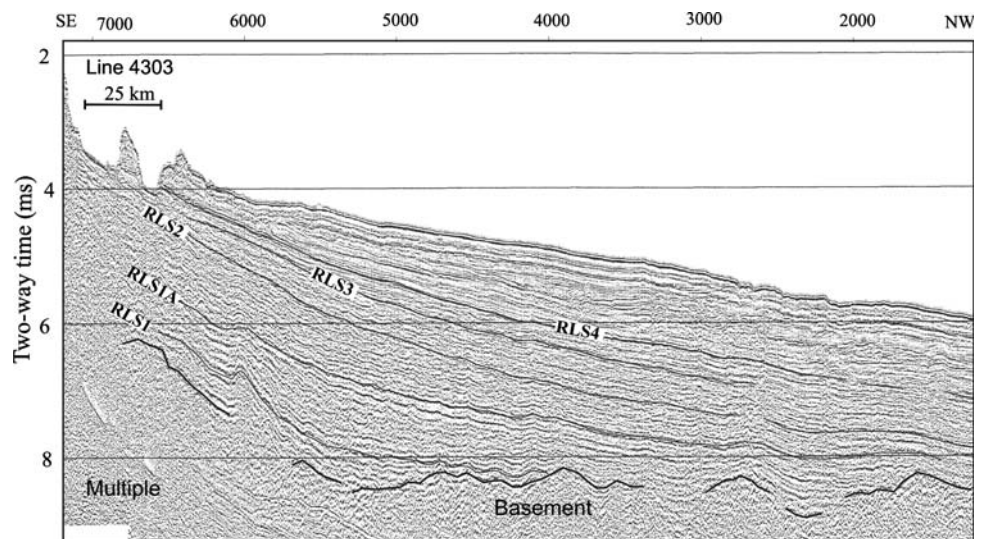


Fig. 18 Seismic stratigraphy and reflection pattern of sedimentary cover across the continental slope and rise of the Riiser-Larsen Sea. See Fig. 1 for the line location



4.2 - mean sequence velocities in km/s

Fig. 19 Seismic stratigraphy of the southern Riiser-Larsen Sea (line drawing along lines 4103 and 4306). Inset shows onlaps/downlaps of horizons RLS1, RLS1a and RLS2 on the basement surface (solid

line), and toplap of horizon RLS3 against horizon RLS4 (dashed line). See Fig. 1 for line location

Horizon RLS2 onlaps basement along the eastern slope of the Astrid Ridge, below the abyssal plain of the north-western Riiser-Larsen Sea, along the southwestern slope of the Gunnerus Ridge, and likely around the outward part of the Kainanmaru Bank (Fig. 19, inset). On MCS lines crossing the northern Gunnerus Ridge, horizon RLS2 onlaps RLS1a (Fig. 17). Horizon RLS2 correlates with the boundary between Units WS3A and WS3B of Hinz and Krause (1982). The unit between horizons RLS2 and RLS1a (or between RLS2 and the basement) ranges in thickness from less than 1.0 km to 1.5–1.6 km and has seismic velocities of 3.6–3.7 km/s.

Horizon RLS3 not only onlaps the older RLS-horizons and basement, but also top-laps against the younger horizon RLS4 (Fig. 19). The unit between RLS2 and RLS3 has seismic velocities 3.0–3.3 km/s and varies in thickness from less than 0.5 km to 0.8–0.9 km, while the overlying unit bounded by RLS3 and RLS4 does not exceed 0.2–0.5 km in thickness and 2.6 km/s in seismic velocities. In a large part of the Riiser-Larsen Sea basin, horizons RLS3 and RLS4 merge in one, and the unit bounded by horizons RLS2 and RLS4 in places exceeds 1 km in thickness. All units in the RLS1a–RLS4 interval are generally characterized by irregular, partly discontinuous internal reflectors that grade into more distinctive acoustic horizons seaward.

Horizon RLS4 is the same horizon as in Kuvaas et al. (2004a) and corresponds to the base of sequence WS-2 of Hinz and Krause (1982). RLS4 is the most prominent of all regional interfaces and marks a significant change in reflection pattern of the sedimentary cover in the study area, and throughout the entire East Antarctic margin (Kuvaas et al. 2004a, b; Leitchenkov et al. 2007). The sequence above horizon RLS4 shows a variety of inferred glaciomarine seismic facies with reflection geometries typical of submarine channel-levee complexes, local debris flows and sediment drifts (Kuvaas et al. 2004a; Solli et al. 2007a). In the Riiser-Larsen Sea, the glaciomarine unit has seismic velocities of 1.8–2.3 km/s. This unit attains maximum thickness in the eastern Riiser-Larsen Sea, where it varies between 1.2 and 1.5 km within drift mounds on the continental rise and is several hundred meters thick in deeply-incised modern channels. In the southern Riiser-Larsen Sea, the thickness of the glaciomarine unit ranges from several hundred meters to 1.0 km beneath the upper continental slope. Detailed structural and stratigraphic characteristics of the unit above horizon RLS4 and insights into the Cenozoic depositional history are given by Kuvaas et al. (2004a) and Solli et al. (2007a, b).

In the absence of rock age information, we infer ages of seismic units in the Riiser-Larsen Sea principally from their onlap geometry onto magnetically dated oceanic basement. We interpret RLS1 as the break-up unconformity based on its deepest stratigraphic position in the

sedimentary section and on comparison with other passive margins (e.g., Kusznieer and Karner 2007). Thus, the strata between RLS1 and the basement are a syn-rift unit. To estimate the age of RLS 1, we extrapolate the age of the oldest oceanic crust (anomaly M24B, ~154 Ma) to the points of onlap of RLS 1 onto the basement (using a half-spreading rate 1.7 km/s; see Fig. 15). Based on this extrapolation RLS1 is 157–158 Ma or older. Horizons RLS1a and RLS2 onlap the oceanic basement around anomalies M18–M19 (143–145 Ma) and M2–M4 (122–126 Ma), respectively and therefore the ages of stratal units below these horizons are younger than the underlying crust formed by spreading.

Horizon 3 onlaps magnetic anomalies 22–24 (49–53 Ma) in the southeastern Indian Ocean off western Wilkes Land and is Early Eocene or younger (Leitchenkov et al. 2007). Horizon 4 is inferred to have an age of Late Eocene to Early Oligocene based on ODP Leg 188 drilling on the continental rise offshore from Prydz Bay (Cooper and O'Brien 2004; Leitchenkov et al. 2007), and is older than Hinz and Krause (1982) postulate for this horizon (named WS-2 in their paper).

The total sediment thickness in the Riiser-Larsen Sea area ranges from 0.1 to 0.5 km on the Gunnerus Ridge, the Kainanmaru Bank and the Astrid Ridge and up to 6.0–7.5 km in depocenters along the southern periphery of the area (Fig. 20). Across the lower continental rise and abyssal plane the thickness is 1.5–2.5 km.

Discussion

Crustal structure and definition of crustal types

The data that we present above indicate considerable variations in crustal thicknesses, acoustic basement characteristics and magnetic pattern, from which we define several tectonic provinces typical of passive continental margins and oceanic basins (e.g., Keen and De Voogd 1988; Menzies et al. 2002). Seismic velocities 5.8–6.1 km/s of the shallowest basement of the Gunnerus Ridge and the Kainanmaru Bank are typical of granitoid rocks of the upper continental crust (Christensen and Mooney 1995). This interpretation is supported by the dredging of inferred autochthonous gneissic debris from the northern Gunnerus Ridge by Saki et al. (1987). Because extensional features such as half-grabens bounded by normal faults are not present along the axial 60–75-km-wide part of the Gunnerus Ridge and on the flat top of the Kainanmaru Bank, we suggest that these areas are underlain by unstretched upper continental crust. Nonetheless, the total thickness of Gunnerus Ridge crust is 25–27 km (Roesser et al. 1996 and this study), which is about 13–15 km thinner than the

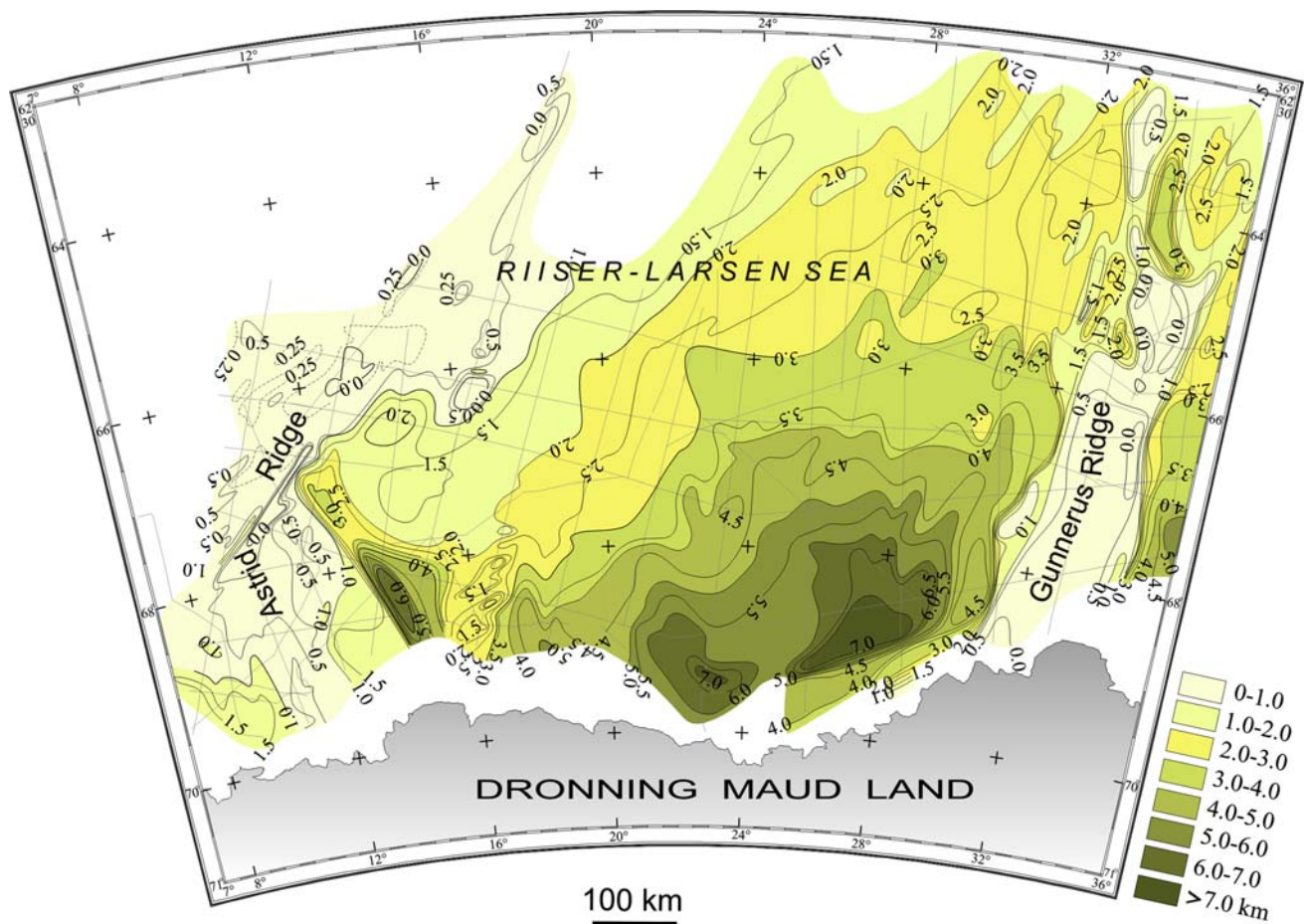


Fig. 20 Total sediment thickness in the Riiser-Larsen Sea area

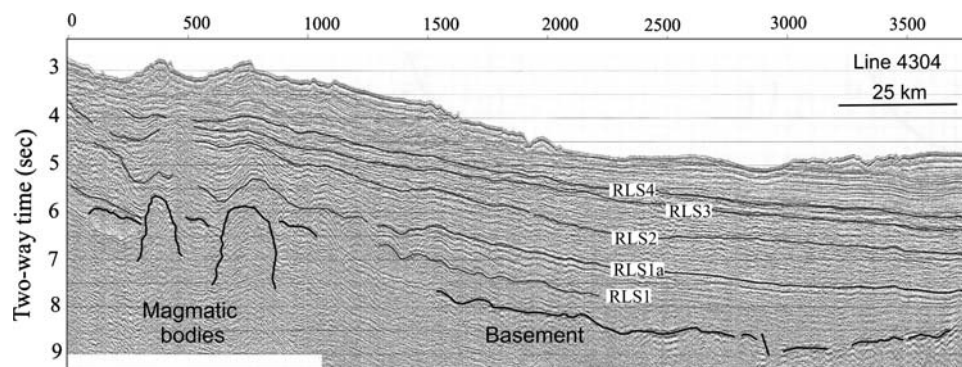
“normal” 40-km-thick continental crust of Dronning Maud Land (Kogan 1972). We believe that crustal thinning can be explained by ductile stretching of lower crust (Kusznir and Karner 2007) or by other geodynamic processes (Artyushkov 1994).

Considerable stretching in the crystalline crust adjacent to the Gunnerus Ridge and the Kainanmaru Bank is indicated by strong block faulting and considerable crustal thinning (Fig. 16b). The zone of crustal thinning coincides with high-amplitude long-wavelength magnetic anomalies (Fig. 14). These anomalies are similar to those observed along margins strongly affected by magmatism (e.g., Holbrook et al. 1994; Gladchenko et al. 1998). For instance the crust of the well-studied eastern North America margin marked by the “East Coast Magnetic Anomaly” is interpreted by Holbrook et al. (1994) to consist of a large thickness (20–25 km) of rift-stage igneous material including volcanic rocks (recognizable on MCS sections as seaward-dipping reflector series), intrusive complexes and underplated high-velocity (7.2–7.5 km/s) magmatic bodies.

MCS data to the east of the southern Astrid Ridge do not reveal a seaward-dipping reflector series typical of volcanic margins (e.g., Planke et al. 2000; Menzies et al. 2002), but magmatic activity is evidenced here by magnetic and gravity anomalies. We interpret the high-density (2.8 g/cm^3) and high-magnetic (magnetization: 0.35–0.7 A/m) block modeled in the upper crust of the southern Riiser-Larsen Sea (Fig. 16a) as composed of mafic intrusions. These intrusions are recognized on seismic sections crossing the margin where they pierce the lower sedimentary cover and are marked by dome-like draping structures (Fig. 18, SP: 6000; Fig. 21, SP: 250–900). Sonobuoy data collected on the upper continental rise do not reveal a high-velocity lower crust typical of igneous underplating (see Fig. 3); however, more detailed studies are needed to characterize the lower crust.

The basement horst between 16°E and 18°E (basement type 10) that is marked by a branch of the inshore magnetic anomaly is in our view an offshoot of the extensive upper crustal intrusive complex inferred along the southern Riiser-Larsen Sea (Fig. 22). This magmatic body may have been emplaced into weakened continental crust during

Fig. 21 - Seismic section (line 4304) across the continental rise of the Riiser-Larsen Sea, showing the basement structure and locations of inferred intrusions that deform the lower sedimentary strata. See Fig. 1 for location



rifting and the primary transverse margin segmentation. The basement horst (magmatic body) lies at the southern continuation of well-recognized oceanic fracture zone (Fig. 22) that likely inherited one of the pre-break-up transfer faults representing this marginal segmentation.

We see a distinct difference between the continental margin of the Lazarev Sea—southern Astrid Ridge and that of the Riiser-Larsen Sea. Hinz et al. (2004) postulate that the former is a typical volcanic margin, whereas we believe the latter is a specific type of rifted margin that is mainly characterized by intrusive emplacements with possibly minor extrusive activity.

Deeply buried (7–8 km b.s.l.) basement in the central Riiser-Larsen Sea (basement type 5 in Fig. 5) has the typical characteristics of oceanic crust (e.g. a highly diffracted surface, lineated magnetic anomalies). Velocities at the basement surface in this area are 5.3–5.6 km/s. These values are similar to those throughout the Indian Ocean sector of the Antarctic (Leitchenkov et al. 2003), and to velocities typical of layer 2B in pre-Oligocene oceanic crust (Carlson 1998). Higher velocities (5.9–6.0 km/s) recorded at two sonobuoy sites (Fig. 5) may be imprecise depth/velocity solutions, or may indicate that layer 2C (sheeted dikes) lies close to the basement surface in those places.

Magnetic anomalies north and north-west of the Kainanmaru Bank are low amplitude and of indeterminate strike. The main reason for denoting this area as oceanic-crust is the contrasting ridge/trough basement morphology (Fig. 8) typical of oceanic fracture zones (Fig. 22). However, a continuation of submerged slices of broken continental crust northward of the Gunnerus Ridge-Kainanmaru Bank spur is certainly not excluded by available evidence.

The Astrid Ridge is interpreted as a volcanic entity (Bergh 1987; Roeser et al. 1986; Hinz et al. 2004). Roeser et al. (1996) postulate that the northern Astrid Ridge is underlain by oceanic crust while Hinz et al. (2004) note that “the Astrid Ridge occupies the transition from Late Jurassic-aged oceanic crust of the Riiser-Larsen Sea in the

east to the volcanic-prone continental margin of the Lazarev Sea in the west”. Hinz et al. (2004) conclude that to the east of the northern Astrid Ridge, volcanic flows (stratified basement in our interpretation) overlap the oceanic crust. We believe that the northern Astrid Ridge formed by excess melt which was produced within the Astrid Fracture Zone when the offset spreading centers were active. This melt caused prolific volcanism that built up the central part of the ridge and also caused formation of thickened oceanic crust to the east (Fig. 22). The boundary between oceanic crust and the rifted volcanic margin of the southern Astrid Ridge cannot be distinguished based on current data. We believe this boundary is represented by the southwestern extension of the Astrid Fracture Zone, which is documented by ridge/trench morphology between 67°S and 67.5°S (Fig. 5, 22).

Many researchers note the practical difficulty of recognizing distinctive properties of oceanic crust and stretched continental crust (e.g., Keen and De Voogd 1988; Leitchenkov et al. 2003; Kuszniir and Karner 2007) and this is also true for the Riiser-Larsen Sea, where geophysical lines are still sparse. Hence, we designate a “zone of uncertainty” for the nature of crust (blank strip in Fig. 22) between the seaward edge of what we believe is highly attenuated continental crust and the landward edge of unequivocal oceanic crust.

The landward edge of oceanic crust in the southern Riiser-Larsen Sea is defined by the southern extent of the basement type 5 (Fig. 5) whose oceanic character is established by identified magnetic anomalies (Jokat et al. 2003; Fig. 22). Our interpretation of the magmatic body at 15–17°E as representative of the rifted margin implies that the “zone of uncertainty” has a minimum width here (Fig. 22). In the southeastern Riiser-Larsen Sea, between 22°E and 28°E, magnetic anomalies are not identified and the position of the oldest oceanic crust can only be presumed from basement velocities 5.4–5.6 km/s (Fig. 5) which are typical of layer 2B (Carlson 1998). The flat-lying basement of type 11 (Fig. 5) that covers the upper continental rise is characterized by two sonobuoy velocities of 6.0 km/s that we attribute to continental-type crust. There

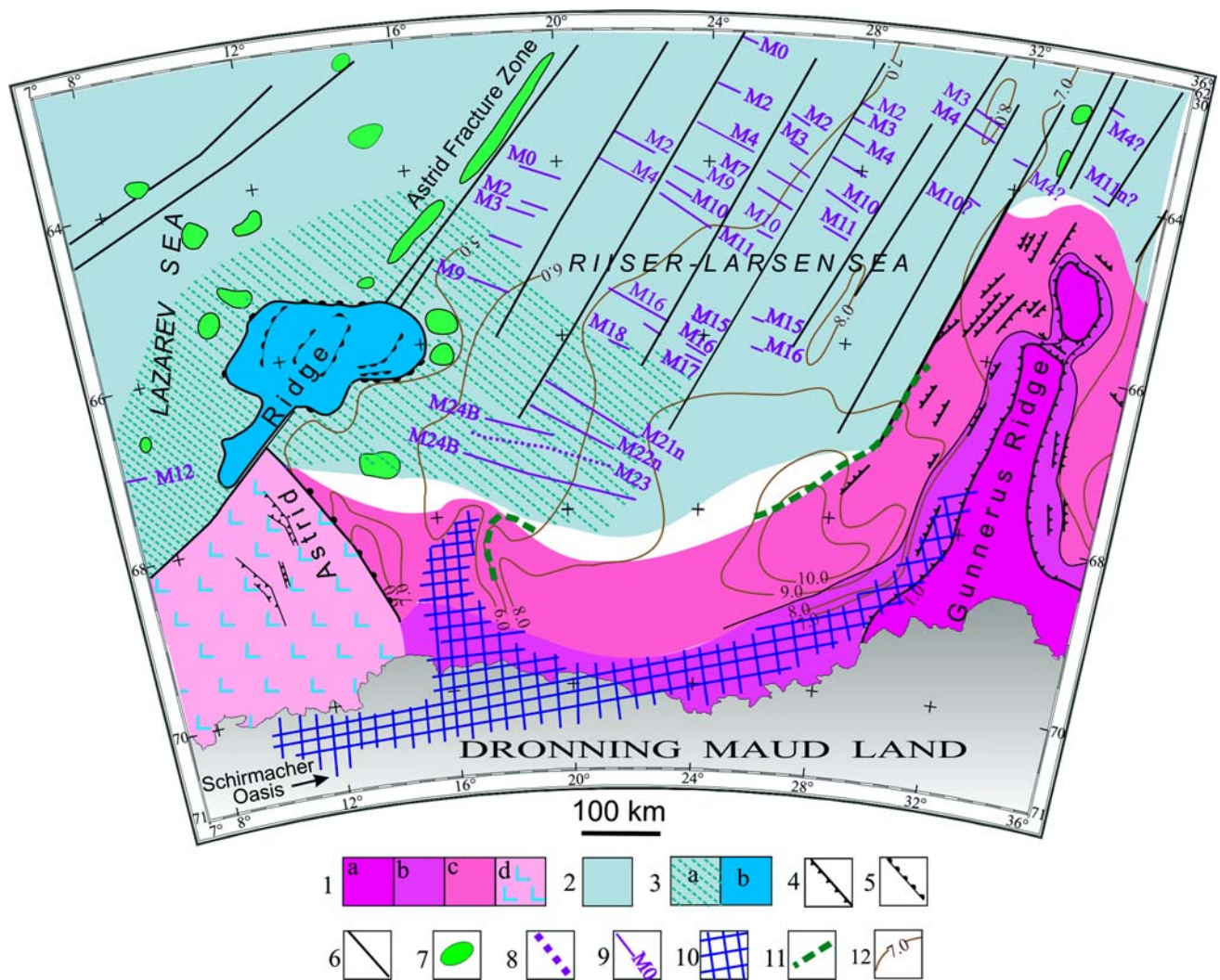


Fig. 22 Tectonic provinces in the Riiser-Larsen Sea area. 1) continental crust: (a) basically not stretched (though moderately submerged and thinned) in the submarine offshoot of the East Antarctic Shield; (b) increasingly stretched oceanward in the shoulder of the marginal rift; (c) strongly stretched in the deep part of the marginal rift; (d) undifferentiated by degree of stretching below the volcanic unit; 2) oceanic crust with normal (5–7 km) thickness of crystalline crustal layer; 3) oceanic crust with thickened (a: 7–10 km;

b: 12–14 km) crystalline crustal layer; 4) normal faults in the basement; 5) inferred volcanic escarpments; 6) oceanic fracture zones; 7) local basement ridges and isometric highs partly exposed in the sea bottom; 8) extinct spreading ridge; 9) identified magnetic anomalies; 10) proposed magmatic intrusions emplaced into the upper continental crust 11) onlap of horizon RLS1 onto the basement 12) total depths to basement (km). The nature of the crust is uncertain in the blank area

are no other characteristics that allow us to define the boundary between highly stretched continental crust and oceanic crust. The zone of uncertainty that we show here may indeed be wider.

In the eastern Riiser-Larsen Sea, the stretched continental crust can be identified on the west side of the Gunnerus Ridge and the Kainanmaru Bank, where basement is block-faulted over a wide zone (Figs. 6, 7, 21). This zone is marked by a regional free-air gravity low of up to -60 mGal. The western edge of the gravity low corresponds to the change in the magnetic anomaly pattern (Fig. 14) and to the linear divide between seismically

distinct terrains differing in basement surface relief and reflectivity parameters (basement types 3 and 5 in Fig. 5). Based on structural trends (flow-lines) defining the direction of movement between Africa and Antarctica (e.g., Marks and Tikku 2001; König and Jokat 2006), we suggest this linear divide is the strike-slip fault of a sheared margin and the boundary between stretched continental and oceanic crust. North of the Kainanmaru Bank our inferred distinction between different crustal types is based on the change in morphology of acoustic basement (Fig. 5) and on the southern terminations of fracture zones identified on the satellite-derived gravity map (Figs. 2, 22).

The edge of extended continental crust can also be defined using the seaward distribution of the syn-rift unit that is overlain by horizon RLS1 (break-up unconformity). Only few MCS lines in the southeastern Riiser-Larsen Sea allow certain identification of places where RLS1 onlaps the basement (see Figs. 18, 19, inset) but in all cases they are landward of the inboard edge of unequivocal oceanic crust. Onlap of RLS1 onto the basement is plotted on Fig. 22 to define the possible seaward extent of the stretched continental crust.

We subdivide stretched continental crust of the Riiser-Larsen Sea into two zones with different crustal characteristics. In the first zone, the basement steeply descends from 2–3 km to 7–8 km b.s.l. and the crust thickness decreases from ~25 km to 12–14 km (basement type 2); while in the second zone, the basement has a relatively flat surface with minor but still recognizable block-faulted topography at 7–10 km depth b.s.l. (basement types 3, 4a and 11). We refer to these zones as the shoulder and the bottom of the marginal rift, respectively (Fig. 22).

The geometry of the marginal rift, with the northeast-southwest trend along the Gunnerus Ridge-Kainanmaru Bank and with the east-west trend along the southern Riiser-Larsen Sea, shows a typical dog-leg pattern common in initial lithospheric rupture followed by continental break-up (Lowell 1990). This dog-leg rift pattern is probably a part of triple junction, the third branch of which can be marked by a prominent negative anomaly (–40 mG) reaching 50–75 km in width and more than 500 km in length and extending from the southeastern Riiser-Larsen Sea south-southeastward to approximately 73–74°S (Fig. 2, Inset).

Timing of major events

Break-up of central Gondwana is commonly associated with emplacement of the Karoo mantle plume that caused lithospheric extension (White and McKenzie 1989). This resulted in extensive magmatism on the conjugate margins of Africa and Antarctica between 200 Ma and 170 Ma which culminated about 182–183 Ma (e.g., Spaeth and Schull 1987; White and McKenzie 1989; Cox 1992; Grantham 1996; Duncan et al. 1997). The age of rifting and igneous activity in the Riiser-Larsen Sea is still unclear. Seismic lines 4303 and 4304 (Figs. 18, 21) show that intrusions deform the syn-rift unit in the lower part of the sedimentary section, suggesting that they were intruded in the early rift stage. The only evidence of intrusive activity in the Riiser-Larsen Sea area comes from Jurassic dolerite dyke swarms that are mapped onshore in the Schirmacher Oasis (situated within the high-amplitude inshore magnetic anomaly; Fig. 22) and in the

Petermann Range (situated southward of magnetic anomaly). ^{40}Ar – ^{39}Ar incremental-heating ages for dykes from the Petermann Range average 183 ± 2 Ma (Rolf et al. 2005), which agrees with the peak of Karoo-type igneous activity. A Sm–Nd whole rock age of 171 Ma from a dolerite dyke at Schirmacher Oasis (Belyatskii et al. 2006) helps set the age range of 170–183 Ma for the emplacement of plume-related plutonic rocks presumably causative of the inshore magnetic anomaly in eastern Dronning Maud Land.

Hinz and Krause (1982) correlate the formation of the volcanic margin in the Weddell Sea and Lazarev Sea with the Early-Middle Jurassic rocks of western Dronning Maud Land. Cox (1992) proposed two successive orthogonally-oriented stages of rifting/break-up between Africa and Antarctica and interpreted volcanic wedges in the Weddell Sea and in the Lazarev Sea as occurring in older and younger stages, accordingly. The older stage in the Cox interpretation corresponds with the widespread Karoo magmatism while the younger may have started at about 170 Ma. Hinz et al. (2004) detect two volcanic wedges in the Lazarev Sea and infer that they correspond with two episodes of transient volcanism. They believe that the seaward wedge and the volcanic edifice forming the Astrid Ridge were emplaced in the Early Cretaceous coevally with the formation of the volcanic Mozambique Ridge.

Jokat et al. (2003) identify anomaly M12 (~134 Ma) immediately north of the seaward wedge and conclude that “the Astrid Ridge marked a continental margin from M24N to approximately M12 [time]”. In the Jokat et al. (2003) and König and Jokat (2006) models, prior to M12 time the Mozambique Ridge had not yet separated from Dronning Maud Land to form the Lazarev Sea ocean basin. Alternatively, we suggest that an ocean basin did exist prior to M12 time, that anomaly M12 formed along the new plate boundary, and that the volcanic rocks which constructed the Mozambique Ridge also covered the older ocean basin crust. This scenario would equate the period of rifting in the Lazarev Sea and the Astrid Ridge with that in the Riiser-Larsen Sea, and would reconcile the different ages of rifting postulated in the Jokat et al. (2003) model. Their interpretation implies that the African plate did not act as a single plate before M12 time. Such assumption is not required in our model implying that the northern Astrid Ridge may have formed initially in Middle-Upper Jurassic (155–160 Ma) time when the oldest spreading between Africa and Antarctica commenced. Furthermore, the ridge may have been fully emplaced by Early Cretaceous (M12 time) when the volcanic edifice of the Mozambique Ridge started to move away from Antarctica along the Astrid Fracture Zone (König and Jokat 2006).

Conclusions

- (1) The study area includes the Riiser-Larsen Sea and two submarine ridges (the Gunnerus and Astrid Ridges) that bound the sea on the east and west. The geophysical data collected here allow mapping of the major crustal features and tectonic provinces. Eleven types of acoustic basement are identified in the Riiser-Larsen Sea area based on depth, morphology and seismic pattern. These basement types are partly consistent with the signature of the magnetic anomaly field that provides useful supplementary information on magmatic and seafloor spreading history of the study area.
- (2) In addition to previously recognized sea-floor spreading magnetic anomalies in the western Riiser-Larsen Sea, we identify a similar linear succession from M2 to M16 in the eastern Riiser-Larsen Sea between 24° and 28°E. Half spreading rates for this succession range from 1.2 to 2.2 cm/a. Magnetic data in the southwestern Riiser-Larsen Sea also reveal a symmetric magnetic anomaly succession from Chron M24B (~154 Ma) to 24n (~152.2 Ma) with a central anomaly M23 (~151.5 Ma). This symmetric succession is obliquely truncated by younger lineations M22–M22n. From this observation, we believe that at about M23 time the ridge jumped to a new position.
- (3) The crustal thickness in the study area varies from about 35 km below the Riiser-Larsen shelf to 26–28 km below the Gunnerus Ridge, 12–17 km below the Astrid Ridge and 9.5–10 km beneath the ocean basin. In the upper crust of the inshore zone (marked by high-amplitude magnetic anomaly), a 50-km-wide block with increased density (2.8 g/cm³) and magnetization (0.35–0.7 a/m) is modeled from potential field data and interpreted as a zone of mafic intrusions.
- (4) Five seismic horizons labeled up-section as RLS1, RLS1a, RLS2, RLS3 and RLS4 bound six seismic units in the sedimentary cover of the Riiser-Larsen Sea. The ages of most seismic units are derived from their onlap geometry onto magnetically dated oceanic basement in the Riiser-Larsen Sea and in the other parts of the southern Indian Ocean. We interpret horizon RLS1 as the break-up unconformity, which is about 157–158 Ma years old. Horizon RLS1a, RL2 and RLS3 onlap the oceanic basement around anomalies M18–M19 (143–145 Ma), M2–M4 (122–126 Ma) and 22–24 (49–53 Ma), respectively and therefore the ages of units lying between these horizons are not older than dated spreading crust. The upper horizon RLS4 is the most prominent regional interface that marks a significant change in

reflection pattern of the sedimentary cover, and RLS 4 is interpreted as glacial-onset unconformity of Late Eocene to Early Oligocene in age.

- (5) The continental margin in the southern Riiser-Larsen Sea is inferred to be affected by intensive pre-breakup magmatism, but in contrast to the margin situated to the west this was dominated by plutonic activity with possibly only minor extrusive component. In the western Riiser-Larsen Sea, intrusive rocks were emplaced into thinned continental crust in the early rift stage (possibly between 170 and 183 Ma). In the southern Riiser-Larsen Sea the seaward edge of stretched and attenuated continental crust and the landward extent of unequivocal oceanic crust are mapped using structural and geophysical characteristics. However, there is a zone where the nature of crust is unclear. In the eastern Riiser-Larsen Sea the boundary between oceanic and stretched continental crust is more definite and we interpret this boundary as a strike-slip fault along a sheared margin. We also propose that the southern Astrid Ridge formed during rifting in the Riiser-Larsen Sea, whereas the northern Astrid Ridge may have formed initially when earliest spreading between Africa and Antarctica commenced.

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