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Formation of the Transantarctic Mountains related to extension of the West Antarctic Rift system

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ABSTRACT

The Transantarctic Mountains are a major rift-related mountain belt bisecting the Antarctic continent. The range is located on the tectonic boundary between non-cratonic West and cratonic East Antarctica. Formation of the mountain range and a possible relation with the West Antarctic Rift system are unclear. In this study, we find a new explanation for uplift of the Transantarctic Mountains and suggest a relation between uplift of the range, formation of a small crustal root, depression of the hinterland Wilkes Basin, and the formation of the adjacent West Antarctic Rift system. Numerical models show that upon extension of the Antarctic lithosphere, the West Antarctic Rift system is formed on the tectonic boundary between East and West Antarctica. Convergence of crustal material results in crustal thickening and uplift of the Earth's surface on the cratonic side of this rift zone, and formation of the Transantarctic Mountains. Some models predict a depression in the hinterland of the mountain range at the location of the Wilkes Basin. These models predict that the Wilkes Basin is a non-extensional basin caused by flexure of the lithosphere. This study shows that inherited lithosphere structures play an important role in localization of both extensional and convergent deformation.

1. Introduction

The West Antarctic Rift system (WAR) resulted from extension between East and West Antarctica starting in the Cretaceous (see Fig. 1 for tectonic context). Despite numerous studies on plate reconstruction (e.g., Cande et al., 2000), crust and mantle structure (e.g., Bannister et al., 2003; Lawrence et al., 2006a,b,c), and elastic, kinematic and dynamic modeling (e.g., Stern and ten brink, 1989; Lawver and Gahagan, 1994; ten Brink et al., 1997; Huerta and Harry, 2007; Bialas et al., 2007), evolution of the rift system and how it is related to the Transantarctic Mountains (TAM) uplift is unclear. The spatial relationship between the TAM and WAR is obvious (Fig. 1), but timing of TAM uplift and denudation and WAR extension are not well related. Earlier significant extension during the Cretaceous was accompanied by relatively little denudation (Fitzgerald, 1992), while most of the recent denudation that was quite extensive predates minor Cenozoic extension (Karner et al., 2005; Huerta and Harry, 2007). The tectonic evolution of this area is important for the understanding of the climatic and glacial history of the Antarctic continent. For example, the geothermal heat flux plays a role in basal melting of ice sheets (Hulbe and MacAyeal, 1999), and tectonic activity may have implications for Antarctic Ice Sheet dynamics (Van den Berg et al., 2006).

In this study, we address the dynamic evolution of the tectonic boundary between West and East Antarctica with a thermo-mechanical modeling approach. This tectonic boundary is likely an inherited laterally varying lithosphere structure from earlier tectonic events (Kalamarides et al., 1987) that preceded the Cretaceous to Cenozoic extension phase that formed the WAR. Our models show that extension of this laterally varying lithosphere structure has resulted in focused deformation along the inherited tectonic boundary, whereby rifting occurs adjacent to dynamic and kinematic uplift of a mountain belt. We attempt specifically to model the evolution of the TAM and WAR in the region near Ross Island, where a recent seismic study (Lawrence et al., 2006a,b,c) imaged the present crust and mantle structure. The presented models provide new insight in the relation between West Antarctic Rift system evolution, formation of the Wilkes Basin, and uplift of the intervening Transantarctic Mountains.

2. Constraints on the evolution of the West Antarctic Rift system and Transantarctic Mountains

The TAM form the morphological and geological boundary between the East Antarctic Proterozoic craton and the now highly extended West Antarctic lithosphere formed by an assemblage of blocks (Jankowski and Drewry, 1981). The TAM are the largest and longest rift-related mountain belt in the world (Fig. 1); with a length of

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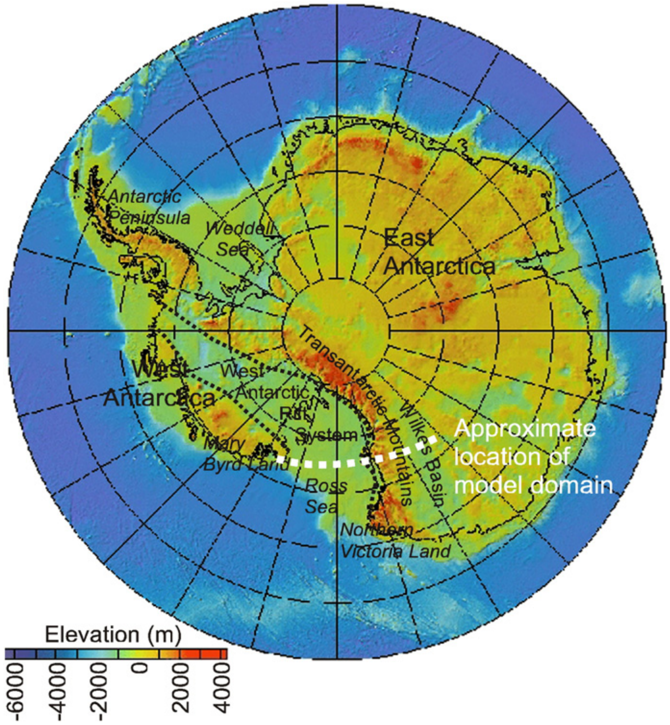

Fig. 1. Subglacial topography and bathymetry map of Antarctica (Lytte et al., 2000). The black dotted lines are the inferred boundary of the West Antarctic Rift system from Rocchi et al. (2002). The Antarctic continent is subdivided into East and West Antarctica by the Transantarctic Mountains. East Antarctica is probably a stable craton. The white dashed line denotes the approximate location of the modeled transect.

almost 3500 km and mountain peaks over 4 km high extending from the Atlantic Ocean side of Antarctica to the Pacific Ocean. Information on the timing of erosion of the TAM is available from fission track analyses. Apatite fission track thermochronology from Victoria Land and the South Pole (Fitzgerald and Gleadow, 1988; Stump and Fitzgerald, 1992; Fitzgerald, 1992; 1994; Balestrieri et al., 1994; Fitzgerald and Stump, 1997) suggests that different blocks of the TAM have experienced separate block movements providing different amounts of uplift and consequent erosion at different times, ranging from early Cretaceous to Cenozoic in southern Victoria Land. In northern Victoria Land, denudation started probably around 50 Ma. Also the amount of denudation appears to vary, ranging from 3 km in the western part of Victoria Land to 10 km in the southeastern coastal areas. Behrendt and Cooper (1991) suggested that uplift of the TAM may have continued until recent times while uplift of the Roberts Ridge may have ended much earlier (Hamilton et al., 2001). Most exhumation of the Transantarctic Mountains however took place after 55 Ma (Fitzgerald, 2002); before 55 Ma the exhumation, that began in the Cretaceous, was relatively limited.

Given that the onset of rifting is poorly constrained, it is less certain how uplift of the TAM is related to extension between East and West Antarctica and formation of the West Antarctic Rift system. The WAR is marked by a topographic trough that is up to 1000 km wide and 3000 km long (Fig. 1) adjacent to the TAM. It is an asymmetric rift zone flanked on one side (Marie Byrd Land) by relief recording up to 3 km of uplift (LeMasurier and Rex, 1989) and on the opposite site by the TAM. Well data in the Ross Sea (e.g., Hannah, 1994) failed to recover sediments that record the onset of rifting, so that more indirect methods have been employed to establish the onset of extension. Plate reconstructions based on magnetic and gravity data suggest that ~180 km of extension in the western Ross Sea embayment occurred during the Eocene–Oligocene (Cande et al., 2000). Early and Late Cenozoic extensional activity has also been identified by Hamilton et al. (2001). Fitzgerald et al. (1986) inferred that rifting started in Late Cretaceous based on the age of rift-related sediments on the conjugate New Zealand margin. Lawver and Gahagan (1994) propose that rifting started in the middle Cretaceous with an early phase of extension ~105 Ma. The locus of rifting appears to have migrated laterally over time through the WAR. For example, the Ross Sea region underwent a multi-phase evolution (Fig. 1). The early, Cretaceous phase of extension affected a wide zone (750–1000 km) between the East Antarctic craton and Marie Byrd Land (Behrendt et al., 1991), while later extension affected a smaller zone adjacent to East Antarctica (Huerta and Harry, 2007).

The Wilkes Basin is located in the interior side of the TAM (Fig. 1), oriented parallel to the TAM with a width of several hundred kilometers. The Wilkes Subglacial Basin is distinguished from surrounding areas by a negative gravity anomaly, low topography, and a negative magnetic anomaly (Drewry, 1976). Interpretations of these anomalies suggest that either 1) A sedimentary layer may fill depressions caused by rifting in the Subglacial Basin (Drewry, 1976; Ferracciolo et al., 2001; Studinger et al., 2004). Rifting resulting in Wilkes Basin formation would have been more recent than major TAM denudation or 2) the Beacon-type sediments may thin toward a flexural trough formed concurrently with TAM uplift (Stern and ten Brink 1989; ten Brink and Stern 1992; ten Brink et al., 1997). In this scenario, Wilkes Basin formation through flexure of the lithosphere would be linked to the Cenozoic uplift of the TAM (Stern and ten Brink, 1989; ten Brink et al., 1997).

The several uplift mechanisms that have been suggested for the formation of the TAM can be divided into different groups: thermally driven uplift, mechanically driven uplift, a combination of these two, or remnants of the collapse of a West Antarctic plateau. Smith and Drewry (1984) attribute uplift of the TAM to heating of the lithosphere by anomalously warm asthenosphere temperatures underneath the area induced by rifting processes. Berg et al. (1989) suggest that magma injections into the middle and lower crust of the rift zone increased temperatures that resulted in thermal expansion of the mountain belt. Rifting is inferred to have been focused by a mantle plume according to Behrendt (1999). In this model, lateral heat transfer from the warm, thin lithosphere beneath the rifted areas affected the cold lithosphere from East Antarctica resulting in the high rift shoulder.

ten Brink and Stern (1992) and Stern and ten Brink (1989) have proposed a flexural origin model for formation of the TAM. In these flexural models, the end of a broken plate is uplifted, forming the TAM, whereby the plate has been broken by a deeply penetrating normal fault. Uplift then results from a combination of processes, such as lateral variations in temperature or density structure, erosion, and footwall uplift. Fitzgerald et al. (1986) suggest a lateral and depth dependent asymmetric rifting model for formation of the TAM/WAR region, in which the TAM is underlain by a shallow lithosphere/asthenosphere boundary, created by a detachment zone dipping beneath the mountain belt. Lawrence et al. (2006a) conclude that uplift can best be explained by a combination of a buoyant thermal load and flexural uplift.

ten Brink et al. (1997) propose that uplift of the TAM is not related to the main phase of extension of WAR formation. This idea is based on data that show that the main phase of extension between New Zealand and West Antarctica predates the main phase of TAM uplift by at least 30 My. Furthermore, the lack of deep extensional basins along much of the TAM (ten Brink et al., 1993) and the coincidence of coeval uplift with plate reorganization in the South Pacific suggest that uplift of the TAM is associated with breaking the lithospheric plate and changing the effective elastic thickness during the Eocene transtensional deformation phase (ten Brink et al., 1997).

An alternative hypothesis concerning the origin of the Transantarctic Mountains (Huerta and Harry, 2007; Huerta, 2007; Bialas et al., 2007) suggests that this mountain range is the abandoned margin of a collapsed Mesozoic West Antarctic plateau. Tectonic reconstructions indicate that West Antarctica has been the site of repeated orogenic phases. It was part of the active margin during Gondwana assembly (Goodge, 2002). Along strike of the pre-stretched Ross Embayment, the Lachlan Fold Belt is located in southeastern Australia; an orogen with thickened crust (e.g., Foster and Gray, 2000). These reconstructions indicate that West Antarctica including the Transantarctic Mountains might have been part of an elevated, thickened crust plateau in the Mesozoic. There is increasing geologic evidence that supports this plateau scenario, including indications of drainage reversal (Huerta, 2007).

3. Constraints on crust and mantle structure from seismic studies

Prior to 2000, little seismic data were available for imaging the crust and mantle beneath the Transantarctic Mountains, and adjacent West Antarctic Rift system. Consequently, the only techniques capable of imaging to mantle depths in the region were 1) continental-scale tomographic images using broadband stations located around the perimeter of Antarctica and 2) limited small-scale broadband analyses using a handful of stations in the Ross Sea region. The continental-scale images (Danesi and Morelli, 2001; Ritzwoller et al., 2001; Sieminski et al., 2003; Morelli and Danesi, 2004) with ~1000 km scale resolution delineated a relatively sharp boundary beneath the Transantarctic Mountains dividing the seismically fast East Antarctica from the seismically slow West Antarctica. The small-scale studies used surface waves (Bannister et al., 2000) and receiver functions (Bannister et al., 2003; Kanao et al., 2002) from a few seismometers to reveal a 10–15 km thicker crust and >5% seismically faster mantle beneath the TAM than beneath the westernmost WAR.

From 2000 to 2003 the Transantarctic Mountain Seismic Experiment (TAMSEIS) placed 41 broadband sensors in and around the TAM in the Ross Sea region, providing data for 3D imaging of the crust and mantle. From this experiment, it is known that the crust thickness increases from ~20 km in West Antarctica to ~38 km beneath the crest of the TAM and thins back to ~35 km beneath East Antarctica (Lawrence et al., 2006a). Surface wave (Lawrence et al., 2006b) and body wave (Watson et al., 2006) tomography illuminate a slow mantle seismic velocity anomaly beneath Ross Island and the WAR and fast seismic velocity anomalies beneath East Antarctica, with the transition between the two occurring beneath the TAM at 50–150 km inland from the coast. Seismic attenuation (or energy loss) is low in East Antarctica and increases markedly toward the Ross Sea between 50 and 150 km from the coast. The reciprocal velocity and attenuation are consistent with a ~300 K temperature increase, which would cause a ~1% density drop over the same scale (Lawrence et al., 2006c). The observed low attenuation and fast seismic velocities from 50 to 300 km depth beneath East Antarctica indicate that East Antarctica has little-to-no asthenosphere, which suggests that a high-viscosity lithosphere may extend to ~250 km beneath East Antarctica. Conversely, the high attenuation and slow seismic velocities beneath the Ross Sea region of the WAR suggest the presence of a hot, well-developed asthenosphere and only a thin lithosphere.

4. A rift-related model for the Transantarctic Mountains

In this paper we use numerical models to study deformation of the tectonic boundary between West and East Antarctica and adjacent regions. Specifically, we examine how the inherited structure and boundary conditions affect the spatial and temporal evolution of crustal thickening and thinning.

Based on geologic and geophysical constraints mentioned above, a Mesozoic plateau for West Antarctica with thickened crust is used as the initial condition for the model. Isostatic equilibrium in the models then results in an elevated plateau at the onset of extension. In these models, the lithosphere is extended for a finite duration between Late Cretaceous and present day. Because the exact period of extension is still under debate, several models are tested with varying duration of the extensional phase. Other parameters that we tested that are not well constrained are discussed in the following section.

The finite-element model is two-dimensional and represents a vertical transect through the lithosphere. It consists of three layers (upper crust, lower crust and mantle lithosphere) that are composed of triangular elements. The Lagrangian approach is used, and the code solves for viscoelastic deformation of the lithosphere with a parameterization of plastic behavior. A more detailed description can be found in Van Wijk et al. (2001) or Van Wijk and Cloetingh (2002). A Maxwell body is used to describe the viscoelastic behavior of the lithosphere, and a power-law relation between stress and strain is used (Ranalli, 1995). In the models, density is dependent on temperature following a linear equation of state. Temperature is calculated at each time step using the heat flow equation. Temperatures are calculated on the same grid as the velocity field. Advection of heat is accounted for by the nodal displacements. In all experiments, the grid is non-equidistant with the smaller element spacing toward the center of the domain (~1.6 km grid spacing in center of domain, increasing to ~40 km at the sides of the domain).

Brittle behavior is included in the models through a parameterization. This deformation mechanism is active when a critical stress level is reached. In the model, the critical stress level is defined by the Mohr–Coulomb criterion. Stresses are adjusted for every time step when the criterion is reached; frictional sliding and fault movement are not explicitly described by this criterion.

The initial model configuration is shown in Fig. 2. The domain is divided into an upper crust that has a granite composition, a lower crust that has a plagioclase composition, and an olivine-based mantle (Table 1). Orientation of the domain is chosen to be across-strike of the Transantarctic Mountains and West Antarctic Rift system (Fig. 1). Prior to model runs a state of thermal and isostatic equilibrium is assumed. This results in initial surface topography, as shown in Figs. 2 and 3. West Antarctica is an elevated plateau at the onset of extension. The initial temperature structure (Fig. 2) is calculated using the boundary conditions (T=0 °C at the surface and 1300 °C at the base, zero heat flow through the sides) and heat production in the crust as a heat source. The initial temperature structure adopted in these models (Fig. 2) shows that the East Antarctic lithosphere is, at the onset of extension, thicker than the West Antarctic lithosphere. This reflects the cratonic nature of East Antarctica (Fitzgerald, 2002). Extension of the lithosphere is modeled by pulling the left and right sides of the model domain outward.

We assume in these models that the WAR formed through passive extension (i.e. not induced by a mantle plume) of the lithosphere resulting from extensional plate boundary forces. Antarctica is almost completely surrounded by mid-ocean spreading ridges that have continuously added oceanic lithosphere to the plate. Consequently, the proportion of old oceanic Antarctic lithosphere has increased over time (Sandiford and Coblentz, 1994), changing the potential energy of the plate and creating deviatoric tension. While not explored here, other authors have attributed rifting in Antarctica to mantle plume processes (Weaver et al., 1994; Storey et al., 1999), citing domal uplift and hotspot-like volcanic rocks in Marie Byrd Land (LeMasurier and Landis, 1996), and geochemical similarity between WAR volcanic rocks and hot spot volcanism (Behrendt et al., 1994; Hole and LeMasurier, 1994) as evidence. We adopt the passive rifting model in this study because the relation between the spatial and chronological distribution of plutons and dike swarms on the western Ross Sea shoulder points toward a relation between magmatism, rifting, and plate tectonic processes (Rocchi et al., 2003, 2005).

4.1. Parameter study

A range of initial and boundary conditions were tested to explore their influence on the extensional evolution of the WAR and the vertical tectonic history of the TAM. It is important to test these

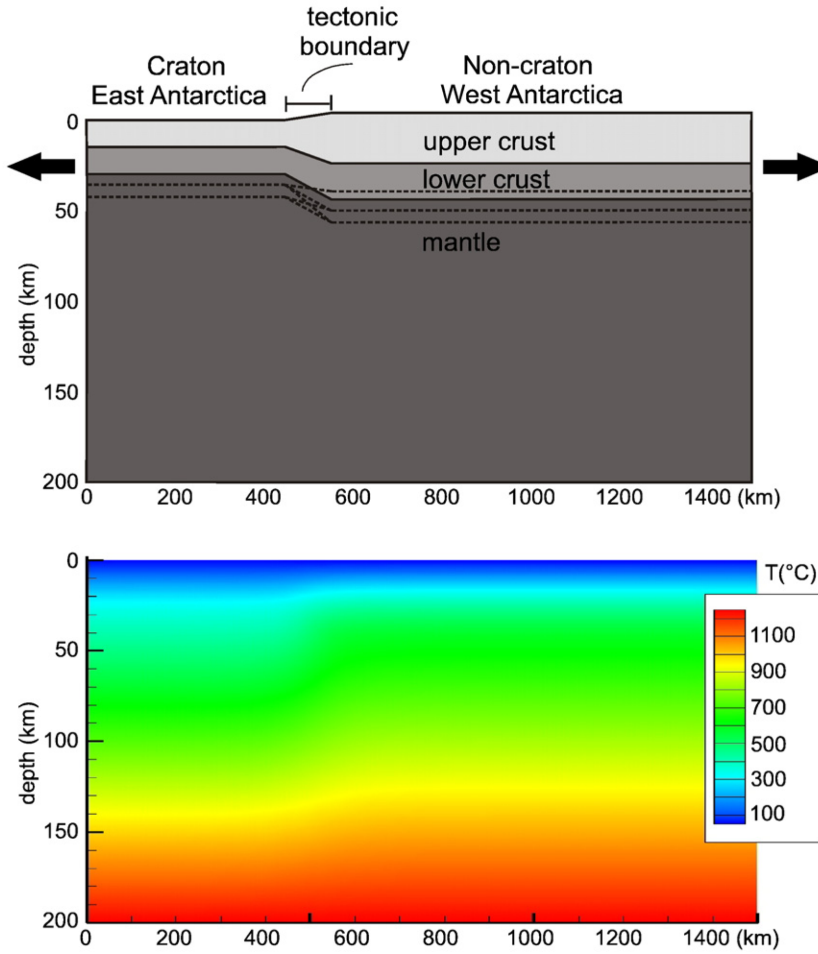

Fig. 2. Sketch of the initial model setup (upper panel) and an example of the initial temperature field (lower panel). The dashed lines indicate the different crustal thickness distributions that are tested. Apart from the Moho depth and topography, we varied the extension rate and extension duration.

parameters; they are not well constrained and small variations could affect the tectonic development of the models. The initial, pre-rift strength difference between East and West Antarctica is varied in the different tests to obtain insight in its role during deformation of the lithosphere. This is done by varying the initial crustal thickness in West and East Antarctica. The nature and origin of the pre-rift boundary between East and West Antarctica are debated. Bannister et al. (2003) suggest a mid-crustal compositional change beneath the TAM. Kalamarides et al. (1987) find isotopic differences in the lower crust across this boundary. Prior to the episode of extension related to Gondwana breakup, this boundary was likely altered by 1) repeated rifting, 2) Mid-Late Cambrian transpression and back-arc volcanism, and 3) Paleozoic terrane accretion (Elliot, 1975; Dalziel and Elliot, 1982; Dalziel, 1992). The initial crustal thickness in West Antarctica was varied between 37 and 46 km, the crustal thickness in East Antarctica was varied between 30 and 34 km. The East Antarctic craton is modeled with a relatively thick lithosphere. The thickness of the lithosphere was varied between 160 and 200 km (Fig. 2 for example setup). Extension of the lithosphere is modeled by pulling the left and right sides of the model domain outward. As the rate of extension is poorly constrained, we varied this parameter between 1, 6 and 10 mm/yr total extension velocity for different test scenarios. An extension rate of ~7 mm/yr corresponds to the estimated 180 km of extension during the Eocene/Oligocene suggested by Cande et al. (2000). Also the duration of the extensional phases is not well constrained, and we varied this parameter as well (from 30 to 85 My).

Within the parameter space tested, all simulations predict a consistent tectonic evolution. In Fig. 3 a selection of the models is shown, where the crustal thickness and extension velocity are varied over relatively small ranges. When the lithosphere is extended a rift basin is formed in West Antarctica adjacent to the tectonic boundary with East Antarctica. A mountain range is formed adjacent to the rift zone, and several of the models predict a surface depression in the hinterland of the mountain range, possibly corresponding to the Wilkes Basin. This consistent evolution was found in all models that we tested, illustrating that the formation of a rift zone (WAR) adjacent to uplift and formation of a mountain range (TAM) is a robust result of the pre-existing asymmetric lithosphere. Some of the models predict

Table 1
Thermal and rheological parameters used in the models

Parameter	Upper crust	Lower crust	Mantle lithosphere
Density [kg/m^3]	2700	2800	3300
Heat production [W/m^3]	$1\cdot10^{-6}$	–	–
Specific heat [J/kg/K]	1050	1050	1050
Conductivity [W/m/K]	2.6	2.6	3.1
Thermal expansion [K^{-1}]	$1\cdot10^{-5}$	$1\cdot10^{-5}$	$1\cdot10^{-5}$
Bulk modulus [Pa]	$3.3\cdot10^{10}$	$3.3\cdot10^{10}$	$4.7\cdot10^{10}$
Shear modulus [Pa]	$5.0\cdot10^{10}$	$2.0\cdot10^{10}$	$1.0\cdot10^{10}$
Activation energy Q [kJ/mol]	186.5	238	510
Material constant A [Pa^{-n}/s]	$3.16\cdot10^{-26}$	$2.08\cdot10^{-23}$	$7.0\cdot10^{-14}$
Power law exponent n	3.3	3.2	3.0
Dilatation angle	0°	0°	0°
Cohesion factor [Pa]	$20\cdot10^6$	$20\cdot10^6$	$20\cdot10^6$
Friction angle	30°	30°	30°

From Tsenn and Carter (1987), Kirby and Kronenberg (1987) and Ranalli (1995).

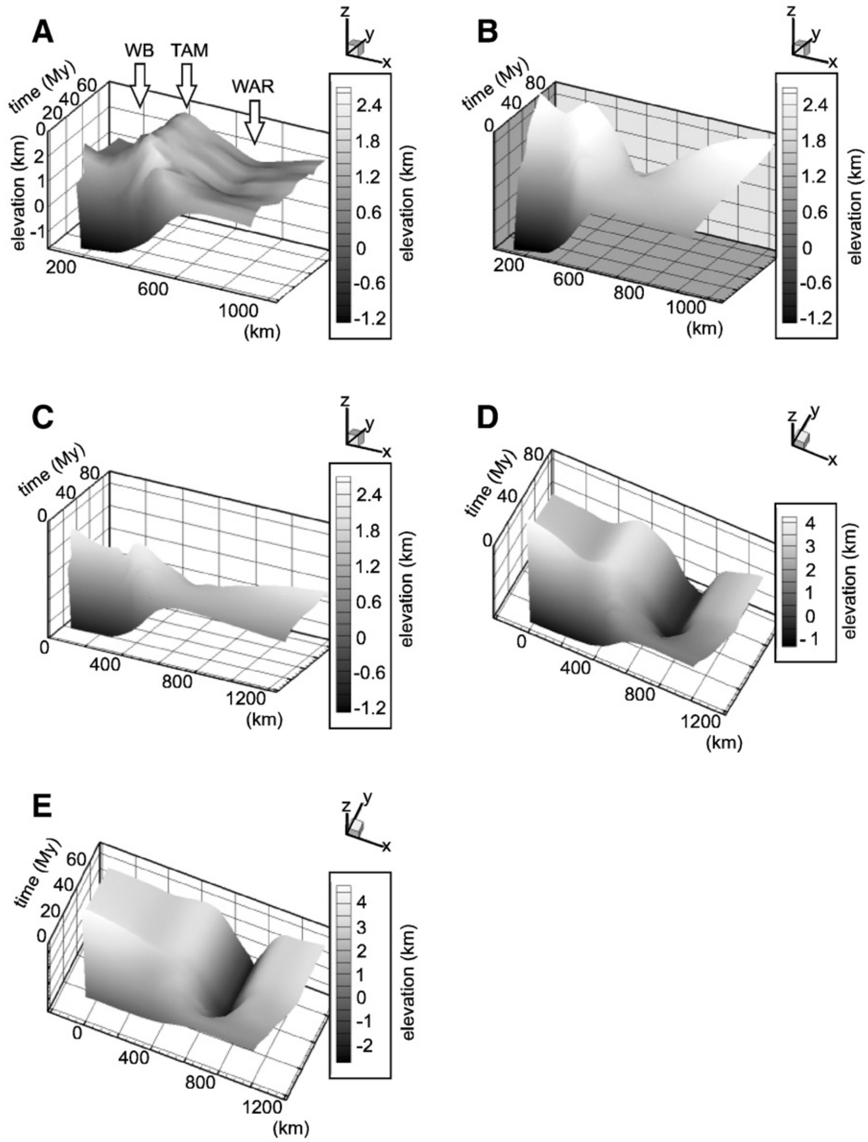

Fig. 3. Sensitivity study. Evolution of surface topography for models A–E. The initial topography (0 My, the onset of extension) reflects the thicker crust in West Antarctica, and results from the initial model condition of isostatic equilibrium. A) Crustal thickness of West Antarctica is 44 km, of East Antarctica 30 km, extension velocity is 1 cm/yr. B) Crustal thickness of West Antarctica is 44 km, of East Antarctica 30 km, extension velocity is 6 mm/yr. C) Crustal thickness of West Antarctica is 44 km, of East Antarctica 34 km, extension velocity is 1 cm/yr. D) Crustal thickness of West Antarctica is 42 km, of East Antarctica 30 km, extension velocity is 1 cm/yr. E) Crustal thickness of West Antarctica is 37 km, of East Antarctica 30 km, extension velocity is 1 cm/yr. WB = Wilkes Basin.

the surface depression related to the Wilkes Basin. In models C and E, the development of this depression is either weak or absent. The magnitude of uplift of the TAM and depression of the WAR vary between the different models, and are dependent on 1) the strength difference between East and West Antarctica, 2) the extension rate and 3) duration of extension. For example, model D shows TAM uplift to 3209 m, while the same experiment with a 10 My longer rift duration predicts uplift to 3911 m (not shown). The Wilkes Basin is predicted to be a flexural basin; no crustal thinning is predicted here. The models predict a clear relation between formation of this flexural low and uplift of the TAM.

4.2. Model for WAR, TAM and Wilkes Basin

The model that best matches the tectonic evolution is Model B (Fig. 3B), with a West Antarctica crustal thickness of 44 km, East Antarctica crustal thickness of 30 km, an extension velocity of 6 mm/yr, and a lithosphere thickness of 180 km below East Antarctica and 150 km below west Antarctica (Figs. 4, 5 and 6). In this model, extension is first widely distributed but localizes later strongly close to the Transantarctic Mountains. The Wilkes Basin is weakly developed through a flexural low behind the mountain range. Model B differs from Model A (Fig. 3A) in that the extensional deformation does not localize as well in Model A. In Model A, however, the Wilkes Basin is more strongly developed.

4.2.1. Strength of the lithosphere: location of rifting

Lateral variations in the strength of the lithosphere control the location of rift basin formation in extensional environments, by focusing deformation in the weakest zones. The tectonic boundary between the weak West Antarctic lithosphere and strong East Antarctic lithosphere may induce such focusing. All the models that we tested predict the formation of a rift basin (WAR) on the non-craton side of the tectonic discontinuity. Calculations of the total lithospheric strength show that this is indeed the weakest zone in the model domain (Fig. 4). The total strength of the lithosphere is obtained by integrating the stress field over the thickness of the lithosphere. We calculated the total lithospheric strength S using $S=\int\sigma(z)dz$ (Ranalli,

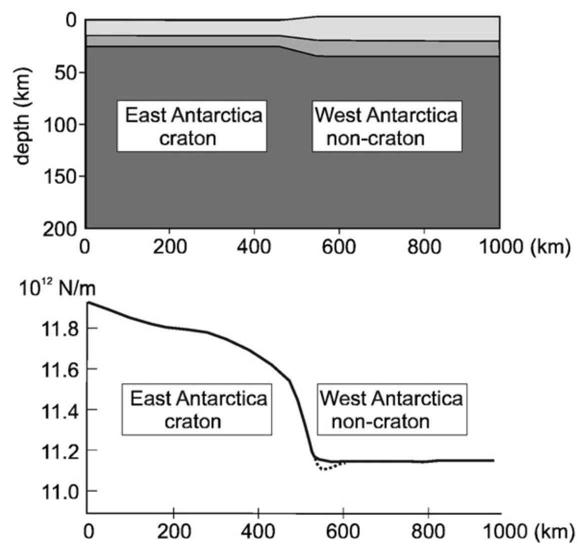

Fig. 4. Total strength of the lithosphere (lower panel) and setup for model A. The region of lowest lithospheric strength at the beginning of the rift phase (black solid line) is very broad. During extension, a small weak zone develops (dotted line, after 5 My) where extensional deformation will focus further.

1995). Fig. 4 shows the strength (lower panel) and initial (pre-rift) lithosphere structure (upper panel). The total lithospheric strength is lowest within the West Antarctic plate proximal to the boundary, which is exactly where the maximum extensional thinning of the crust is observed and predicted (Fig. 6). Thickening of the crust and uplift of the TAM (Fig. 3) is predicted to occur above the tectonic boundary where the strength of the lithosphere increases rapidly toward East Antarctica.

4.2.2. Thermal evolution

After 30–85 My of extension, all models predict that the eastern, cratonic part of Antarctica remains characterized by a thick and cold lithosphere. Fig. 5 shows the thermal results of models A and B. In East Antarctica, the lithosphere is not affected by the rift phase. The lithosphere thickness (defined here by an isotherm) thins rapidly under the western TAM and is thinnest (~95 km in model A, for 1100 °C isotherm) in the part of the WAR adjacent to the TAM. Away from the TAM, the depth of the WAR lithosphere–asthenosphere boundary increases to about 115 km. Surface heat flow values are low below the Wilkes Basin and high on the western flank of the TAM and in the rift zone. For extension persisting for 80 My with an extension rate of 6 mm/yr (model B), the lithosphere beneath the focused rift zone is only about 50 km thick while it is ~100 km thick lithosphere below West Antarctica farther from TAM.

In all models, the thickness of the lithosphere beneath the WAR is controlled by the duration of extension and the extension rate. The various models result in WAR lithospheric thickness ranging from 45 to 110 km. Regardless of extension rate and duration, the models consistently predict extension localized on the non-craton side of the tectonic boundary and the cratonic lithosphere remains undeformed by the rifting process.

4.2.3. Distribution of crustal thinning and thickening

The upper crust of the rift zone thins most in the area closest to the TAM and decreasingly toward the east, however, a very wide zone (500–1000 km) is affected by the rifting process (Fig. 6A,B). The distributed crustal thinning occurs in both the upper and lower crust. Below Wilkes Basin, no crustal thinning is predicted in any of the models (Fig. 6). There is some decoupling between lower crustal deformation and upper crustal deformation; Fig. 6 shows that lower crustal thinning and thickening are more localized. The total amount of whole crustal thinning in the eastern part of the WAR is about 13.7 km for model A and 25.3 km for model B. The total amount of crustal thinning differs greatly between the different models as a result of varying the rate and duration of extension. The region of localized rifting in the WAR is consistently predicted to be asymmetric, causing most extensional deformation to localize close to the craton border/TAM area.

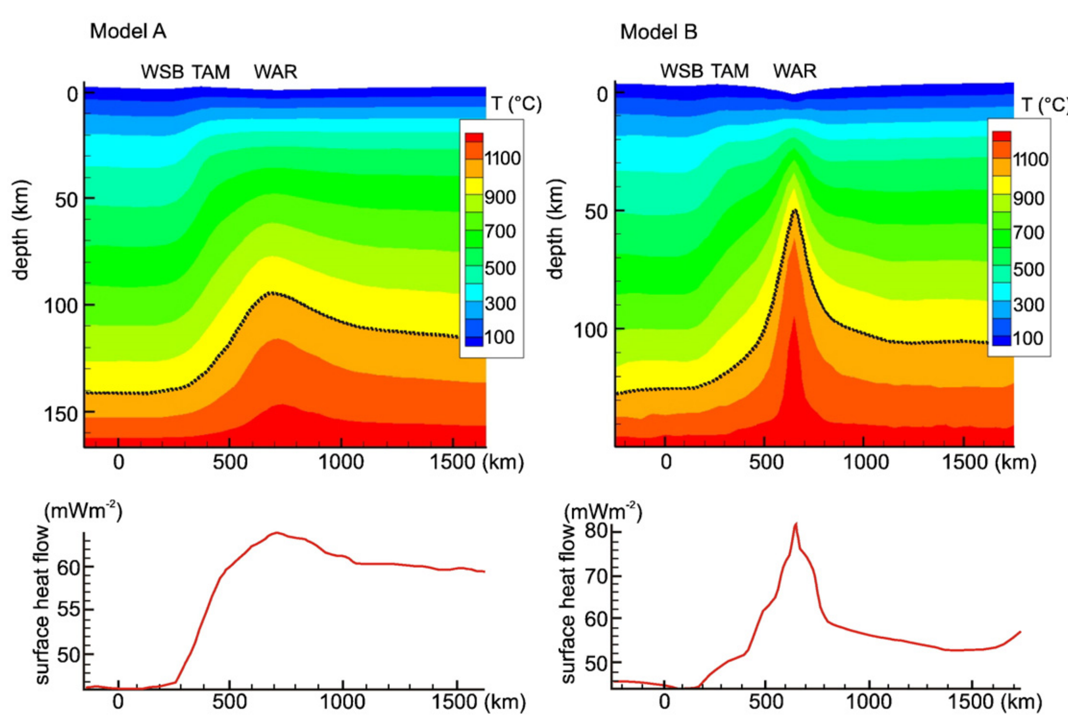

Fig. 5. Predicted temperature field after 30 My of extension (models A and B) and corresponding surface heat flow values. The dotted line indicates the base of the lithosphere, taken as the 1100 °C isotherm. WSB = Wilkes Subglacial Basin, TAM = Transantarctic Mountains, WAR = West Antarctic Rift system.

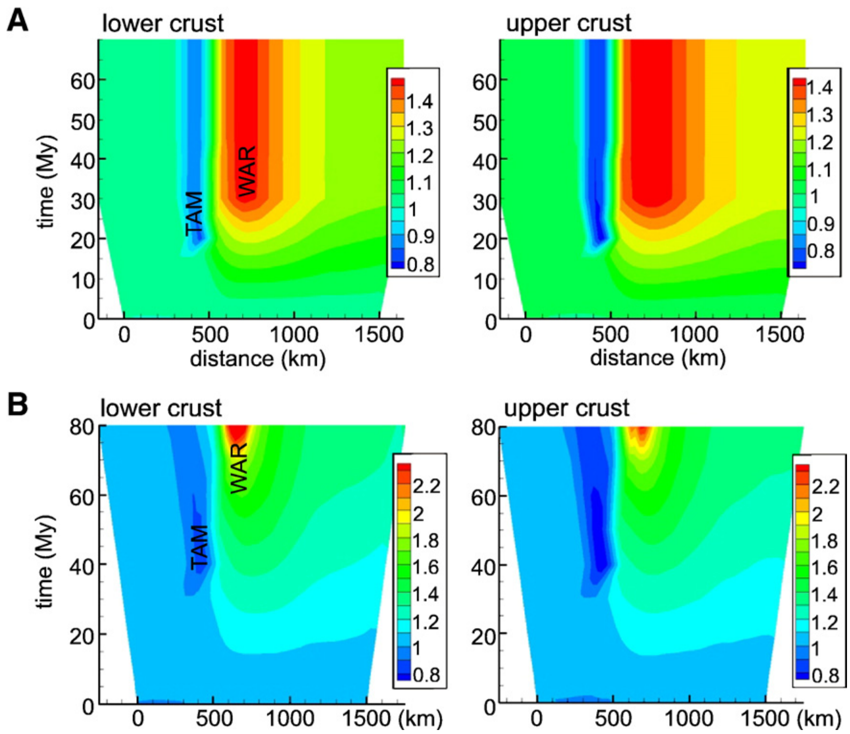

Fig. 6. Evolution of upper and lower crustal thinning. Crustal thinning factors are the ratio between the initial crustal thickness and the current crustal thickness; thinning factors >1 denote crustal thinning, factors <1 denote crustal thickening. A) model A, B) model B. The area where crustal thinning is predicted is very wide; however major thinning occurs in a rather narrow zone close to the TAM. Note the different scales in A) and B).

The majority of crustal thickening occurs in a ~250 km wide zone, resulting in a total of ~8 km of thickening in the various models. The total crustal thickening varies by a few kilometers between the different models as it is dependent on the strength difference across the tectonic boundary between West and East Antarctica, and the amount of extension. Crustal thickening mainly occurs by flow in the middle and lower parts of the crust. During the extensional phase, crustal material is transported away from the rift zone center, but converges at the edge of the craton, due to the great cratonic strength. Analysis of the horizontal component of the velocity field shows a negative velocity gradient in the TAM region meaning that this is a region of convergence. In this area the crust thickens; a crustal root is formed and the surface is uplifted.

Furthermore, Fig. 6 shows that most crustal thickening and uplift of the TAM (Fig. 3) precedes the main phase of rifting and WAR formation. While this pattern is found in all models, the process is particularly pronounced in model B. This prediction is in agreement with data that show that most of the denudation preceded the Cenozoic extension that localized near the mountain range (Huerta and Harry, 2007, and references therein). The Moho extends to greater depths under the TAM during the rift phase, as seen in model A after 15 My of extension (Fig. 6), and is uplifted below the WAR. At the end of the rift phase, a crustal root of several km underlies the mountain range while the crust of the rift zone has been thinned significantly. The amount of crustal thinning depends on model parameters such as extension rate; for example, a ~19 km thick crust is predicted at the end of the rift phase in model B.

4.2.4. Stresses

Fig. 7 shows the predicted horizontal deviatoric stress field in the lithosphere for model A after 15 My of extension. In all models, the strongest layer can be found just below the Moho. The lithosphere has two strong layers; one in the upper crust and one in the upper part of the mantle lithosphere. In West Antarctica the strong part of the upper mantle is confined to a relatively thin layer where stress magnitudes are high. A very thick strong layer is predicted for cratonic East Antarctica but here stress magnitudes are lower. The calculated strength of the lithosphere (Fig. 4) shows that the lithosphere of East Antarctica is considerably stronger than the lithosphere of West Antarctica.

5. Discussion

5.1. Uplift of Transantarctic Mountains

Previous studies (e.g., Smith and Drewry, 1984; Stern and ten Brink, 1989; ten Brink et al., 1997; Studinger et al., 2004; Bialas et al., 2007; Huerta, 2007) have pointed toward three main causes for uplift of the TAM: 1) thermal uplift due to high temperatures at shallow depths, 2) flexural uplift along a broken plate, and 3) remnants of a pre-existing elevated plateau. Our models suggest a source for uplift of the TAM

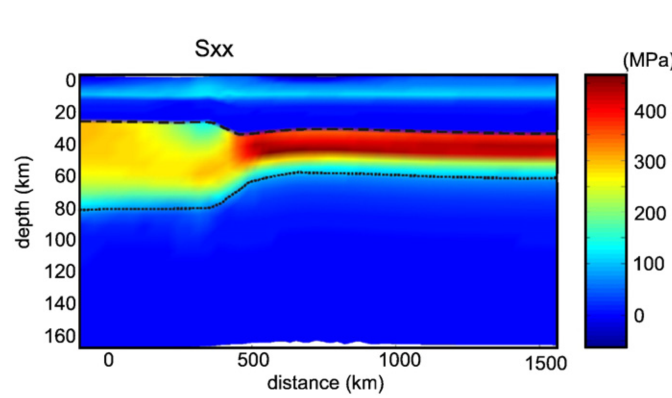

Fig. 7. Horizontal deviatoric stress field in the lithosphere for model A at 15 My. Dashed black line is the Moho. Dotted black line denotes the base of the strong lithospheric layer, corresponding well to the flexural model proposed by Stern and ten Brink (1989).

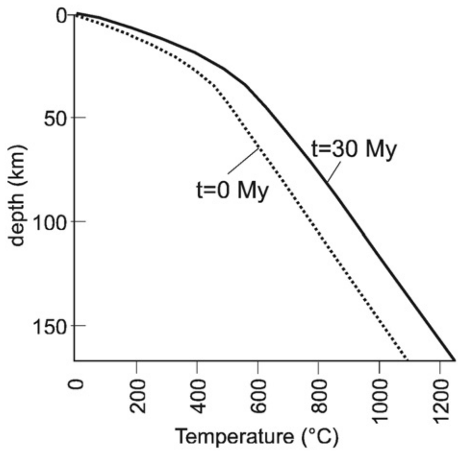

Fig. 8. Geotherm at $t=0$ My (the onset of extension) and after 30 My of extension, of the eastern flank of the TAM.

that was not found in earlier studies: convergence of crustal material that results in thickening of the crust and a small crustal root. Indications for a possible thermal uplift component for the mountains can be found in Figs. 5 and 8. Fig. 5 shows a sharp increase in temperatures below the mountain range with low temperatures under the craton side of the TAM and high temperatures closer to the rift zone. Fig. 8 shows the increase in lithospheric temperature on the eastern flank of the TAM during the rift phase. Thermal uplift related to the rifting event is therefore expected at the eastern front of the TAM, however this would be a transient and would decay with time from maximum extension. Smith and Drewry (1984) proposed a model for thermal uplift of the TAM in which East Antarctica overrides hot oceanic asthenosphere; our models show that high temperatures at shallow depths are expected from the rifting event and do not require additional sources of heat. Thermal buoyancy along with other loads such as erosion, ice sheet effects, and an end-load, has previously been used as the predominant load in flexural modeling of a broken plate configuration (e.g., Stern and ten Brink, 1989; ten Brink et al., 1997). The thick East Antarctic plate/thin West Antarctic plate model used in these flexural calculations is very similar to our model predicted strength distribution (Fig. 5) across the tectonic boundary. A difference between our models and previous flexural models is that the thin crust/weak plate of West Antarctica is, in our models, the result of the rifting phase, instead of a pre-existing condition.

The other main cause of uplift that we found in the models, i.e., a convergence zone on the craton edge, is the result of rift localization adjacent to a craton province. Accumulation of crustal material results in local thickening of the crust some time after the extensional phase has started, but prior to the main phase of rifting and localization (Figs. 6 and 3). Accumulation forms a small crustal root that may explain the 3 to 5 km crustal thickening beneath the TAM, found in the TAMSEIS experiment (Lawrence et al., 2006a). The small crustal root is insufficient to account for the elevation of the TAM; regional unloading and high-temperature low density mantle below the western part of the TAM also contribute to the uplift.

This model result is in agreement with the observed drainage reversal (Huerta, 2007). This model for Transantarctic Mountain origin assumes, like the plateau collapse model (Bialas et al., 2007), that a thickened crust was present below West Antarctica as a result of the orogenic and active margin tectonic history, but differs because additional uplift of the mountains is predicted due to crustal accumulation between the rift zone and cratonic East Antarctica. Such additional uplift due to tectonic forces is not predicted by the plateau collapse model, where the mountain range looses topography during the extensional phase. Another component to contribute to the high elevation of the TAM is isostatic rebound (Stern et al., 2005). Stern et al. (2005) found that isostatic rebound as a response to glacial incisions in the Transantarctic Mountains can account for 2000 m of elevation. The predicted topography of models A and B (around 2.4 km) would, with the isostatic rebound component, match the observed topography of about 4–4.5 km.

5.2. The Wilkes Basin

Prior studies have posited that the Wilkes Subglacial Basin is either a depression caused by rifting (Drewry, 1976), a flexural trough (Stern and ten Brink 1989; ten Brink and Stern 1992; ten Brink et al., 1997), or a structural controlled basin (Studinger et al., 2004). Our models support the flexural trough interpretation. None of our models predict crustal thinning or extension in the TAM hinterland because the strong cratonic lithosphere resists deformation there. Some of the tested models formed surface depressions in the hinterland of the TAM as a result of lithospheric flexure. The magnitude of the predicted surface depression depends on the tectonic evolution of the rift zone and mountain range; it varies in the tested models between ~1 km below sea level (model A) and a local very small depression (model E). Because formation of this flexural depression depends strongly on lateral strength differences between East and West Antarctica, it is likely that lateral variations in the strength difference along strike of the TAM and WAR result in lateral variations in the wavelength and depth of the flexural depression, indicated by observations (Studinger et al., 2004).

5.3. The West Antarctic Rift system

Timing and duration of the rift phase(s) that formed the WAR are not well constrained, and models in which we varied the duration and timing of extension show a wide range in tectonic evolution. A common feature of the models is that the rift zone is always formed at the edge of the craton, with strongest localization close to the craton and wide, distributed rifting farther away. The craton boundary develops to be the weakest location and therefore the preferred place of rift formation with ongoing extension. The cratonic lithosphere remains unaffected by the extensional event.

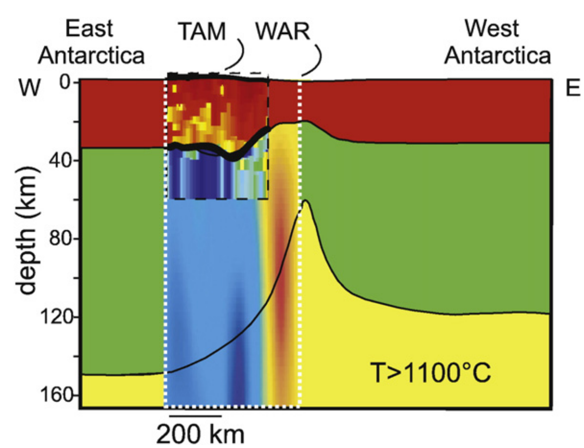

Fig. 9. Interpretation of seismic images of the crust (in black dashed box, Lawrence et al., 2006c) and upper mantle (in white dotted box, Watson et al., 2006), and the result of model B. Fast seismic velocities are blue and slow seismic velocities are red and yellow. Model B is composed of crust (red), cool mantle (green, $T<1100$ °C), and warm mantle (yellow, $T>1100$ °C). This is the interpreted present-day lithosphere structure where the Transantarctic Mountains, West Antarctic rift zone and the Wilkes Basin are all formed as a result of rifting between West and East Antarctica. Characteristic features of this model such as the non-affected craton, the shallow asthenosphere below the rift zone, the slightly thickened crust below the mountains, and the non-rifted hinterland depression are similar to seismically observed structures. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

In the models, we assumed a thick crust prior to rifting in West Antarctica that is thinned to 20 km thickness subsequently during the rift phase. The models thus predict a present-day crustal thickness and structure in West and East Antarctica that is supported by observations (Fig. 9), while the pre-rift configuration that was adopted is very different from the present structure. This is in contrast to flexural models, that start from the present-day lithosphere structure by introducing a broken plate resulting from a rift-related normal fault separating East from West Antarctica. Development of the WAR at the craton edge is then a result of the inherited lithosphere structure in our models, while the WAR is considered a cause for the broken plate structure in some flexural models.

Heat flow data are sparse in the Ross Sea area (Blackman et al., 1987; Berg et al., 1989). The data have an average value of about 70 mW/m^2, and show a general increase in magnitude toward the TAM, reflecting the maximum thinning close to the mountains. Heat flow values in the rift zone predicted by the models vary between different tests, but generally show good agreement (~70–80 mW/m^2) with observations. The models predict very low heat flow values in the Wilkes Basin area (<50 mW/m^2).

6. Uplift of the Transantarctic Mountains: result of rifting at a craton boundary?

Fig. 9 shows the predicted present-day lithosphere structure based on model B. Based on the model results, we interpret the tectonic evolution as follows. Upon extension that started in the Late Cretaceous, the asymmetric pre-rift lithosphere structure has resulted in the formation of an asymmetric, wide rift zone. The cratonic lithosphere is not affected by the rifting event, causing the western side of the rift to end abruptly against the strong lithosphere. A second phase of extension is focused in the Victoria Land Basin, near the East Antarctic craton. Below the eastern flank of the TAM, lithosphere temperatures are elevated, which has contributed to uplift of the TAM. Uplift is further the result of crustal thickening; a crustal root several km thick is formed below the TAM during the extension phase, and of isostatic rebound. Relative timing of uplift in relation to the stretching is still debated; our models suggest that upon initial, moderate extension, the main uplift phase precedes the second localized rift phase. Some authors have suggested two major rift phases in the area, for example, ten Brink et al. (1997) reported Eocene transtensional deformation. Our models predict that such a second rift phases would still focus at the favorable craton edge location.

All models that we tested suggest that uplift of the TAM, extension of West Antarctica and formation of the Wilkes Basin can be tectonically related. Crucial for this finding and for the development of the area is the role played by inherited structures. The inherited tectonic boundary between East and West Antarctica focuses extensional deformation (whether induced by the arrival of a mantle plume head or by plate boundary forces) at the tectonic boundary. Rift-related deformation then includes uplift of the Transantarctic Mountains and formation of the flexural Wilkes Basin in the hinterland. Along strike, both the rifted system and TAM show large variations in the amount of extension and orientation of extensional structures, and uplift rate and timing. The different setup conditions that were tested in the modeling indicate that if such variations exist along strike of the tectonic boundary, they could account for the variation in character and depression/uplift of the WAR and TAM.

7. Conclusions

Numerical models of lithosphere deformation show that formation of the Wilkes depression, uplift of the Transantarctic Mountains, and rifting in the West Antarctic Rift zone can be related to extension of the region along the craton–non-craton boundary that separates East and West Antarctica. The asymmetric pre-rift lithosphere structure of the tectonic boundary focuses extensional deformation in the weakest zone along its side. An asymmetric rift structure is formed, terminating against the craton that remains unaffected by the rifting event. In our models, we find a new mechanism that aids in the formation of the Transantarctic Mountains and local crustal thickening. Accumulation of crustal material at the craton edge results in thickening of the crust, formation of a small crustal root, and uplift of the surface. The models further show that the Wilkes Basin in the hinterland is a flexural depression. Rift-related crustal thinning is not predicted nor expected in this strong hinterland lithosphere.

The models suggest that the inherited lithosphere structure causes extensional deformation to localize at the edge of the East Antarctic craton. The crust of West Antarctica is broadly thinned by the extensional event while maximum thinning occurs near the craton boundary. This results in the abrupt difference in lithospheric strength between East and West Antarctica. Because the craton is basically undeformable, crustal material converges between the rift zone and the craton. Inherited lateral variations in lithospheric structure and strength thus appear to have played a controlling role in the tectonic evolution of the tectonic boundary between West and East Antarctica.

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