Plateau collapse model for the Transantarctic Mountains–West Antarctic Rift System: Insights from numerical experiments

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ABSTRACT

The high elevation and considerable length of the Transantarctic Mountains have led to speculation about their origin. To date, no model has been able to adequately reconcile the juxtaposition of the high, curvilinear Transantarctic Mountains with the adjacent West Antarctic Rift System, a broad region of thin extended continental crust exhibiting wide rift characteristics. We present a first-order investigation into the idea that the West Antarctic Rift System–Transantarctic Mountains region was a high-elevation plateau with thicker than normal crust before the onset of continental extension. With major Cretaceous extension, the rift underwent a topographic reversal, and a plateau edge with thickened crust, representing the ancestral Transantarctic Mountains, remained. In the Cenozoic, minor extension and major denudation reduce the crustal root while simultaneously uplifting peak heights in the mountains. The Cretaceous stage of this concept is investigated using two-dimensional numerical models to determine under what conditions plateau collapse is plausible. Model results indicate that elevation of a remnant plateau edge decreases with increasing initial Moho temperature. Very cold initial Moho temperatures, <675 °C, under the plateau leave a thick plateau edge but do not exhibit wide rifting. A cold to moderate initial thermal structure, Moho temperatures of 675–850 °C, is needed to retain the plateau edge and still exhibit wide rifting in the middle of the plateau. We conclude that this plateau collapse concept is possible using these numerical experiments, and that application of this idea to the West Antarctic Rift System–Transantarctic Mountains system is also supported by geological and geophysical evidence.

Keywords: Transantarctic Mountains, West Antarctic Rift System, rifting, numerical modeling, plateau collapse, Antarctica.

INTRODUCTION

The Transantarctic Mountains are one of the dominant features of the Antarctic continent and define the boundary between cratonic East Antarctica and the West Antarctic Rift System (Fig. 1). The Transantarctic Mountains are more than 3000 km long and have local peak elevations in excess of 4 km, making them the largest and longest rift-related mountain belt in the world. The West Antarctic Rift System is adjacent to the mountains, and is a broad region of thin, extended continental crust 750–1000 km wide, comparable to the Basin and Range Province of North America (Fitzgerald et al., 1986).

Previous quantitative models for the Transantarctic Mountains have mostly been kinematic and include mainly thermal uplift of the mountains (e.g., Smith and Drewry, 1984), a combination of thermal and mechanical uplift related to rifting (e.g., Fitzgerald et al., 1986; ten Brink et al., 1997), and mechanical effects of rifting thick lithosphere (van der Beek et al., 1994; Busetti et al., 1999). Stern et al. (2005) attributed as much as 50% of peak height to the effects of glacial erosion.

While the spatial relationship between the mountains and rift system is obvious, many aspects of their formation remain unclear. Large-magnitude extension in the West Antarctic Rift System during the Cretaceous accompanied relatively little denudation in the Transantarctic Mountains at that time. Conversely, minor Cenozoic extension adjacent to the mountains accompanied or is younger than most of the denudation (Fitzgerald, 2002; Karner et al., 2005; Huerta and Harry, 2007). Geological and geophysical evidence arguing against the often-cited broken plate model for Transantarctic Mountains uplift (ten Brink et al., 1997) is mounting (e.g., Studinger et al., 2004, 2006; Lawrence et al., 2006), most notably, the absence of a laterally continuous crustally penetrating fault at the mountain–rift system interface required by such a model.

Distributed extension in the West Antarctic Rift System may require hot, thin lithosphere (e.g., Buck, 1991), while rift shoulder uplift of the Transantarctic Mountains may require a region of uniformly cold, thick lithosphere (Fig. 2). Distributed wide rifting in the Cretaceous would leave a broad area of hot, thinned lithosphere, similar to

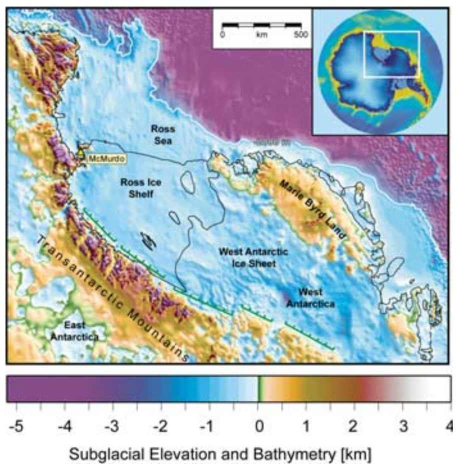

Figure 1. Topographic and bathymetry map of Transantarctic Mountains and Ross Embayment; reference location map is in upper right. Ross Sea is characterized by a series of basins and basement highs and elevated heat flow. Ross Embayment—Ross Sea plus Ross Ice Shelf. WARS—Ross Embayment plus extended parts of West Antarctica.

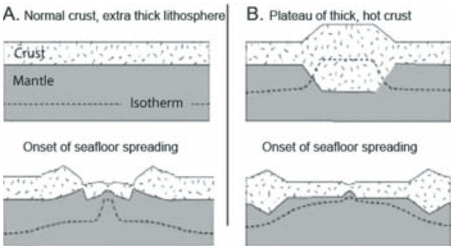

Figure 2. Two classes of models of Transantarctic Mountain uplift and Ross Embayment subsidence. Dashed lines approximate brittle-ductile transition. Top panel shows pre-rift configuration; bottom panels show post-rift configuration. Class A demonstrates extension of cold, thick lithosphere. Rift flank uplift raises the edges of the rift due to thermal effects. Extension is characterized by narrow rifting, and isotherms are shallowed locally. Class B, proposed in this paper, demonstrates extension of a high plateau underlain by hot thick crust. Extension occurs over a broad region in the former plateau, and rift flanks retain much of their elevation through retention of thick crustal roots. Isotherms are shallowed over a wide region.

the Basin and Range. The time between the end of Cretaceous and beginning of Cenozoic rifting is insufficient to cool and thicken the lithosphere under the mountain–rift system interface to the level required for rift shoulder uplift.

As a way to reconcile the structures, distribution of rifting, and the great elevation of the mountains, we explore the possibility that the West Antarctic Rift System was a high-elevation plateau with thicker than normal crust before extension related to plateau collapse. In traditional rift shoulder uplift (Fig. 2A), narrow rifting occurs in cold, thick lithosphere. The rift flanks are elevated due to thermal and mechanical effects, and there is no crustal thickening below the rift flank. In the plateau scenario (Fig. 2B), extension of hot, thickened crust initiates rifting and subsidence in the middle of the plateau. The cooler edge undergoes minimal stretching and is tectonically and erosionally denuded as a result of lateral variations in extension.

We propose a plateau collapse scenario, with the West Antarctic Rift System as the main body of the plateau and Transantarctic Mountains as the plateau edge, to explain the denudational and extensional history of the mountains–rift system. Wide rifting in the Cretaceous extends the rift system, but leaves the Transantarctic plateau edge relatively intact. During the Cenozoic, denudation reduces the crustal root under the mountains while simultaneously enhancing large peak uplift, as described in Stern et al. (2005). Numerical models of the plateau collapse scenario have been previously applied to Basin and Range extension (e.g., Harry et al., 1993). We use numerical models to test the Cretaceous stage, the major phase of extension, of this new hypothesis for the formation of the Transtantarctic Mountains and West Antarctic Rift System and to determine the conditions for which retention of a plateau edge with thick crust is plausible.

TECTONIC SETTING

The Transantarctic Mountains and West Antarctica have been the site of repeated orogenies, including Rodinian assembly, Rodinian breakup, and a transition from a passive to an active margin during Gondwana assembly (e.g., Goodge, 2002). During the Cambrian–Ordovician Ross orogeny the region was part of the active margin of Gondwana (Goodge, 2002, and references therein). Subduction moved outboard to the Pacific margin of Gondwana from at least 320 Ma to ca. 110 Ma (Mukasa and Dalziel, 2000). Reconstructions (e.g., Foster and Gray, 2000) indicate that the along-strike equivalent to the region between Marie Byrd Land and the Transantarctic Mountains, what would have been the majority of the unextended Ross Embayment, from the Cambrian to the Middle Devonian was the Lachlan Fold Belt (orogen) of southeastern Australia, then adjacent to Antarctica. This correlation implies high paleotopography and thickened crust in the West Antarctic Rift System into the Devonian.

Paleocurrent orientations, lithofacies, and stratigraphic relationships of the Devonian–Triassic Beacon Supergroup in the central Transantarctic Mountains suggest that deposition began within two intermontane or successor basins after postorogenic uplift of the Ross terrain (e.g., Isbell, 1999). These data and the presence of sub-Beacon relief in the central Transantarctic Mountains suggest high paleotopography around and between these basins. In southern Victoria Land, the Beacon shows a progression from marine sediment deposition, to coal measures, to alluvial plain sediments (Barrett, 1980, 1991). This history implies the presence of high topography both outboard (Ross Plateau) and inboard (East Antarctica hinterland) of the present-day Transantarctic Mountains and rising paleotopography within the basin, suggesting crustal thickening under the Transantarctic basin during this period.

Voluminous Jurassic tholeiitic magmatism along the Transantarctic Mountains marked the onset of Gondwana breakup (Elliot, 1992). While Jurassic rifting is believed to have occurred within the present-day Transantarctic Mountains–West Antarctic Rift System, no major rift-bounding faults in the mountains or the rift system have been located (e.g., Elliot and Fleming, 2004). The presence of voluminous Jurassic magmatic products along the mountain front in the form of sills and lava flows suggests that this event is the most likely candidate for further thickening of the Transantarctic crust.

Subduction ca. 105 Ma off the Marie Byrd Land active margin ceased (e.g., Weaver et al., 1994, and references therein), transferring New Zealand to the Pacific plate, and marking the onset of major extension in the West Antarctic Rift System (Lawver and Gahagan, 1995). Wide rifting (~400 km) ended with the 84–79 Ma separation of New Zealand and the Campbell Plateau from Antarctica (Stock and Cande, 2002). Denudation (1–2 km magnitude) during the Cretaceous synchronous with this major extension is recorded along the Transantarctic Mountains (e.g., Fitzgerald, 2002, and references therein).

The major phase of denudation, 4–9 km along the Transantarctic front, began in the Eocene, spatially correlative with but temporally preceding minor extension localized in the Ross Sea adjacent to the mountains (e.g., Fitzgerald, 2002, and references therein). The nature and extent of Cenozoic rifting as well as its relationship to the thermochronology data are still uncertain. Narrow rifting in the Terror Rift, transtension in Victoria Land (Wilson, 1995; Rossetti et al. 2006), Adare Trough spreading projected into the Ross Sea (Davey et al., 2006, and references therein), and minimal extension with mostly climate-controlled denudation (Karner et al., 2005) have all been proposed as key mechanisms active in the Cenozoic. Cenozoic alkaline magmatism associated with Cenozoic rifting (e.g., LeMasurier and Thomson, 1990) has a much smaller volume and limited extent compared to Jurassic magmatism, and is not a viable mechanism for thickening the Transantarctic crust.

Studinger et al., (2004, 2006) demonstrated that gravity anomalies across southern Victoria Land and the Scott and Reedy Glacier region are consistent with crustal thickening under the Transantarctic Mountains relative to East Antarctica. Seismic data on the thickness of the Transantarctic crust indicate 40 ± 2 km crust under the mountains and 35 ± 2 km crust under cratonic East Antarctica (Lawrence et al., 2006).

NUMERICAL MODEL

We conducted regional scale two-dimensional numerical models to investigate the conditions conducive for retention of a plateau edge with thickened continental crust after extension. In the models, a viscoelastic-plastic, non-Newtonian layer of thickened continental crust and adjacent crust of normal thickness underlain by a mantle layer is extended by pulling at the edges with a velocity of 1 cm yr^{-1}. Deformation is tracked using an explicit finite-element method similar to the FLAC (fast Lagrangian analysis of continua) technique (Lavier et al., 2000, and references therein). (A complete description of the numerical model is in the GSA Data Repository.[1])

In all models, the initial plateau is 296 km wide, and total model width is 800 km at

[1]GSA Data Repository item 2007177, description of the numerical model, is available online at www.geosociety.org/pubs/ft2007.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

time 0 m.y. Initial plateau crustal thickness is 55 km, and adjacent crust is 32 km thick. Mantle material underlies the crust to 80 km. Initial plateau topography is a rectangle elevated 3 km above the background and is isostatically compensated at depth.

MODEL RESULTS

A variety of extension styles was observed over a suite of 37 numerical experiments, including core complex formation, wide rifting, and narrow rifting. Here we present one example in which the edge of the plateau retains significant crustal thickness to remain an elevated highland after extension. Initial Moho temperature under the plateau, $T_{m,}$ is 680 °C (Fig. 3A).

After 5 m.y. and 100 km of extension (Fig. 3B), the plateau exhibits characteristics of a wide rift. Extension is accommodated by crustal thinning over a wide area in the lower crust that exhibits the highest strain rates. In the upper crust, three basins accommodate most of the extension. Peak elevations have been reduced several hundred meters, and little extension occurs outside of the plateau area.

At 10 m.y. and 200 km of extension (Fig. 3C), several small basins and ridges have developed, but the plateau edges remain several hundred meters above these. Crust has thinned the most under the middle of the plateau, and the edges have undergone the least thinning.

At 20 m.y. and 400 km of extension (Fig. 3D), the plateau edges have retained elevations of ~1.5 km. The crust has been thinned evenly across a broad area formerly occupied by the plateau, and much of the extended region is below sea level. The lower crust has been replaced by strong upper mantle in highly thinned areas. Our model space ends at 20 m.y., analogous to the end of the major phase of extension and minor denudation during the Cretaceous.

A summary of the model results (Fig. 4) demonstrates the thermal conditions required to leave a plateau edge with significant topography and crustal thickness in comparison to the surrounding area. The maximum elevation of the plateau edge is plotted against T_m at 10 m.y. and 20 m.y. There is a roughly linear decrease of plateau elevation with increased T_m, related to the inverse relationship of crustal viscosity and strength with temperature. Nonlinear variation in extensional style due to variations in model thermal structure may be responsible for the scatter of elevations with T_m. A very cold initial T_m, below ~675 °C, allows for a thickened remnant edge, but the extension is characteristic of a narrow rift. An intermediate temperature profile, with T_m ~675–850 °C, exhibits wide rifting or core complex–style rifting and retains a thick edge as a highland, with edge thickness decreasing with increased temperature. Above ~850 °C, no thickened root is retained.

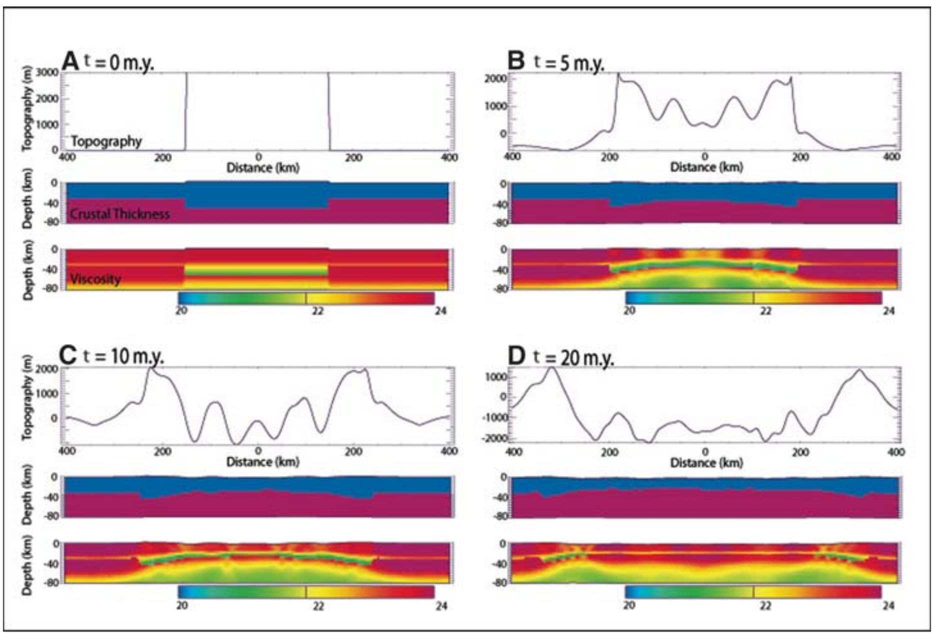

Figure 3. Topography, crustal thickness, and viscosity time slices for 0, 5, 10, and 20 m.y. At 0 m.y. the plateau is represented by a simple box of thickened crust in blue, underlain by olivine mantle in purple. At 5 m.y., basins and ridges have developed in the topography and wide rifting is observed. Lower viscosity regions in upper crust represent shear zones. Lower mantle material has been added to accommodate the space made. At 10 m.y., both lower and upper crust have thinned over a wide region, but plateau edges remain high. At 20 m.y., the region of weak lower crust has thinned significantly, and extension is focused at the rift edge, shown here in the viscosity profile as the lower viscosity upper crust in yellow. High flank edges are underlain by crustal roots.

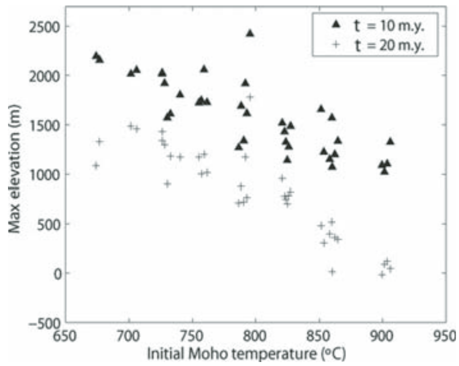

Figure 4. Maximum topography at plateau edge vs. initial Moho temperature under plateau for 10 m.y. and 20 m.y. Both plots show a rough linear trend of decreasing topography with increased Moho temperature, implying that a cold to moderate initial Moho is needed to retain a high plateau edge.

DISCUSSION

The necessary thermal conditions for a plateau collapse scenario in the Transantarctic Mountains–West Antarctic Rift System could be achieved by low to average concentrations of radiogenic elements, crustal rocks with high thermal conductivities in the plateau, or some combination thereof.

These models are not designed to precisely emulate the lithospheric architecture across the mountains–rift system, but simply test a concept. All models shown (Fig. 3) are symmetrical; i.e., there is a plateau remnant on either side. While the Transantarctic Mountains represent one edge of a plateau collapse, the opposite edge of the plateau was represented by the active margin of Gondwana (New Zealand–Marie Byrd Land) and was rifted away with plate reorganization following the breakup of Gondwana.

At the end of the Cretaceous, the major extensional phase, our conceptual model leaves an extended West Antarctic Rift System and relatively unextended Transantarctic Mountains plateau edge with a crustal root of ~12–16 km and elevations of 1–2 km. Significant denudation, ~4–9 km, along the mountain front and decreasing inland, has occurred since the start of the Eocene. This denudation would reduce the crustal thickness of the mountains, yet gravity and seismic studies indicate the presence of an ~5 km root under the Transantarctic Mountains today compared to cratonic East Antarctica. If denudation relates directly to erosion, a crustal root ~9–14 km thick would be required at the end of Cretaceous extension, as is present in our model. All mechanisms of Cenozoic extension, whether Adare Trough spreading, transtention, narrow rifting in Terror Rift, or climate-controlled denudation without major extension, are feasible within the context of our conceptual model. Synchronous with Cenozoic denudation reducing the crustal root of the Transantarctic

Mountains, the peak elevations are increased as much as 50% by glacial incision (Stern et al., 2005), leading to the peak heights of as much as 4.5 km observed today.

CONCLUSIONS

Our numerical experiments demonstrate that rifting a plateau can leave a remnant edge at its flank that retains greater crustal thickness. Coupled with the Cenozoic denudation history of the Transantarctic Mountains, this root would be comparable to that seen under the mountains today. Initially very cold conditions, initial Moho temperatures T_m < 675 °C, do not produce distributed extension and crustal thinning as observed in the Ross Embayment. A moderate temperature profile, T_m ~675–850 °C, is needed to retain crustal thickness and elevation at plateau flanks and exhibit wide rift characteristics in the extended plateau. Topography and crustal thickness are not retained at plateau flanks under initially hot, T_m > 850 °C, conditions. A plateau collapse scenario agrees with the geological history of the region and erosional studies attributing significant peak height increase to the effects of glacial incision.

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