Isostatic rebound due to glacial erosion within the Transantarctic Mountains

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ABSTRACT

In temperate climates, ~25% of peak elevations in mountain ranges can be created by isostatic rebound as a response to erosional incision. Significantly more relief generation and peak uplift are, however, possible for glacial erosion in a polar climate. We incorporate regional isostasy using flexure of an elastic plate to show that isostatic rebound as a response to glacial incision can account for as much as 2000 m or 50% of peak elevation in the central Transantarctic Mountains. Differences in relief of at least 5500 m over lateral distances of just 40 km are evident within the central part of the 3000-km-long mountain range. Such strong relief is possible because a polar climate since the middle Miocene has resulted in freezing conditions at high elevations, which acted to preserve the peaks, whereas wet-based glaciers at low elevations have produced optimal conditions for enhanced glacial incision. Because isostatic rebound results in permanent peak uplift, this mechanism provides an explanation of why the Transantarctic Mountains are one of the higher and more long-lived continental rift margins on Earth.

Keywords: isostasy, flexure, Transantarctic Mountains, glacial erosion, isostatic rebound.

INTRODUCTION

The process of isostatic uplift in response to the incision of valleys is at the center of the debate between the roles of climate and tectonics in generating relief (Molnar and England, 1990; Watts et al., 2000). Although modeling studies point to topographic relief being created by glacial erosion under some conditions (Tomkin and Braun, 2002), an empirical test of the phenomenon in mid-latitude glacial conditions shows limited relief production (Small and Anderson, 1998). In this study we quantify glacial incision of the central Transantarctic Mountains (Fig. 1) and show that as much as 2000 m of peak uplift can be attributed to the isostatic response to incision by the large outlet glaciers that drain the East Antarctic Ice Sheet. This uplift is the largest rebound attributed to incision, and we therefore ask the following: What percent of relief and peak elevation does this rebound represent? What makes the Antarctic conditions favorable for the production of relief? Can the unusual length, longevity, and height of the Transantarctic Mountains (Stern and ten Brink, 1989) be attributed to isostatic rebound as a response to glacial incision?

BACKGROUND

At 3000 km long and with peaks reaching 4500 m high, the Transantarctic Mountains are the highest continental rift-flank uplift on Earth (Stern and ten Brink, 1989; Sugden and Denton, 2004). They formed at the boundary between East and West Antarctica (Fig. 1), where there are abrupt changes in crustal thickness, upper-mantle S-wave speeds, and surface elevation (Bannister et al., 2003; ten Brink et al., 1993). However, the Transantarctic Mountains differ from most mountain ranges in two principal ways: they formed as an uplifted rift-flank in an extensional environment (Fitzgerald, 1992), rather than by compression, and glacial erosion has deeply incised the mountains to depths of as much as 1500 m below sea level (Drewry, 1983; Rose, 1982).

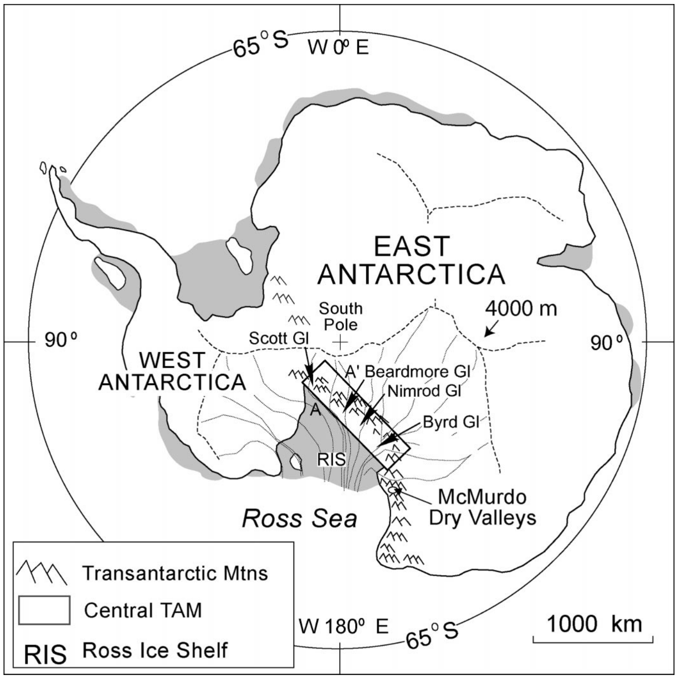

Figure 1. Map of Antarctica showing Transantarctic Mountains (TAM) at boundary between East and West Antarctica. Area of study in central Transantarctic Mountains and main outlet glaciers (Gl) of Beardmore, Nimrod, Scott, and Byrd are shown. Shaded areas are permanent ice shelves. A-A′ shows position of topographic profile shown in Figure 2.

Exhumation in the Transantarctic Mountains (Fig. 2) initially began in the Cretaceous, but most exhumation, on the basis of fission-track data, took place after 55 Ma (Fitzgerald, 1992). Timing of the incision is unknown, although there is evidence from offshore drill cores that at

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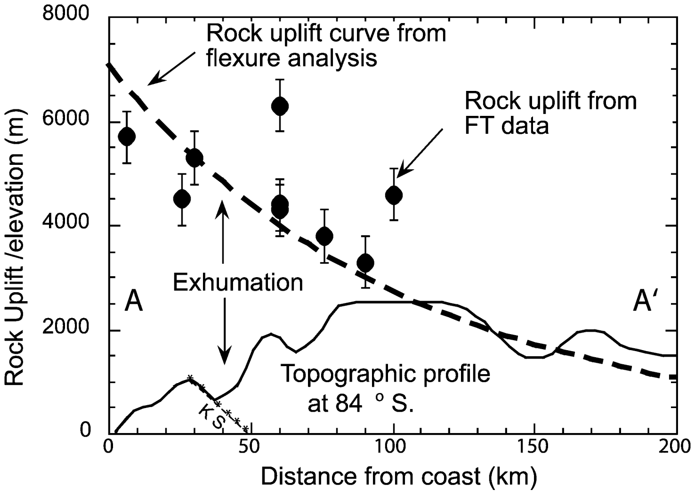

Figure 2. Plot of rock uplift based on interpretation of fission-track (FT) data from Scott and Beardmore Glaciers, and McMurdo Dry Valleys (from Fitzgerald, 1992, 1994; Fitzgerald and Stump, 1997). Also shown is surface topography at 84°S (A-A′) in Figures 1 and 3A. Dashed line is flexural profile based on gravity data, subglacial topography, dip of Kukri erosion surface (KS), and adopted variation in effective elastic thickness (T_e) of 5–85 km (ten Brink et al., 1997).

least 2 km of incision had occurred by early Oligocene time (Smellie, 2001).

OBSERVED SUBGLACIAL TOPOGRAPHY

Contour maps of subglacial topography (Drewry, 1983) for a 1200 × 300 km section of the Transantarctic Mountains from 80°S to 85°S form the basis for the analysis presented here (Fig. 1 and Appendix DR1[1]). Deep glacial troughs cut through the Transantarctic Mountain front (Fig. 3A), and these troughs contain glaciers as wide as 35 km that continue back into the mountains for distances of 100–200 km. The 5 km digitization (Appendix DR1; see footnote 1) of the contour data smoothes the extreme elevation differences represented on the digital elevation model (DEM). However, the original data show that the maximum relief is between the deep glacial troughs of the Beardmore and Nimrod Glaciers (−1250 m) and the peak of Mount Kirkpatrick (4528 m). This difference of ~5800 m occurs over a lateral distance as small as 40 km. Incision within the Transantarctic Mountains was initially fluvial (Sugden and Denton, 2004), but judging from valley form and given the depth of downcutting below sea level, the significant mass removal was most likely by glacial processes (Hicock et al., 2003). Similar relief (~5500 m) has been created by the gorges that cut through the Himalayas (Montgomery, 1994).

ISOSTATIC RESPONSE TO EROSION

Previous analyses (Kerr and Gilchrist, 1996; Stern and ten Brink, 1989) of the isostatic response to erosion in the Transantarctic Mountains were based on the total rock removed as estimated from fission-track data; this study is concerned with only the response due to incision. We estimate the isostatic response to incision by fitting a smooth summit-accordance surface to the dominant peaks within the study area (Fig. 3B). Subtraction of observed subglacial topography from this surface yields a map of the mass that is missing because of incision (Fig. 3C). Incision estimated this way is the minimum that has occurred since rock uplift began, because rock has also been eroded from the mountain peaks (Fitzgerald, 1992) (Fig. 1), thus lowering the

[1]GSA Data Repository item 2005038, Appendix DR1, additional information on methods and interpretation, is available online at www.geosociety.org/pubs/ft2005.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

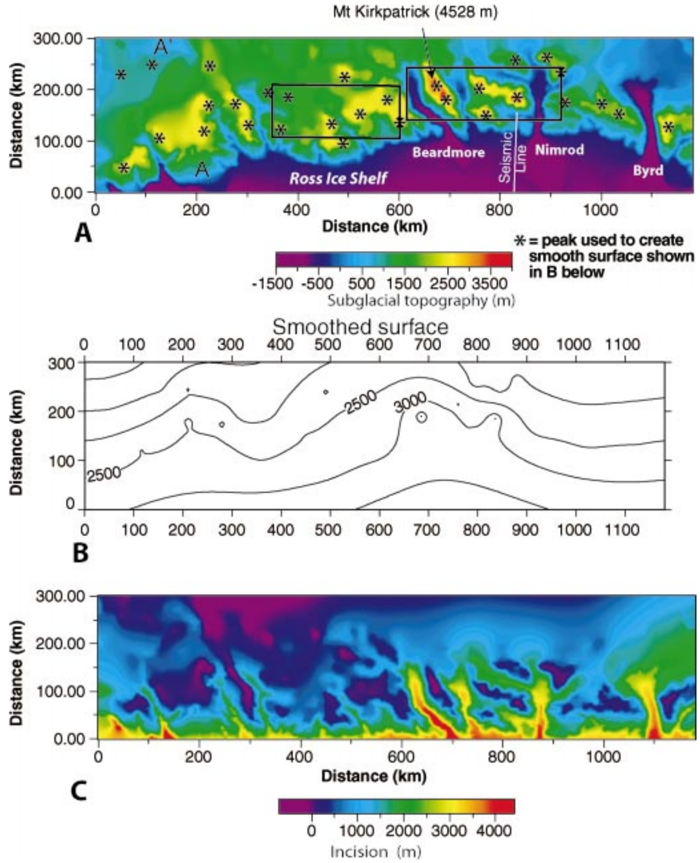

Figure 3. A: Digital elevation model of central Transantarctic Mountains and subglacial topography based on original radio echo-sounding (Drewry, 1983) data that were contoured and digitized into 5 × 5 km grid. Area of study shown in Figure 1. Transantarctic Mountain front occurs at abrupt juxtaposition of Transantarctic Mountains with submarine topography of Ross Ice Shelf. Seismic line is that presented by ten Brink et al. (1993). Star symbols represent peaks selected automatically on basis of slope and elevation criteria, which are used to construct peak-accordance surface shown in B. Rectangular areas represent areas for which mean elevations are calculated (plotted in Fig. 5B). B: Peak-accordance surface calculated from peaks marked in A by fitting smooth surface (Appendix DR1; see footnote 1) to elevations of peaks shown in A. C: Glacial incision calculated from difference between B and A. Coastline has been straightened as discussed in text.

datum for our summit-accordance surface. We remove the curvature from the mountain front (Fig. 3C) by using a simple digital transformation that moves each y-coordinate of the DEM to a common value at the Transantarctic Mountains front. This procedure is necessary to test the free-edge end-member model that requires, from a numerical standpoint, that it be a straight edge (Stern et al., 1992). This transformation does not affect the predicted rebound, as the deviations from a straight boundary (~100 km) are less than the typical flexural wavelength of rebound (Fig. 4).

A three-dimensional loading and flexure code (Stern et al., 1992) calculates the isostatic response to the three-dimensional incision (Appendix DR1; see footnote 1). Mass removed by erosion is considered as a negative load for which Earth compensates by rebounding, the pattern of rebound being controlled by both the wavelength of erosion and the flexural rigidity for the lithosphere (Walcott, 1970; Watts, 2001). Flexural rigidity (D) is related to the effective elastic thickness (T_e) by $D = ET_e^3/12\ (1 - \nu^2)$, where E is Young's modulus and ν is Poisson's ratio. Isostatic rebound is calculated on a 5 × 5 km grid by using four rheological models that represent a range of proposed tectonic scenarios for the Transantarctic Mountains and adjacent regions

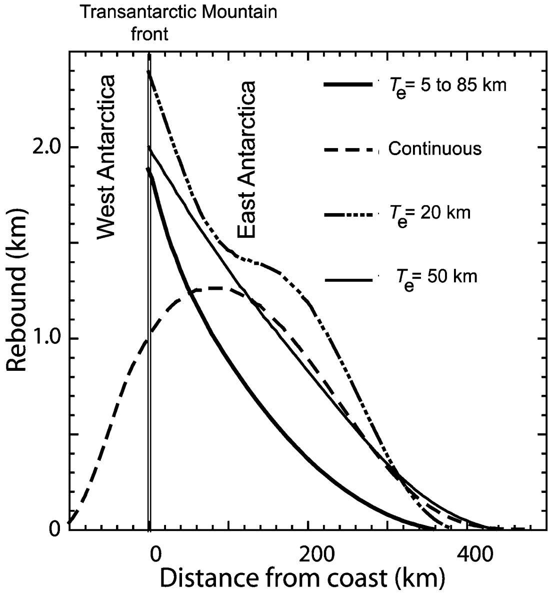

Figure 4. Predicted rebound for profiles through Beardmore-Nimrod area based on three-dimensional flexure modeling (Stern et al., 1992). Three models with free edge and one with continuous plate are shown. For continuous plate model, effective elastic thickness, T_e, is 20 and 40 km in West and East Antarctica, respectively. Preferred model is heavy solid line where T_e = 85 km except within 120 km of front, where it drops to T_e = 5 km. Note that T_e = 20 km model allows short-wavelength variations in incision to be expressed in predicted rebound. For other models, flexural rigidity suppresses these variations.

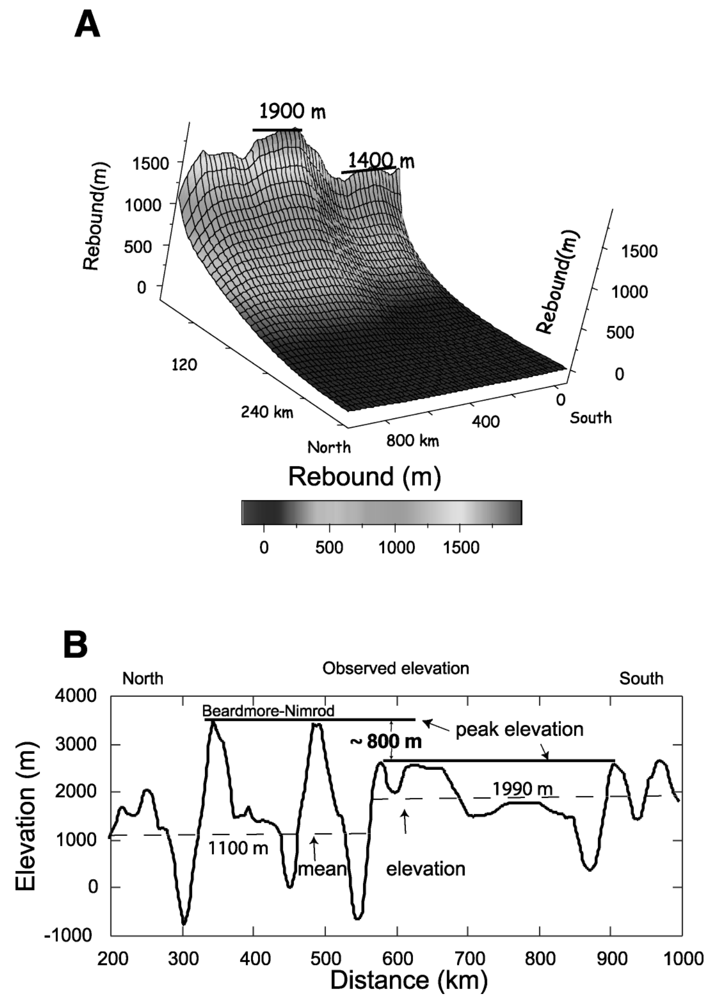

Figure 5. A: Rebound response in three dimensions for effective elastic thickness, T_e = 5 to 85 km model in Figure 4. Note ~500 m difference between Beardmore-Nimrod region and region to south. B: Observed topography along profile ~40 km back from Transantarctic Mountains front, showing ~800 m jump in summit accordance between Beardmore-Nimrod area and area to south. Mean elevation for each region (1100 and 1990 m) is found by averaging topography in rectangular areas shown in Figure 3A. Note drop in mean elevation of mountains for region with highest peaks.

(Busetti et al., 1999; Kerr and Gilchrist, 1996; Stern and ten Brink, 1989; ten Brink et al., 1997). Models with a free edge (Fig. 4) show maximum rebounds at the mountain front of 1900–2400 m, whereas the continuous-plate model (T_e values of 40 and 20 km in East and West Antarctica, respectively) shows a maximum rebound of 1300 ± 100 m.

Our preferred model is the free-edge one in which T_e = 85 km for most of East Antarctica, but 120 km from the mountain front, T_e drops in a linear manner to T_e = 5 km at the front (i.e., the boundary with West Antarctica). This model is preferred because a drop to T_e = 5 km is required to match the 2°–3° dip of the Kukri erosion surface (Fig. 1) 40 km back from the mountain front (Stern and ten Brink, 1989; ten Brink et al., 1997). Furthermore, the variation in T_e of 85–5 km is in keeping with that found by flexural studies for other cratonic margins of Gondwana (Gunnell and Fleitout, 2000; Zuber et al., 1989). Continuous-plate models, however, produce lower dips than are observed for the Kukri erosion surface (Stern and ten Brink, 1989) and require that rebound from incision in East Antarctica will be distributed as a ~1000-m-high, 100-km-wide uplift into West Antarctica (Fig. 4). Evidence for such a distributed rebound is not evident in the sedimentary stratigraphy adjacent to the central Transantarctic Mountains. Here a seismic reflection profile (Fig. 3A) shows a relatively flat lying sedimentary section that is only upturned within 30 km of the mountain front (ten Brink et al., 1993).

REBOUND, PEAK ELEVATIONS, AND RELIEF

Regardless of what flexural rigidity model is adopted, the maximum predicted rebound is in the Beardmore-Nimrod region, where glacial incision is most intense. Predicted rebound is ~500 m higher here than for the region to the south (Fig. 5A), a trend mirrored in the topography of the peak summits (Fig. 5B). What we see here, therefore, is the often predicted (Molnar and England, 1990; Montgomery, 1994), but rarely demonstrated, process of incision producing a reduction in the mean elevation of the mountains, yet an increase in peak elevations. Specifically, the mean elevation in the Beardmore-Nimrod region is only 1100 m, yet the smoothed peak elevation is ~3500 m (Figs. 3A and 5B). In contrast, for the area to the south, mean and peak elevations are 2000 and 2600 m, respectively.

An isostatic rebound for the preferred free-edge model of 1900 ± 100 m is ~50% of peak elevation (taking 4000 m as the mean height of the main peaks), or 36% of maximum relief (5500 m). The continuous-plate model predicts ~1300 ± 100 m of rebound, or 32% of peak elevation and 22% of relief. Thus, between 32% and 50% of peak elevation in the central Transantarctic Mountains is accounted for by rebound as a consequence of glacial incision. This result contrasts with the response to glacial incision in temperate regions where rebound rarely exceeds 25% of the peak elevation (Gilchrist et al., 1994; Montgomery, 1994).

CLIMATE AND RELIEF PRODUCTION

Maximum peak elevation is limited in temperate climates by the interplay between relief-generating and relief-reducing mechanisms (Gilchrist et al., 1994; Whipple et al., 1999). Greater relief in the Transantarctic Mountains is, nevertheless, explicable within the context of Antarctic cooling through Cenozoic time. The largely ice-free conditions of the early Cenozoic ended with the first big ice sheet ca. 34 Ma, and was followed by ice sheets with temperate margins fluctuating on Milankovitch frequencies (Naish et al., 2001). Shortly after 15 Ma a persistent ice sheet developed on East Antarctica (Flower and Kennett, 1994), maintaining a frozen landscape in the high Transantarctic Mountains (Sugden and Denton, 2004). However, deep-sea $\delta^{18}O$ isotope records show 40 k.y. oscillations of $\sim$0.6%, most likely representing significant Antarctic ice-volume fluctuations, into the Pliocene (Barrett, 1996). Such rapid oscillations are the very conditions that could enhance erosion (Zhang et al., 2001), particularly at the margin of the ice sheet where the advance and retreat of the major wet-based (Robinson, 1984) outlet glaciers would impart an effective reaming action to the bedrock.

Optimal conditions for creating maximum relief, termed ''selective linear erosion'' (Sugden and John, 1976), beneath ice sheets occur when the basal melting is concentrated in troughs and adjacent peaks remain preserved in a frozen climate. On the basis of the climate evidence discussed here, these conditions are most likely to have been first achieved in the Transantarctic Mountains in the middle Miocene and to have continued through to the present day. Geochemical evidence in the McMurdo Dry Valleys (Fig. 1) records the antiquity of frozen peaks that weather at $<$0.2 m/m.y. in a hyperarid climate (Summerfield et al., 1999), yet the main outlet glaciers through the Transantarctic Mountains are wet based and erosive (Sugden and Denton, 2004). We therefore propose that the unique combination of glacial erosion in a polar climate and orbitally forced glacial fluctuations explain why the Transantarctic Mountain range is the highest and one of the more long-standing continental rift margins on Earth.

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