Geochemistry Geophysics Geosystems G3
AN ELECTRONIC JOURNAL OF THE EARTH SCIENCES
Published by AGU and the Geochemical Society

Article
Volume 7, Number 12
30 December 2006
Q12011, doi:10.1029/2006GC001294
ISSN: 1525-2027

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Bedrock platforms within the Ross Embayment, West Antarctica: Hypotheses for ice sheet history, wave erosion, Cenozoic extension, and thermal subsidence

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[1] Ice-penetrating radar (mostly airborne) and marine seismic surveys have revealed plateaus and terraces about 100–350 m below sea level beneath parts of the Ross Embayment, including the West Antarctic ice sheet, the Ross Ice Shelf, and the eastern Ross Sea. These surfaces cover many thousands of square kilometers and are separated by bedrock troughs occupied by the West Antarctic ice streams. Prominent plateaus are under Edward VII Peninsula, Siple Dome, and Roosevelt Island. Marine seismic data and gravity data over the buried plateaus support an interpretation that they were caused by erosion into basement. The flat and level nature of the surfaces that were formed by erosion, are near the same depth over large distances, and fringe the buried rugged bedrock topography of Marie Byrd Land supports an interpretation of marine rather than glacial erosion. Marine seismic reflection profiles over one of the plateau remnants show it as an angular unconformity cut into gently dipping sediments of late Oligocene age, draped with thin, flat-lying sediments, implying that the plateau is early Miocene or younger. The younger limit for plateau erosion follows from the interpretation of the mechanism: erosion must predate permanent ice along the coast associated with the growth of the West Antarctic ice sheet, probably prior to 10 Ma and possibly prior to 14 Ma. The plateaus along the Siple Coast, with depths around 350 meters, do not rebound to close to sea level and wave base for models of removing past and present ice load. This discrepancy can be explained if Ross Embayment lithosphere has been cooling and subsiding since significant extension in Eocene to Oligocene time. Our interpretation requires crustal subsidence in the Ross Sea, which in turn implies higher bedrock elevations in the past within this region. Higher-standing topography presented an opportunity for the accumulation of local ice sheets and caused sea level fluctuation prior to late Cenozoic time.

Components: 10,873 words, 13 figures.

Keywords: Antarctic; geomorphology; subsidence.

Index Terms: 9310 Geographic Location: Antarctica (4207); 1824 Hydrology: Geomorphology: general (1625); 8109 Tectonophysics: Continental tectonics: extensional (0905).

Received 27 February 2006; **Revised** 11 July 2006; **Accepted** 13 October 2006; **Published** 30 December 2006.

Wilson, D. S., and B. P. Luyendyk (2006), Bedrock platforms within the Ross Embayment, West Antarctica: Hypotheses for ice sheet history, wave erosion, Cenozoic extension, and thermal subsidence, *Geochem. Geophys. Geosyst.*, *7*, Q12011, doi:10.1029/2006GC001294.

1. Introduction

[2] The Ross Sea and Ross Ice Shelf of Antarctica occupy a region known as the Ross Embayment, a geographic region of sea and ice shelf that separates the subcontinents of East and West Antarctica (Figure 1). The Ross Embayment comprises the geological provinces of the West Antarctic rift and Ross Sea rift [*Behrendt et al.*, 1991a, 1991b; *Tessensohn and Worner*, 1991]. The continental margin of the Ross Sea includes features unique to the continent, including an over-deepened shelf at a depth of around 500 meters, and evidence of multiple advances and retreats of continental glaciations during the Cenozoic [*Anderson*, 1999].

[3] Reconnaissance mapping of the Ross Sea margin has largely been accomplished [*ANTOSTRAT*, 1995] except in the far eastern parts of the sea adjacent to western Marie Byrd Land (wMBL). Onshore beyond the grounding line of the ice sheets, little has been known of the topography of the bedrock beneath the ice until recently. Radar sounding from the ice surface and from aircraft has mapped sub ice topography in selected locations (BEDMAP: http://www.antarctica.ac.uk/bedmap/ [*Lythe et al.*, 2001]). One mapped area of bedrock includes parts surrounding the West Antarctic Ice Streams, fast-moving glaciers (kilometer per year) that drain the parts of the West Antarctic Ice Sheet into the Ross Ice Shelf.

[4] A record of West Antarctic glacial history has been interpreted from study of the glacial marine sediments overlying basement in the Ross Sea [*Anderson*, 1999]. These studies have revealed eight or more stratigraphic and seismic sequences [*ANTOSTRAT*, 1995; *Hinz and Block*, 1983]: the Ross Sea Seismic Sequences (RSS-) numbered from oldest RSS-1 to youngest RSS-8, separated by unconformities (RSU-) numbered from youngest RSU1 to oldest RSU7 [*Luyendyk et al.*, 2001]. Sequences RSS-2 (late Oligocene–early Miocene) and younger are glacial marine or glacial in origin. The unconformities are variously interpreted as due to erosion by grounded ice [*Anderson and Bartek*, 1992] or erosion or nondeposition due to fluctuating sea level [*DeSantis et al.*, 1999].

[5] In this paper we discuss plateaus and terraces found at about 100–350 m below sea level beneath parts of the Ross Embayment including the West Antarctic ice sheet, the Ross Ice Shelf, and the eastern Ross Sea. We define plateaus as nearly flat surfaces separated from surrounding lower elevations by at least one steep side, and terraces as nearly flat surfaces interrupting slopes. We use the term platform to include both plateaus and terraces.

2. Observations

[6] Airborne radar surveys reported by *Luyendyk et al.* [2003] and *Behrendt et al.* [2004] have dramatically increased the data coverage for bed topography along much of the margin of the Ross Embayment (Figures 2 and 3). Two remarkably flat surfaces within this new coverage stand out amid generally rough bed topography: beneath Siple Dome (Ice Stream C-D divide) and beneath the divide between Ice Streams E and F, here termed the Blue Plateau. The Siple Dome bed plateau includes an area about 40 $\times$ 80 km with the bed elevation between 325 and 355 meters below sea level (mbsl). The Blue Plateau is a longer, crescent-shaped feature with length of 200 km, a maximum width of 40 km, and most of the bed surface between 230 and 260 mbsl. In both cases, subtle ridges with dimensions about 10 $\times$ 20 km rise about 20–40 m above the adjacent plateaus. Both plateaus have steep boundaries with the beds of the adjacent ice streams at depths of 500–700 m.

[7] The bed at Roosevelt Island appears to include a plateau of comparable scale, but mapping is much less complete (Figure 4). Observations are limited to ground-based radar in the southeast part of the island [*Conway et al.*, 1999], and seismic soundings from 1961–1963 reported by *Greischar et al.* [1992] that had position and elevation control from transit surveying (C. Bentley, personal communication, 2005). Most of the soundings from the northern half of the island are in the range 185–205 mbsl, with an eastern subset showing deeper, northeast-sloping surface (Figure 4). Below the southern half of the ice island, most seismic and radar bed observations are in the range 135–

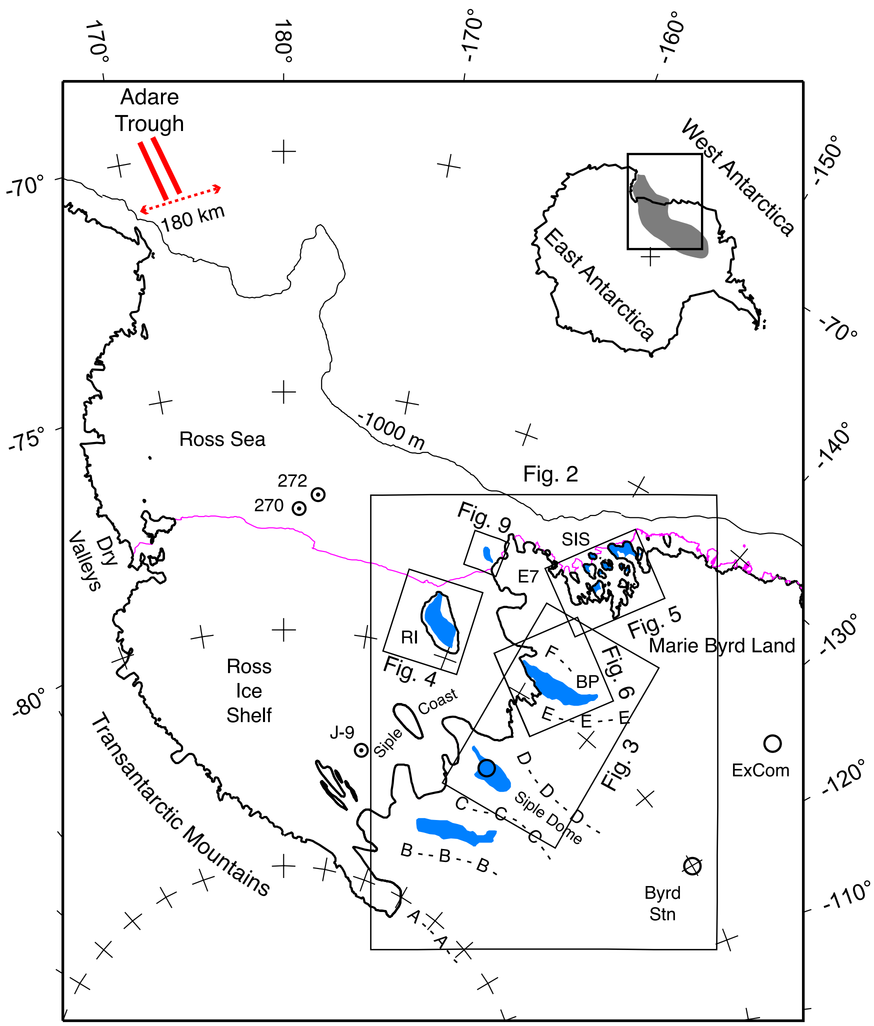

Figure 1. Location map showing positions of plateaus and terraces (blue), ice streams (dashed lines A–F), and locations of subsequent figures. Gray shade in inset shows low-elevation continental crust of the Ross Embayment. Interpretation of 180 km of seafloor spreading on the Adare Trough at 44–27 Ma from *Cande et al.* [2000] is shown by dashed red arrow. Single circles show sites with ice-thickness history discussed in the text at Executive Committee Range, Byrd Station, and Siple Dome. Double circles show sediment core sites. Abbreviations: RI, Roosevelt Island; E7, Edward VII Peninsula; SIS, Sulzberger Ice Shelf; BP, Blue Plateau; ExCom, Executive Committee Range.

155 mbsl. The lack of soundings at depths of 155–185 m suggests two separate, nearly flat surfaces rather than a gentle northward slope. Our interpretation of the bed topography of the island differs substantially from that of the BEDMAP compilation [*Lythe et al.*, 2001]. Lythe et al. generally derived bed topography by subtracting ice thickness measurements from surface elevations derived from an Antarctic-wide digital elevation model (DEM) [*Liu et al.*, 1999]. For the Roosevelt Island area, the DEM does not contain recent data and island elevations appear to be mislocated by about 20 km. Our compilation instead assumes the original elevation measurements to be accurate.

[8] Airborne radar in the vicinity of Sulzberger Ice Shelf [*Luyendyk et al.*, 2003] shows several fairly flat surfaces at depths near 100 m (Figure 5). The surfaces are rougher than the previous examples, but are generally smoothest at depths of 60–120 m. Most of the surfaces have a gentle, down-to-northwest slope, and relations between the surfaces are more consistent with nearly parallel slopes at similar depths rather than discontinuous pieces of a single, planar surface. In several cases, moderately steep topography abruptly rises hundreds of meters above the flatter terraces, generally on the southeast side of the terrace.

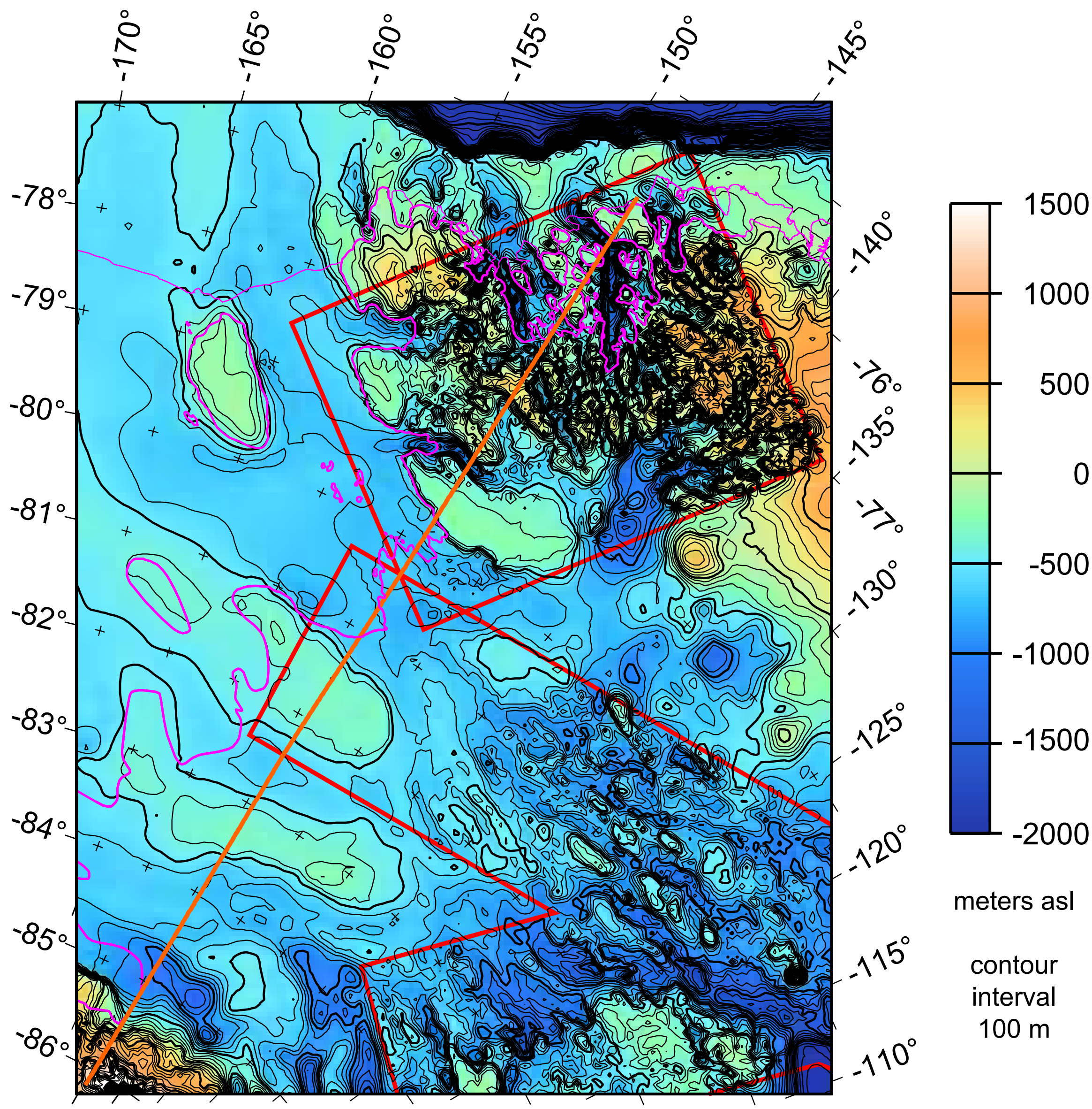

Figure 2. Bed topography and bathymetry in the eastern Ross Embayment, revised from the BEDMAP compilation [*Lythe et al.*, 2001] including data from *Luyendyk et al.* [2003], *Behrendt et al.* [2004] (red boxes), and a new compilation of bathymetry data [*Luyendyk et al.*, 2002, Figure 11]. Location shown in Figure 1. Orange line shows profile for Figure 13.

[9] First-generation airborne radar [*Rose*, 1979] suggests that the some of the bed surface underlying the divide between Ice Streams B and C is at comparable depth to the Siple Dome bed and may be similarly flat. Uncertainties are much larger than for more recent data due to barometric measurement of aircraft elevation and analogue recording technology. A suggested flat and level surface of dimensions about 40 × 100 km is largely at depths of 360–420 m (Figures 1 and 2).

[10] The relation of the bed topography and the gravity data presented by *Luyendyk et al.* [2003] suggests that the Blue Plateau surface has been eroded into both basement and sedimentary deposits (Figure 6). The southern part of the plateau surface between 146°W and 149°W corresponds to a negative gravity anomaly which is well modeled by a sediment-filled trough of at least 1-km thickness. East of 146°W and west of 149°W, the north edge of the gravity low aligns with the south edge

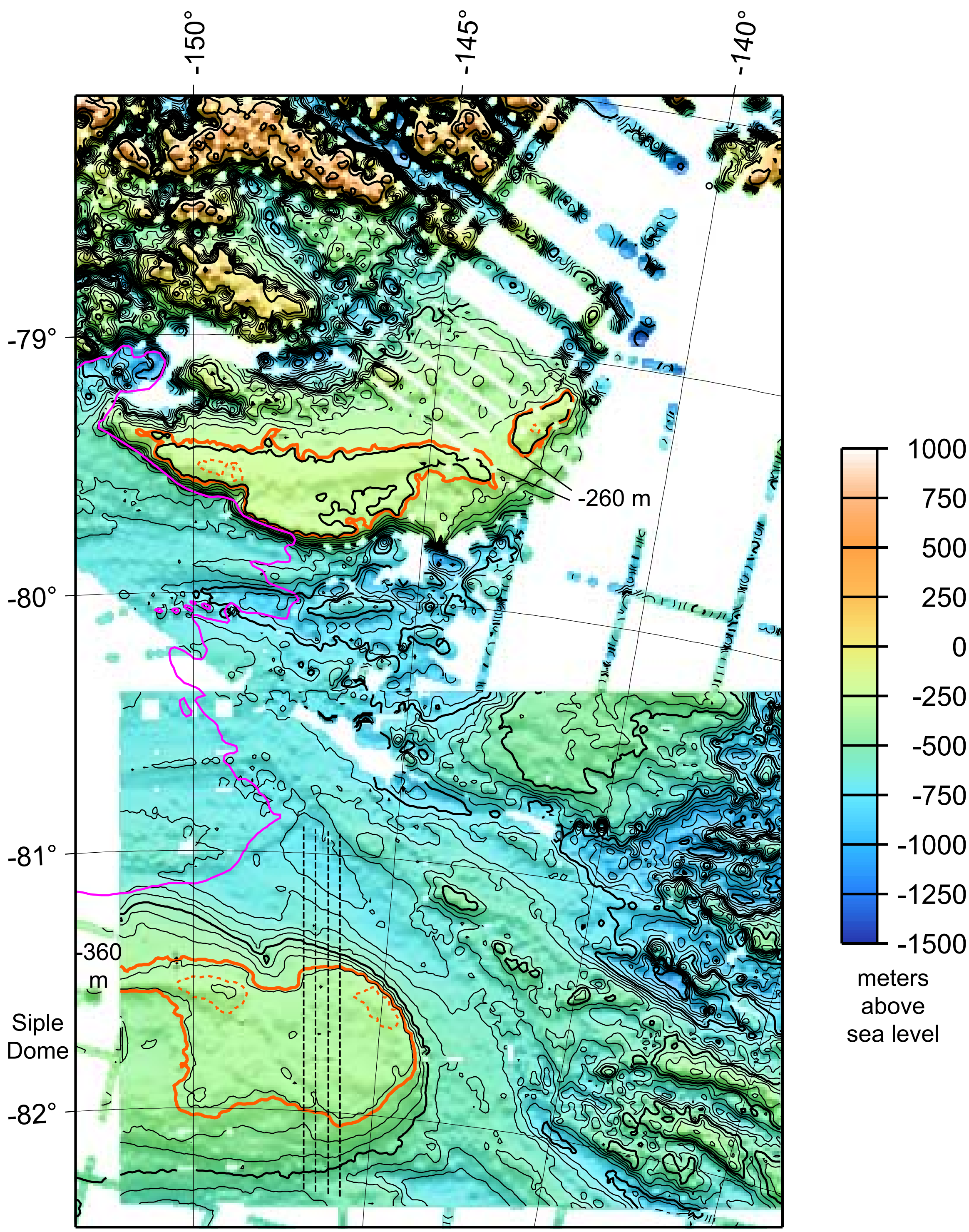

Figure 3. Shaded-relief image of base-of-ice elevation measured by airborne radar in the eastern Ross Embayment. Contour interval is 50 m below sea level, 250 m above sea level. Data sources are *Rose* [1979] (thin tracks), *Luyendyk et al.* [2003], and *Behrendt et al.* [2004]; grounding line (magenta) modified from *Luyendyk et al.* [2003] and *Jacobel et al.* [1994]. Orange contours show lower (solid) and upper (dashed) limits of 40-m elevation range defining the major plateaus. Location shown in Figure 1. Dashed lines show profiles analyzed in Figure 7.

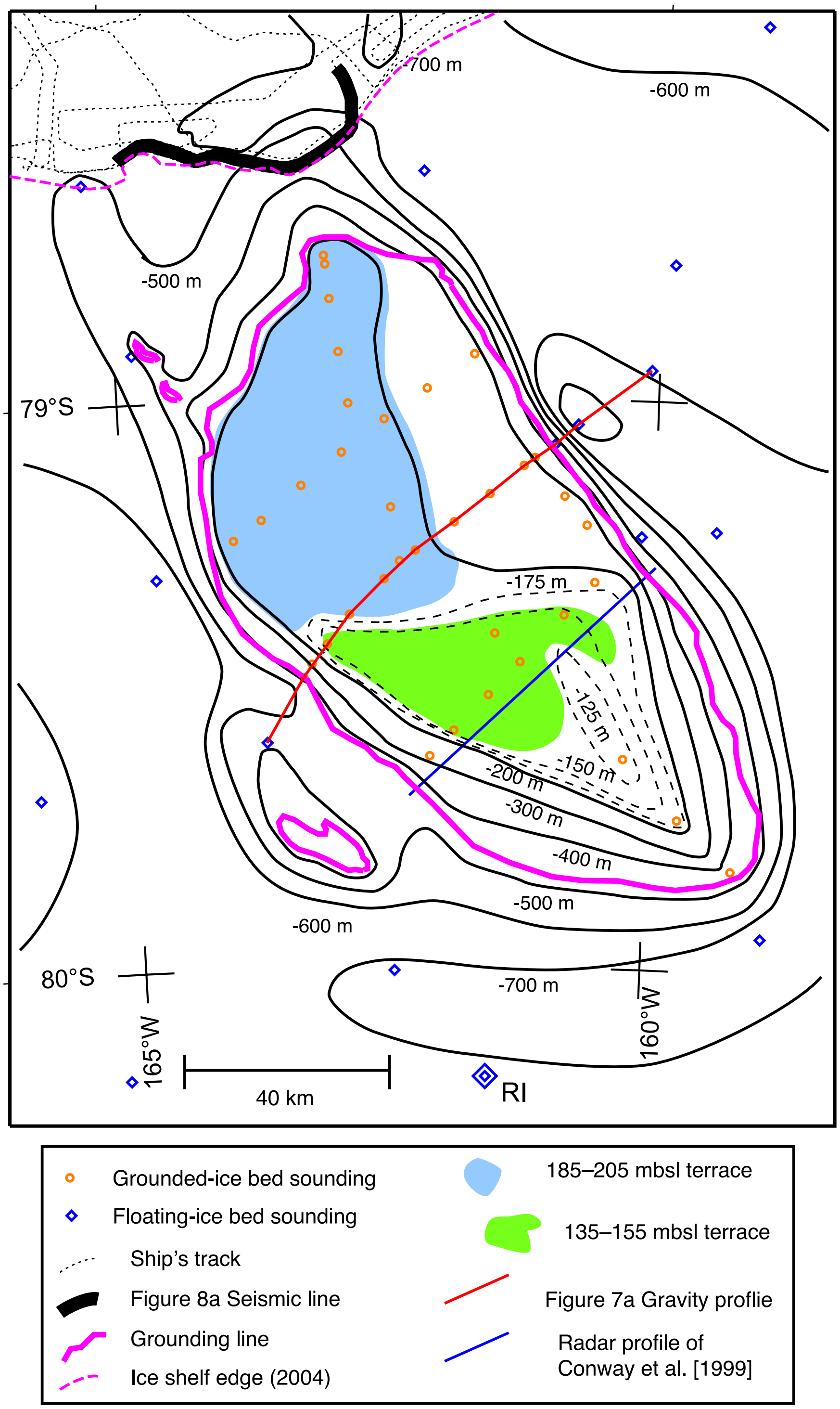

Figure 4. Data locations and hand-contoured bathymetry and bed topography at Roosevelt Island. Soundings are based on seismic data from the compilation of *Greischar et al.* [1992]. Ross Sea bathymetry is from cruises NBP0301 and NBP0306. Grounding line is from inspection of MODIS satellite imagery. RI is the center of a seismic survey showing 700 m of sediment overlying acoustic basement [*Greischar et al.*, 1992]. Location shown in Figure 1.

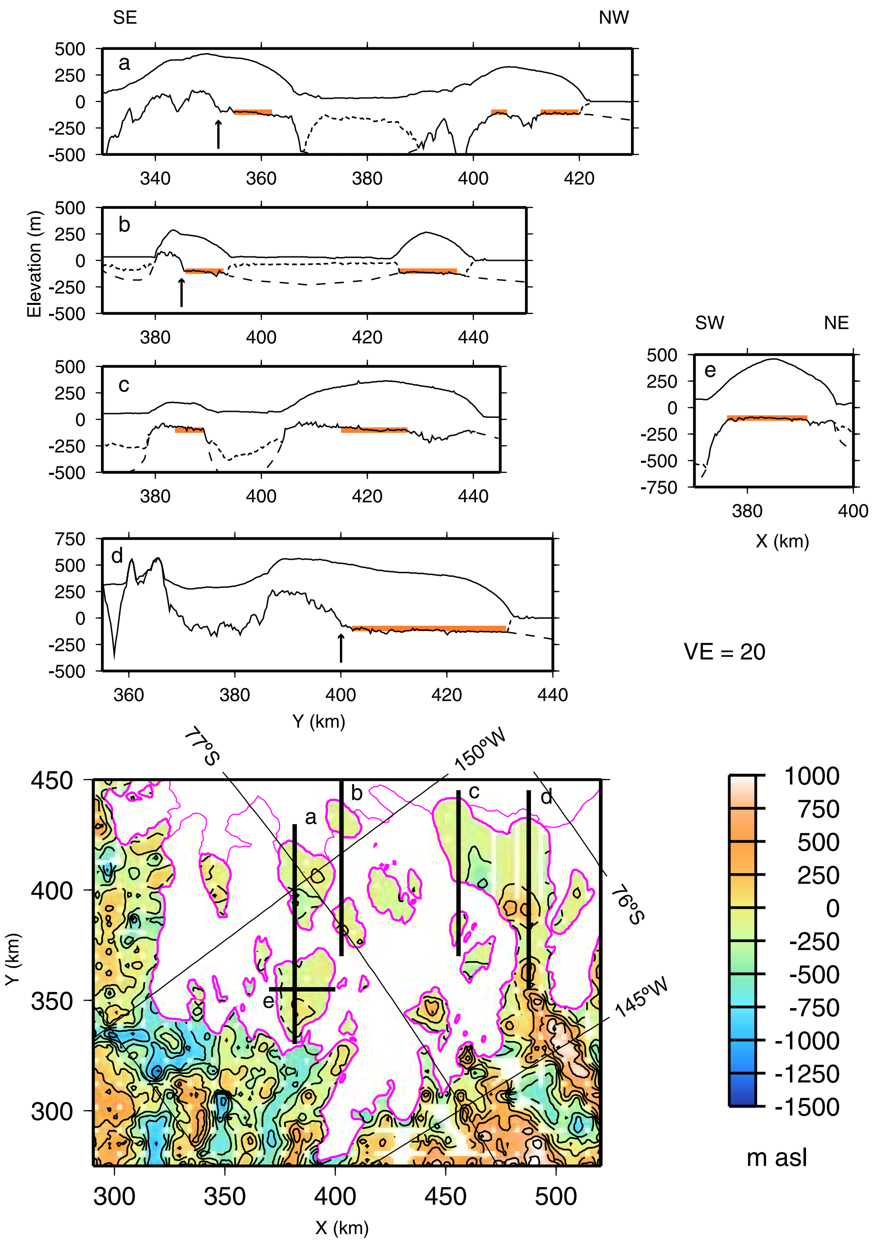

Figure 5. Airborne radar profiles and map showing contoured bed elevation and profile locations within and near the Sulzberger Ice Shelf. Short-dashed lines in profiles show observed base of floating ice; long-dashed lines show approximate depth to bedrock, estimated from gravity anomalies [*Luyendyk et al.*, 2003]. Horizontal orange lines highlight nearly flat surfaces at 60–120 mbsl. Steep slopes at the southeast end of the platforms on profiles a, c, and d are interpreted as relict sea cliffs (arrows). Location shown in Figure 1.

of the plateau, suggesting that less resistive sediments have been removed from these areas by widening of the bordering ice stream trough. Any variation of the plateau surface with interpreted rock type is, at most, subtle, with the basement surface possibly slightly shallower and rougher at the vertical scale of 10–20 m.

[11] Modeling of gravity data over Siple Dome and Roosevelt Island yields similar results. A model for aerogravity data [*Bell et al.*, 1999] over Siple Dome using bed topography from *Behrendt et al.* [2004] shows that the gravity field is best matched by contrasts in the bedrock that are likely sediments within crystalline rock (Figure 7a). It cannot

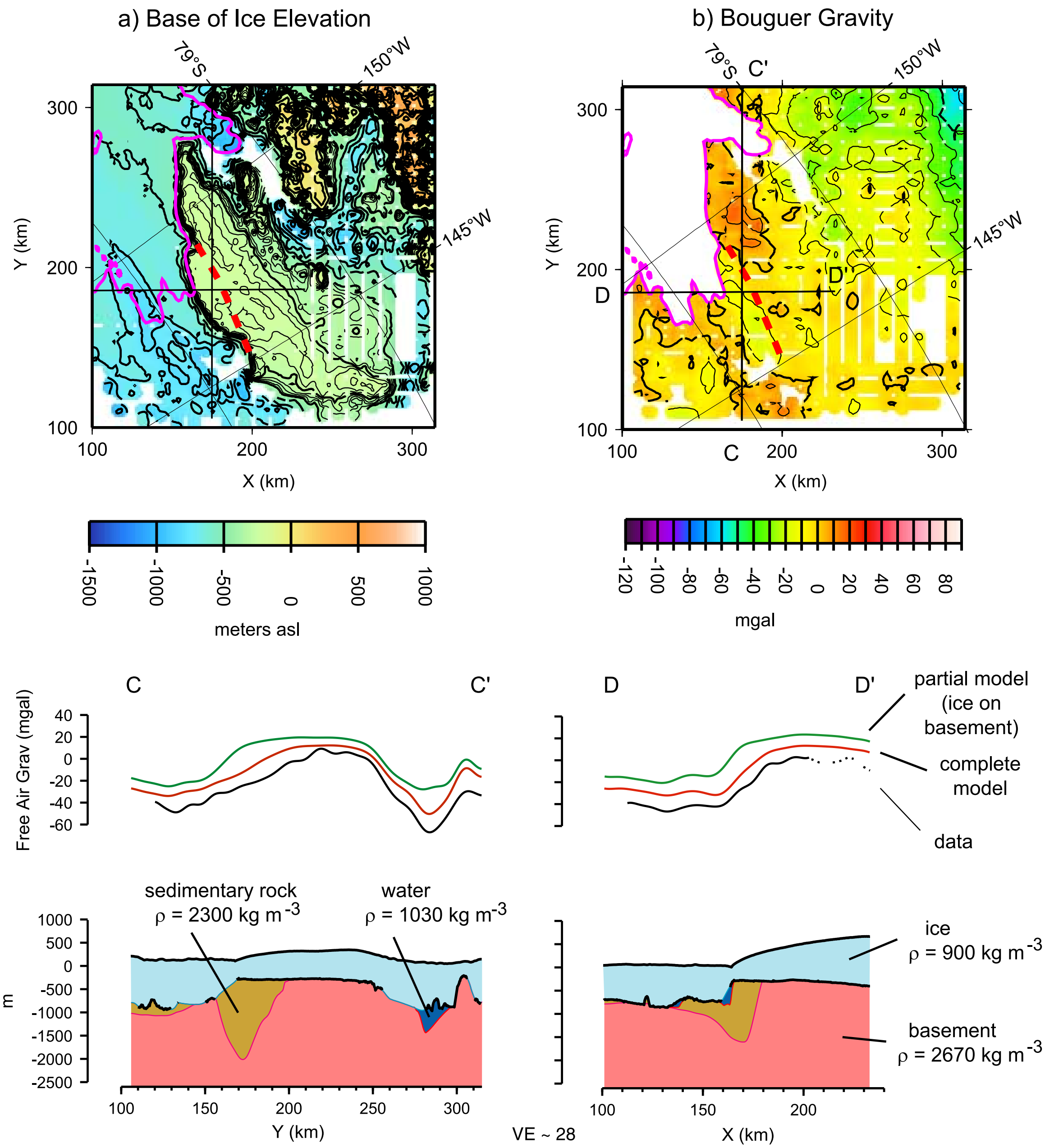

Figure 6. Topography and gravity relations for the Blue Plateau. Location shown in Figure 1. (a) Bed topography is contoured at 100 m plus 25 m (fine) between 200 and 400 mbsl. (b) Bouguer gravity anomaly, accounting for both ice topography and bed topography [*Luyendyk et al.*, 2003], shows a −10 to −20 mgal low at the south edge of the plateau. Comparison of profile data (C-C′ and D-D′) with the 3-D gravity model of *Luyendyk et al.* [2003] shows the anomaly is well modeled by at least 1-km thickness of sedimentary rock. Partial model (green) has ice and basement layers, and complete model (red) includes sediment and water bodies, with gravity predictions offset vertically for clarity. Red dashed line shows inferred contact on plateau surface between thick sediments and thin or absent sediments. Flat topography independent of rock type is more consistent with marine erosion than glacial erosion.

be explained by uniform density bedrock topography. The anomaly at the north edge of the plateau requires high-density bedrock. Modeling of on-ice gravity survey over Roosevelt Island also demonstrates that the gravity field cannot be explained by uniform-density topography (Figure 7b). The edges of the anomaly require high-density bedrock, previously noted by *Greischar et al.* [1992]. A low-density unit, likely sedimentary, is required over the central portion of the plateau. The high densi-

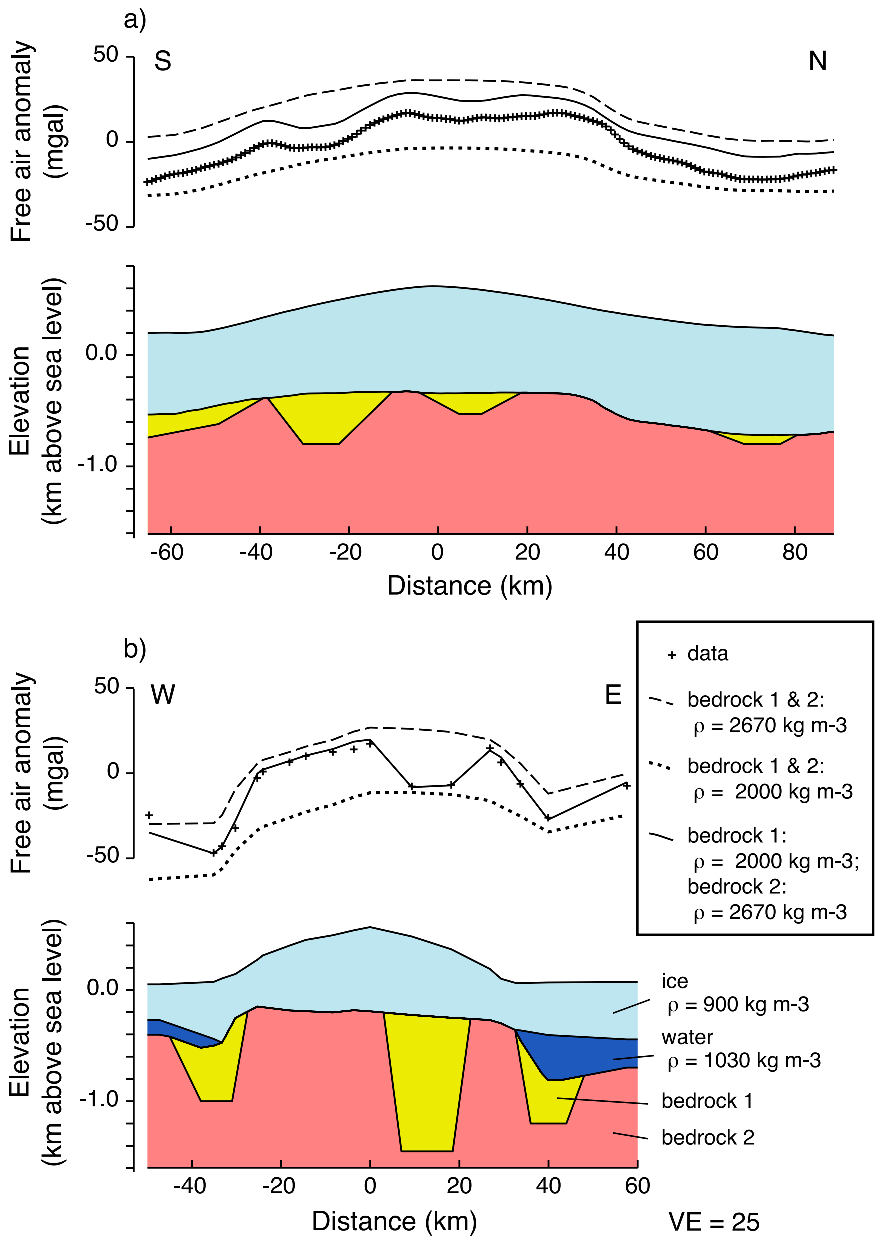

Figure 7. Gravity data and models of (a) Siple Dome and (b) Roosevelt Island. Gravity predictions are shown for uniform bedrock densities of 2000 and 2670 kg m^{-3} and for low-density sediments (yellow = 2000) filling structures in high-density basement (pink = 2670). Data profiles for Siple Dome (Figure 7a) are from an airborne survey, sampling grids of gravity [*Bell et al.*, 1999], ice surface, and bedrock topography data [*Behrendt et al.*, 2004] along four parallel flight lines separated by about 5.3 kilometers (location in Figure 3), then averaging the profiles to minimize noise in airborne gravity. Data from Roosevelt Island (Figure 7b; location in Figure 4) are from a surface survey, with precise relative elevations [*Greischar et al.*, 1992]. Density models include sloping Moho discontinuities at ~20-km depth to account for regional gradients, with slopes of 1% down to south in Figure 7a and 4% down to west in Figure 7b. Uniform low-density bedrock cannot explain the amplitudes of the anomalies. Therefore bedrock plateaus cannot be sub-ice deltas.

ties modeled at the plateaus edges in both these locations argue that crystalline rock forms the edge of the plateaus. Therefore they are eroded, not constructed by sediment accumulation, and cannot be sub-ice deltas. Seismic profiles near Roosevelt Island (Figure 8a) show basement outcrop near the shallowest part of the profile, supporting the interpretation that the island topography is at least partly eroded into basement.

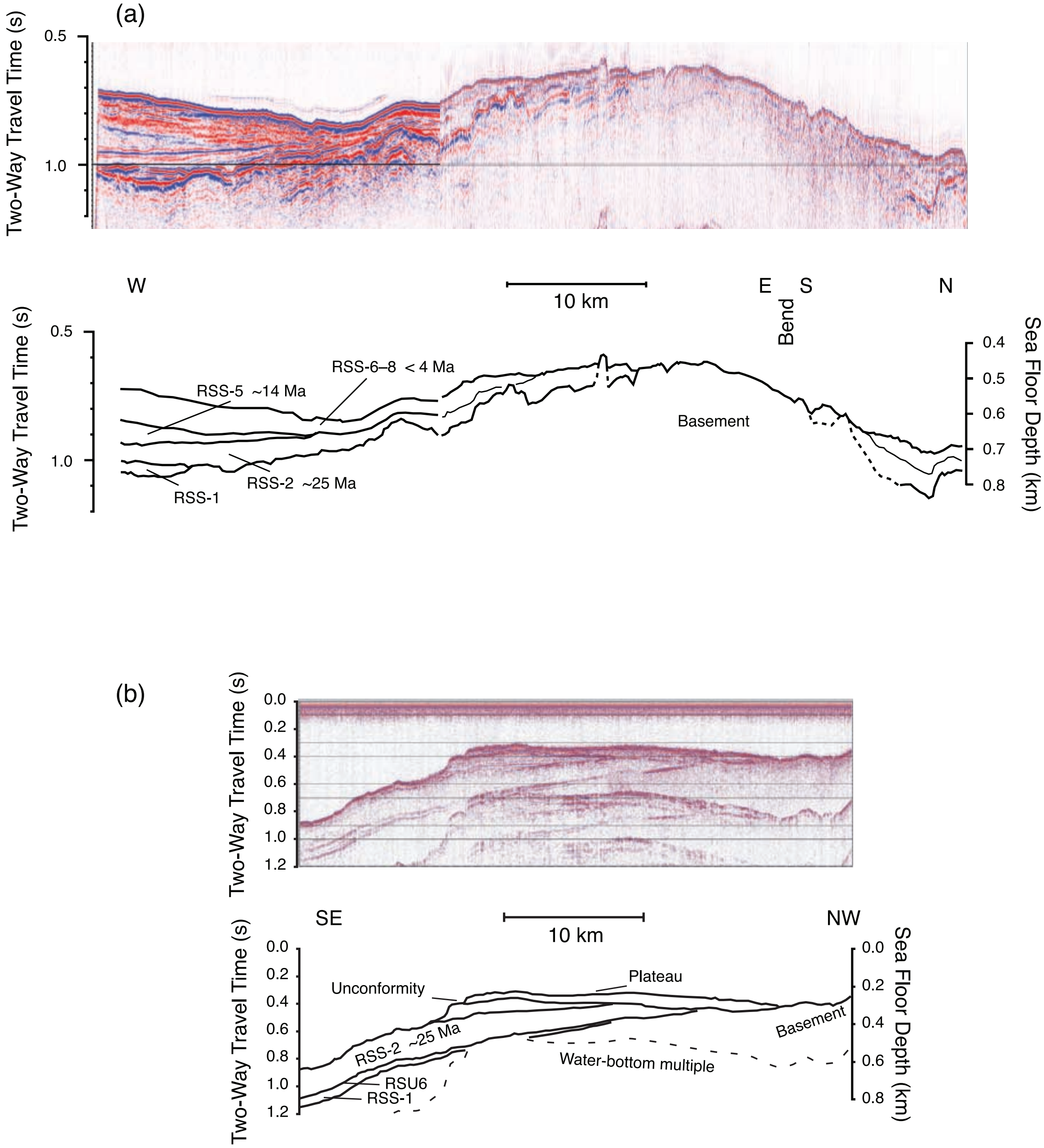

Figure 8. Multichannel seismic data and interpretations for the southeastern Ross Sea. (a) Composite time section north of Roosevelt Island showing basement outcrop at seafloor, indicating that Roosevelt Island is a basement high. Lines NBP0306-110 and NBP0301-28-29; vertical exaggeration ~28X at seafloor. Location in Figures 4 and 10. (b) Time section of line NBP9601-26 showing truncation of upper Oligocene–lower Miocene seismic stratigraphic unit RSS-2, showing that terrace and plateau formation occurred no earlier than early Miocene. Vertical exaggeration ~13X at seafloor. Location in Figures 9 and 10.

[12] All three examples show flat surfaces cut into variable density, and presumably variable type, bedrock. Eroding a continuous flat surface into contrasting rock types is more consistent with marine erosion than glacial erosion, as we discuss below.

[13] Limited sonar mapping in the southeast corner of the Ross Sea near Cape Colbeck [*Luyendyk et al.*, 2001] (Figure 7) shows a small plateau sharing many features in common with the Blue Plateau. Depths along three profiles in an area about 10 × 30 km are exclusively 230–255 m, west of which

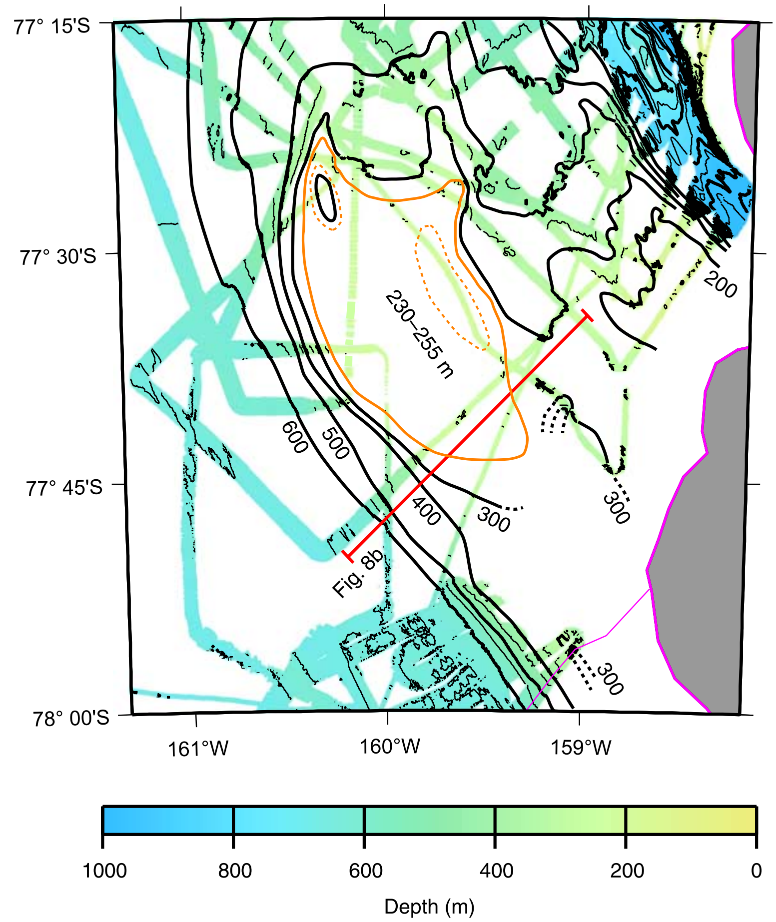

Figure 9. Bathymetry near Cape Colbeck. Location shown in Figure 1. Hand-drawn contours at 100-m interval fill the gaps between automated contours of multibeam data at 50-m interval. Data are from *Luyendyk et al.* [2001] and unpublished cruises NBP9402 and NBP0301. A small (~10 × 30 km) terrace at depth 230–255 m (outlined in orange) is perched above the former ice stream trough, in a setting closely analogous to the Blue Plateau (Figure 3). Location of the seismic profile shown in Figure 8b is highlighted in red.

depths drop steeply to a broad trough with a floor at about 650 m (Figures 8b and 9). The trough is clearly the bed of a former ice stream, passing east of Roosevelt Island and connecting to the current downstream end of Ice Streams D and E between Siple Dome and the Blue Plateau (Figure 2). Seismic reflection data collected in conjunction with bathymetric surveys (Figure 8b) offer some constraint on the age of this surface. This plateau comprises flat-lying sediments capping a level surface cut into dipping sediments that are identified as Ross Sea Stratigraphic Sequence 2 (RSS-2) [*Luyendyk et al.*, 2001], which correlate to upper Oligocene and lower Miocene sediments sampled at DSDP Site 270 (Figure 10). Therefore the surface here was cut during or after the early Miocene.

[14] These plateaus are younger and at lower elevation than the West Antarctic Erosion Surface (WAES) mapped throughout wMBL and correlated to a similar Cretaceous age surface in New Zealand [*LeMasurier and Landis*, 1996]. *Luyendyk et al.* [2001] have correlated the WAES to Ross Sea unconformity RSU7, which is stratigraphically below the plateau near Cape Colbeck and separates RSS-1 upper and lower [*Luyendyk et al.*, 2001].

3. Discussion

[15] We have mapped flat surfaces formed by erosion that are generally far from the shelf edge and near the grounding line, and hundreds of meters below sea level. What needs to be explained is what process cut the surfaces so that they are nearly flat

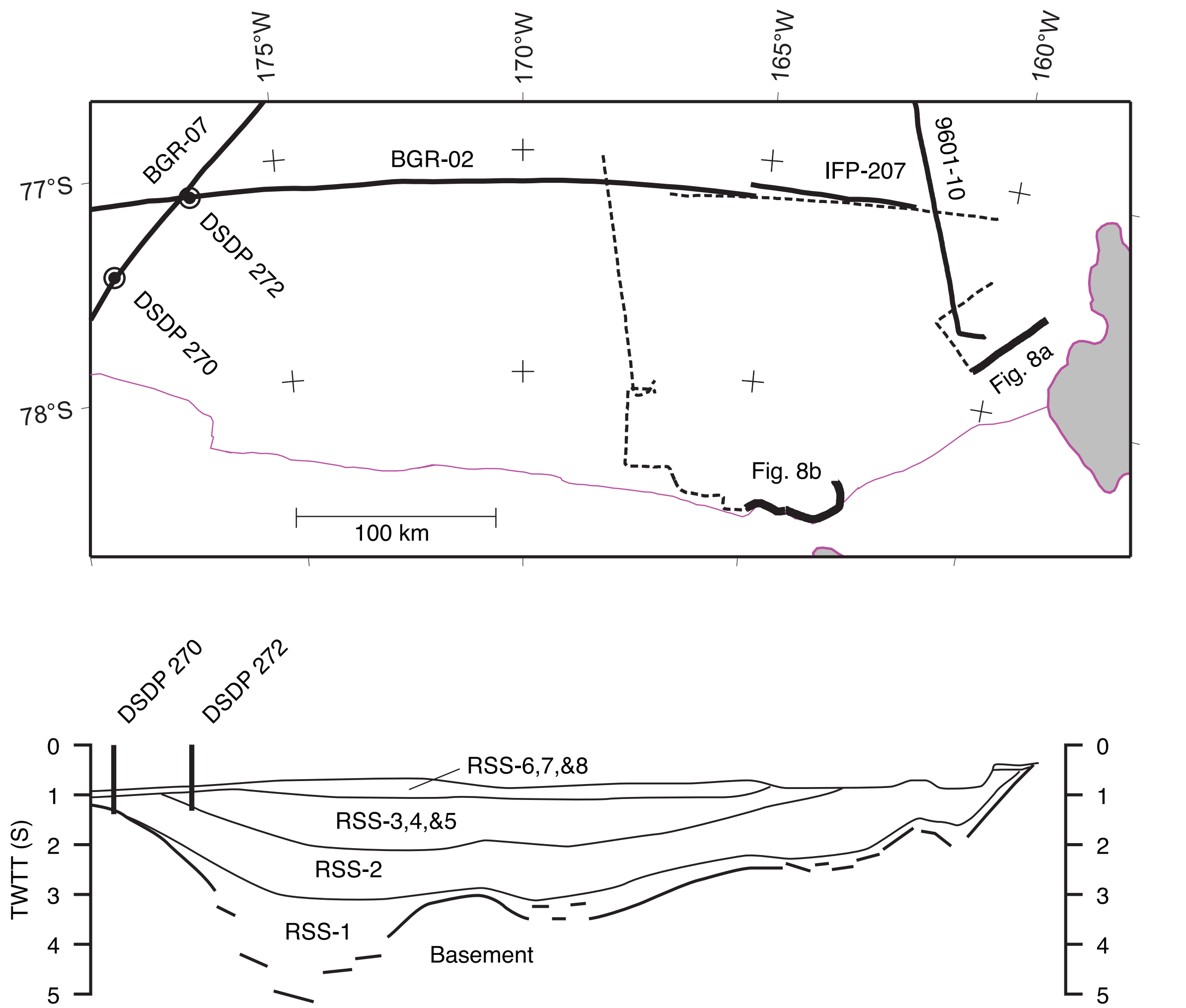

Figure 10. Track map and schematic cross section of the Eastern Basin, illustrating correlation of the Figure 8 seismic sections to DSDP Sites 270 and 272. Bold lines in track map show published sections from *ANTOSTRAT* [1995] and *Luyendyk et al.* [2001]; dashed lines show unpublished NBP9601 and NBP0306 lines used in the correlation. Units RSS-3–RSS-5 are Miocene; RSS-6–RSS-8 are Pliocene-Pleistocene.

and level, why most are so far inland, and why they are below sea level. As a start we review some earlier findings about glacial geomorphology on the Antarctic shelf and some polar margins in the northern hemisphere. We then argue that in some contexts, an origin by marine erosion is a simpler interpretation than an origin by glacial erosion, followed by a discussion of the possible timing of marine erosion. We end with an analysis of subsidence caused by cooling after heating from lithospheric extension that explains the current depths.

3.1. Comparison of Antarctic Shelf Geomorphology With Other Polar Shelves

[16] Glacial erosion typically results in considerable relief on continental shelves. The present-day continental shelf of the Ross Sea (north of the Ross Ice Shelf) shows significant relief due to glacial erosion. *Anderson* [1999] devotes a chapter in his book on Antarctic marine geology to the geomorphology of Antarctic shelves. The Antarctic shelves have great depths, irregular topography, and a landward sloping profile. Beyond the shelf break in the Eastern Basin a broad continental rise appears to have built from sediments scraped off the shelf during Neogene glaciations (Figure 11). Flat-topped banks are found in some places near the shelf edge and are due to ice sheets "advancing across seaward-thickening wedge of sedimentary strata" [*Anderson*, 1999]. In contrast, deep, glacially carved troughs that follow geological boundaries such as contacts and faults, characterize the inner shelf. Deep glacial erosion removes material from the landward shelf and deposits this near the shelf

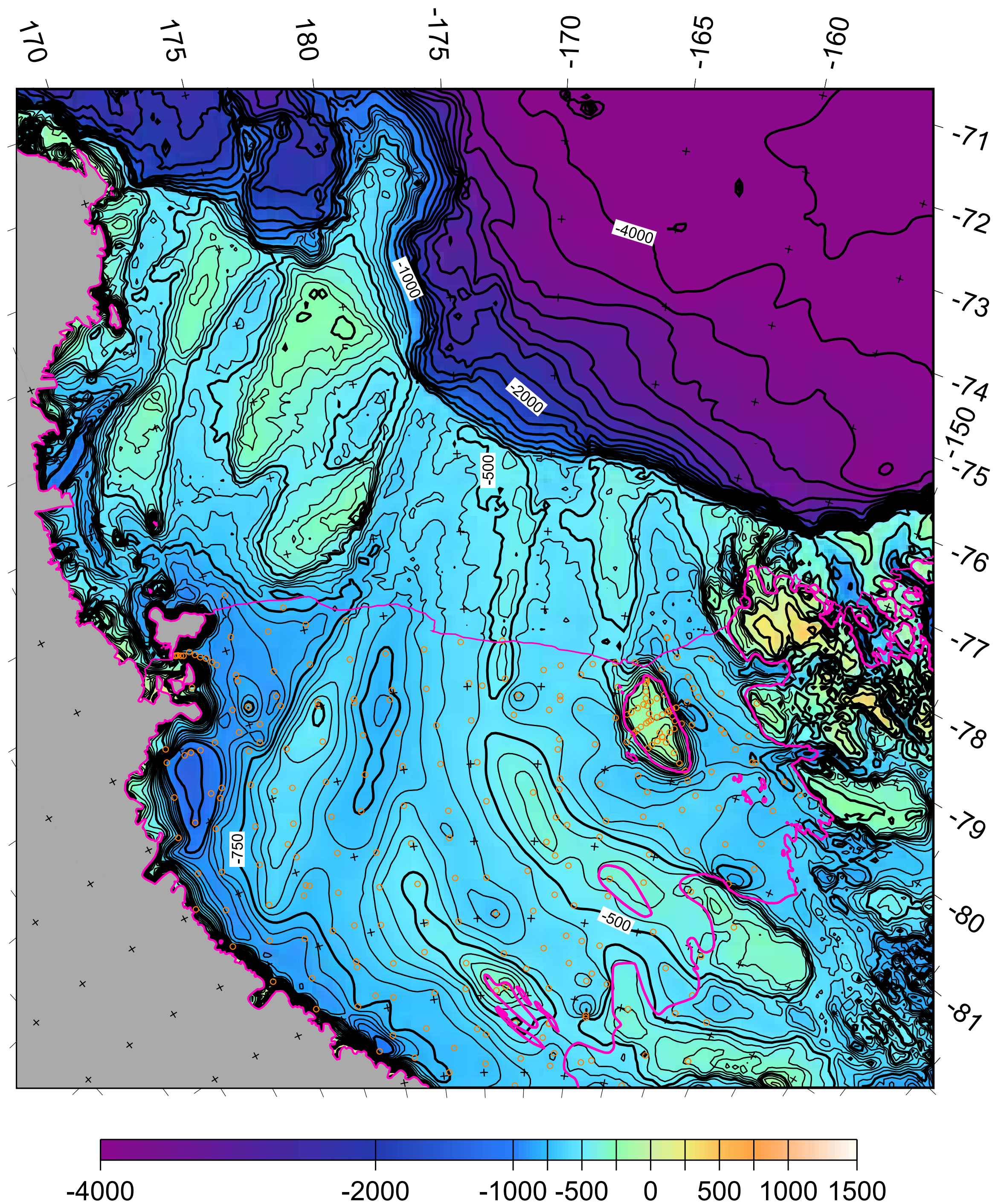

Figure 11. Bathymetry and bed topography for the Ross Sea and Ross Ice Shelf area after *Luyendyk et al.* [2002]. Contour interval is 250 m plus 50 m between 0 and 1000 mbsl. Data sources as in Figure 2, plus deep-water contours from *Davey* [2004] and F. J. Davey and F. Nitsche (Bathymetric grid of the Ross Sea, Antarctica, 2005; available at ftp://ftp.ldeo.columbia.edu/pub/fnitsche/RossSeaBathymetry/). Small orange circles show RIGGS project soundings [*Greischar et al.*, 1992]. Steep sides and flat tops of plateaus in southeast are not present in central and western Ross Sea.

edge and is responsible for the landward slope in the bathymetry.

[17] The broad polar shelf of the Barents Sea north of Scandinavia shows a similar bathymetric relief as the Ross Sea shelf although generally shallower at 100–400 m (IBCAO; http://www.ngdc.noaa.gov/mgg/bathymetry/arctic/arctic.html; accessed 10/20/2005 [*Jakobsson et al.*, 2000]) (Figure 12).

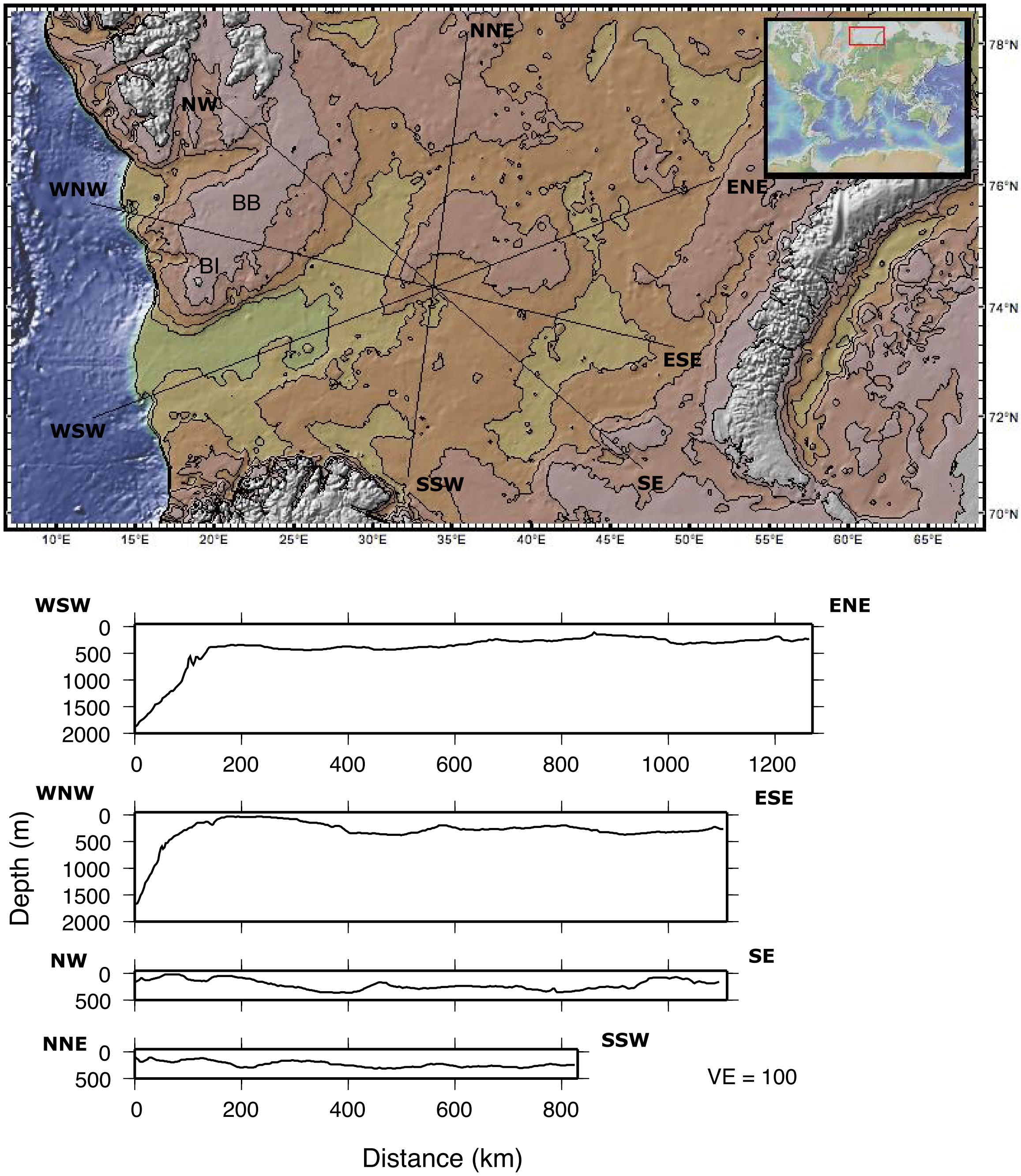

Figure 12. Map of bathymetry of Barents shelf produced via GeoMapApp (http://www.marine-geo.org/geomapapp/; accessed 10/20/2005); 100 m contours, 0–500 m; data source IBCAO (http://www.ngdc.noaa.gov/mgg/bathymetry/arctic/arctic.html; accessed 10/20/2005 [*Jakobsson et al.*, 2000]). Also shown are bathymetric profiles at various azimuths. The banks have uneven tops and gently sloping sides except for Bjornoya bank (BB), a shallow, basement-cored high actively eroding at Bjornoya Island (BI). Extensive deep, flat-topped plateaus such as we mapped in the Ross Embayment are not evident.

Widespread areas of Mesozoic sedimentary deposits here [e.g., *Worsley et al.*, 2001] are more analogous to the Ross Embayment than other polar shelves in cratonic areas with extensive basement near sea level. Banks at 25–200 m depth are separated by broad basins 100–200 km wide and 300–400 m deep. Two major troughs up to 500 m deep feed huge fan complexes beyond the shelf edge. The inner shelf here has been deeply eroded by ice sheets [*Solheim et al.*, 1996].

Shallow banks have uneven topography and lack flat tops (Figure 12), except for Bjornoya bank, which has extensive areas near 25-m depth and is continuing to form by wave erosion, as attested by 400-m sea cliffs in basement at Bjornoya Island [*Worsley et al.*, 2001]. Since glaciation began about 2.5 Ma [*Solheim et al.*, 1996], approximately 1.3 million km^3 of material has been eroded off the Barents shelf by advancing ice sheets [*Rasmussen and Fjeldskaar*, 1996]. Much of this material has been deposited as prograding sedimentary wedges by advancing ice sheets, creating flat-topped fan heads at the mouths of the major troughs on the outer shelf [*Faleide et al.*, 1996].

[18] A review of bathymetry mapped on other polar shelves shows that trough and bank topography to be common, but generally restricted to the outer shelf. Deep troughs, previously occupied by ice streams, dissect these shelves. Surrounding banks are not flat-topped except in some cases near the shelf edge. On the Labrador shelf, banks are 100–200 m deep on the outer shelf, and have about 50 m relief on their tops. They are exclusively on thick passive margin sediments presumed to be mostly early Tertiary in age [*van der Linden and Srivastava*, 1975; *van der Linden et al.*, 1976] covered by minor thickness (~50–100 m) of overlying glacial deposits. Outlet troughs up to several hundred meters deep separate the banks from each other, and shallower troughs on the inland side separate the banks from the coastal basement piedmont. Overall the margin was shaped by passive margin sedimentation, later modified by deep glacial erosion of troughs. Remnant banks have only been slightly modified by deposition of glacial tills.

[19] The submerged Antarctic plateaus bear a resemblance to the *strand flats*, widespread coastal platforms that are characteristic of the coasts of Norway and other polar margins, as noted by *Nansen* [1922]. The Norwegian strand flats are exposed up to 50 meters above sea level and inland some 50 kilometers [*Klemsdal*, 1982]. Their origin remains controversial. Discussing the strand flats of the Antarctic Peninsula and nearby islands, *John and Sugden* [1975] and *Anderson* [1999] mention many competing theories including wave erosion, glacial erosion, and erosion by floating ice, all constrained to operate near sea level. Though acknowledging possibly complex histories, *John and Sugden* [1975] summarized that the platforms of the Antarctic Peninsula area "have morphological characteristics which point unequivocably to a marine origin."

3.2. Mechanism of Formation of Plateaus and Terraces

[20] The plateaus we mapped are close to the grounding line, far from the outer edge of the shelf, and they were formed by erosion rather than deposition (Figures 6, 7, and 8). The Blue, Roosevelt, and Siple Dome plateaus discussed above are 250–600 kilometers inboard of the shelf edge and all formed by erosion. Two obvious candidates for a mechanism to form these surfaces are glacial erosion and marine erosion at wave base to form a wave-cut platform. Glacial erosion is potentially a simple interpretation in that ice is generally regarded as having covered the Antarctic continent for nearly all of the last several million years and may have been extensive for most of the time since ~30 Ma. However, for several reasons discussed below relating erosion processes to the patterns of plateau occurrence, we favor wave erosion. As with essentially any discussion of erosion surfaces, the need to guess the topography prior to erosion leaves ample room for differences of opinion.

[21] Could the platforms have formed in the present glacial regime? Evidence is very clear that they could not. Ice currently grounded on the platforms is moving very slowly [e.g., *Joughin et al.*, 1998], therefore not causing significant erosion. The well-studied case of Siple Dome offers strong evidence that the bedrock plateau there (Figure 3) is not a result of recent glacial action. A GPS survey of ice-surface motion shows exclusively slow, downhill motion, normal to the east-west ice divide except adjacent to the ice streams [*Nerenson et al.*, 2000]. The character of the radar reflection from the base of the ice indicates that the ice is frozen to the bed [*Gades et al.*, 2000]. Modeling of the internal layering of the ice mapped by ground-based radar demonstrates nearly steady state conditions for several thousand years, with internal deformation of the ice driven exclusively by stresses resulting from surface topography and no motion of the ice across the bed [*Nerenson et al.*, 2000]. Under these conditions, neither erosion nor deposition will occur with ice frozen to the bed.

[22] Were conditions very different at Last Glacial Maximum (LGM), could the bedrock have been beveled then by glacial erosion? Several lines of evidence, especially from the ice core recovered at Siple Dome from 1997–1999, indicate that differences were minor compared to the present. The

Siple Dome ice core penetrated the entire 1004-m thickness of ice, with the age of the ice determined by counting annual layers to 8.2 ka at a depth of 514 m and by correlating gas content with other dated cores to 30 ka at about 800 m [*Taylor et al.*, 2004; *Waddington et al.*, 2005]. Using climate constraints from other ice cores, *Waddington et al.* [2005] showed that the current age-depth pattern can only be reasonably explained if the surface elevation was only 200–400 m higher than present at 25–16 ka, and thinning to present thickness by 2 ka. *Parizek and Alley* [2004] argue that the isotopic composition of the entire ice core, to ~90 ka, requires deposition at relatively low elevations and precludes transport of ice from interior highlands to the core site. *Steig et al.* [2001] interpret the ice core record from Byrd Station (80°S, 120°W; Figure 1) in western Marie Byrd Land as showing only 200 m of thinning since 8 ka. Changes in ice thickness measured at relatively inland nunataks also show that changes have been minor. *Ackert et al.* [1999] interpret a lowering of the ice surface of 45 m since 10 ka in the Executive Committee Range (Figure 1). All available evidence suggests that ice was not much thicker than at present while the air temperatures were colder, implying that the ice was frozen to the bedrock plateaus during LGM as it is today. Therefore ice grounded on the divides (plateaus) between ice streams would be expected to be just as immobile as at present, and not eroding the bed.

[23] Is it plausible that the platforms eroded sometime in the early history of the West Antarctic Ice Sheet (WAIS)? We believe that the relations around Roosevelt Island suggest that glacial erosion is not likely. The bedrock of the southern Ross Sea west and northwest of Roosevelt Island contains thick late Oligocene to middle Miocene sedimentary rocks [*ANTOSTRAT*, 1995], with the younger units in the sequence pinching out approaching Roosevelt Island at depths deeper than 700 mbsl (Figure 8a), well below the 200 mbsl level of the terrace on northern Roosevelt Island and only 30 km away (Figure 4). Sediments northeast of Roosevelt Island are poorly dated, but a sedimentary basin had already formed in the area by late Oligocene (Figure 8b). Geophysical evidence summarized by *Greischar et al.* [1992] at station RI south of Roosevelt Island (Figure 4) shows sediment thickness varying from 300–800 m. Sediments under the Ross Ice Shelf are only minimally sampled, but it is a conservative extrapolation of relations in the Ross Sea to identify Roosevelt Island as a structural high that stood hundreds of meters above surrounding sedimentary basins in the late Oligocene through middle Miocene. Although ice sheets may well have advanced into the Ross Sea sedimentary basins during this interval [*Anderson and Bartek*, 1992; *Bart*, 2003], and warmer temperatures may have led to more extensive wet-based (erosive) ice than in the present climate, a mechanism for glacial erosion to carve a flat surface on the top of a basement high above the adjacent basins is lacking.

[24] There are many examples from glaciated northern-hemisphere sites of flat plateaus above regions of obvious glacial erosion. *Sugden and John* [1976, p. 195] term such areas as landscapes of selective linear erosion, where "ice erosion has been concentrated in a trough or series of troughs and has left the intervening slopes or plateaux unmodified." They describe a generally sharp transition between striated steep slopes and plateau areas that "may be regolith covered and devoid of glacial erosion forms." Their examples are from diverse settings including New York, East Greenland, Iceland, and Scotland. These patterns are reasonable consequences of the consensus view that glacial erosion requires a shear traction to drive ice across the bed as well as a supply of entrained rock in the base of the ice, and is greatly accelerated by the presence of water to lubricate and flush rock flour. Once troughs start to form in a region of subdued topography, the new slopes will enhance traction, and the supply of rock fragments and water will both be concentrated at the base of the troughs by downhill motion. Similar positive feedback effects have been discussed as a mechanism of forming ice stream troughs on the Barents shelf [*Siegert and Dowdeswell*, 1996].

[25] There is a widespread perception that glacial erosion is an effective agent for generating flat bedrock surfaces. This perception appears to result from the association of flat topography and recent glaciation of the Canadian and Baltic Shields. However, it is not the consensus of glacial geomorphologists that glaciation is responsible for the shield topography. The review of *Embleton and King* [1975] instead shows that for most parts of the Canadian Shield, glacial erosion has been minor, limited to removing platform sediments to expose the Paleozoic unconformity. In certain areas, for example the Lake Superior basin, glacial erosion was certainly much more important, but these areas are not especially flat. In areas of high relief, for example the Alps or the Transantarctic Mountains [*Stern and Baxter*, 2002] glacial erosion

acts primarily by deepening and widening existing valleys, with the highest topography becoming increasing rugged as adjacent valleys approach each other.

[26] Cenozoic glaciation in West Antarctica has generally not resulted in a leveling of the preexisting bedrock topography. In the Late Cretaceous widespread extensional rifting strongly modified structures formed during earlier convergent orogenesis [*Davey and Brancolini*, 1995; *Luyendyk et al.*, 2001]. Parts of West Antarctica may also have be subject to Eocene-Oligocene extension [*Cande and Stock*, 2004; *Cande et al.*, 2000]. A basin-and-range morphology with range length of 25–40 km, range spacing of $\sim$20 km, and peak-to-trough amplitude of $\sim$1000 m, consistent with a horst-and-graben origin, dominates the bed topography of western Marie Byrd Land [*Luyendyk et al.*, 2003]. In the adjacent eastern Ross Sea near 160°W, *Luyendyk et al.* [2001] mapped basin-and-range basement topography buried under Oligocene and Miocene sedimentary deposits. Knowledge of basement structure under the Ross Ice Shelf is, of course, extremely limited, but at the five sites with adequate seismic and gravity data to determine basement relief reviewed by *Greischar et al.* [1992], all showed basement relief of at least 400 m, buried by lower-density sediments. The similarity of buried basement topography in the eastern Ross Sea to bed topography in western Marie Byrd Land implies that glacial erosion has not substantially reduced the amplitude of the onshore bed topography. In fact, onshore trough depths extending well below sea level suggest that glacial erosion has been most effective at removing sediments from the troughs and not in leveling the ridges. The similarity of relief also implies that intervening areas that are now basement platforms more than 30 km wide can reasonably be expected to have previously supported ranges rising several hundred meters above the current level. In terms of a mechanism that explains which range peaks are preserved or eroded and which trough fills are preserved, a marine erosion mechanism offers a simpler scenario than a glacial erosion mechanism.

[27] The platforms show several features that would be expected to result from wave erosion in the absence of coastal ice. The position of several platforms fringing the wMBL highlands is exactly what would be expected if the extent of the highlands had been somewhat reduced by wave erosion. No platforms are present in locations that would be strongly sheltered from wave action in the absence of sea ice or grounded coastal ice. Subtle slopes on the platform surfaces have highest topography on the northeast to southeast regions that would be least exposed to wave action (Figures 3–5 and 9). The bed topography in the Sulzberger Ice Shelf area in particular shows features that can only be simply interpreted as products of wave erosion, especially the sharp transitions between smooth terraces and higher rough topography (relict sea cliffs, Figure 5). The very gradual southward increase in depth of the platforms suggests formation at a common elevation followed by gentle regional tilt.

3.3. Timing Constraints on Age of Formation of Plateaus and Terraces

[28] Tight constraints for the age of formation of the plateaus and terraces are lacking. Oligocene foraminera from lower RSS-2 at DSDP Site 270 indicate an age of roughly 25 Ma for the eroded sediment imaged in Figure 8b [*Hayes and Frakes*, 1975]. A review of Cenozoic global climate inferred from deep-ocean cores [*Zachos et al.*, 2001] describes the record after formation of large Antarctic ice sheets in the earliest Oligocene: "These ice sheets persisted until the latter part of the Oligocene (26 to 27 Ma), when a warming trend reduced the extent of Antarctic ice. From this point until the middle Miocene ($\sim$15 Ma) global ice volume remained low and bottom water temperatures trended slightly higher..., with the exception of several brief periods of glaciation (e.g., Mi-events)...This warm phase peaked in the late [to] middle Miocene climatic optimum (17 to 15 Ma), and was followed by a gradual cooling and reestablishment of a major ice sheet on Antarctica by 10 Ma..." The $\delta^{18}O$ proxy reaches its Neogene minimum in many records just prior to the Mi-2 event, $\sim$16 Ma [*Miller et al.*, 1991], reflecting some combination of warm water temperature and low ice volume. Local records offer limited constraints on the possibly different histories of East and West Antarctic Ice Sheets. As judged from the presence of ice-rafted debris in the deep ocean east of the Antarctic Peninsula (Ocean Drilling Program (ODP) Leg 113), the WAIS reached close to its present extent about 8 Ma [*Kennett and Barker*, 1990]. More recent drilling west of the Antarctic Peninsula (ODP Leg 178), however, shows abundant ice-rafted debris in the oldest recovered sediments at 10 Ma [*Barker and Camerlenghi*, 2002].

[29] The record from a few short cores taken through the southern Ross Ice Shelf (site J-9, Figure 1) offers a unique perspective [*Harwood et al.*, 1989]. The recovered diamictites contain abundant small clasts of middle lower Miocene (~21–18 Ma) diatomite, a single clast and significant matrix material of early middle Miocene diatomite (~16–15 Ma), and matrix material containing middle upper Miocene (~8–6 Ma) diatomaceous material. Because diatom growth requires sunlit open water, Harwood et al. interpret an "open marine environment in West Antarctica with intense productivity and limited glacial ice at sea level in the Ross Sea embayment," at least for limited time intervals indicated by the age of the recovered diatoms.

[30] The interval 20–14 Ma corresponds to Ross Sea units RSS-3–RSS-5, sampled as glacial marine sediments at DSDP Site 272 [*DeSantis et al.*, 1999; *Hayes and Frakes*, 1975]. Unconformities at the top of these units are of uncertain origin, interpreted by some as a result of plowing by grounded ice sheets [*Anderson and Bartek*, 1992; *Bart*, 2003]. The depth of the unconformities at 700 mbsl and deeper near Roosevelt Island (Figure 8a) is very difficult to reconcile with an origin near sea level. Therefore these unconformities must have been cut by grounded ice. Our interpretation of a wave-cut origin for platforms as far inland as Siple Dome requires intervals adding up to probably at least several hundred thousand years without ice covering the coastline to have sufficient time to erode these large surfaces. (The estimate of cumulative time is based on observing terraces of width on the order of hundreds of meters at active margins, where the rapid uplift limits the duration of wave erosion to ~10 kyr at a given elevation during highstands or lowstands.) Unconformities formed by grounded ice in deep water must erode at a different time in the glacial cycle from inland, wave-cut platforms. Because the southern ocean was warmer prior to 14 Ma than at present [e.g., *Shevenell et al.*, 2004], it is reasonable to speculate that during intervals of glacial retreat, ice would not blanket the coastal areas, possibly similar to portions of the Greenland coast today. In the coldest or wettest intervals, ice accumulating at high elevations could follow paths similar to the modern ice streams to erode the bed of the Ross Sea. Within these limited constraints, the ice-free conditions at the coast necessary for our preferred interpretation of wave erosion are at least plausible in limited intervals anytime in the interval 20–10 Ma. An ice-free coast for significant time intervals is probably easiest to reconcile with the global record of warm temperatures and high sea level during the Miocene climatic optimum at 17–15 Ma [*Miller et al.*, 2005; *Zachos et al.*, 2001].

3.4. Tectonic Subsidence and Postglacial Rebound

[31] One obvious test of the hypothesis that the plateaus and terraces were formed by wave erosion is whether plausible mechanisms exist to bring the plateaus from near early to middle Miocene sea level to their current depths of 100–350 mbsl. Early to middle Miocene sea level was close to present sea level [*Miller et al.*, 2005], so we do not consider eustatic changes. A mechanism for true subsidence appears necessary, and possible candidates are ice loading, sediment loading, crustal thinning, and lithospheric cooling. Mapping of sediment distribution in the Ross Sea shows that significant accumulations of middle Miocene and younger sediments are restricted to the northern and western Ross Sea, so we do not consider the sediment loading mechanism further. An interpretation of crustal thinning of Miocene or younger age is difficult to reconcile with the general interpretation of lack of extension of either oceanic lithosphere surrounding the Antarctic continent or Ross Sea sediments for that age range, so we also discount that mechanism, and investigate an explanation using a combination of ice loading and lithosphere cooling and contraction.

[32] Our interpretation of up to 350 m of subsidence for the wave cut platform can be tested quantitatively using simple ice loading and thermal contraction models. The subsidence caused by the increase in ice load is largely determined by the difference in weight between the current ice load and the Miocene ocean load. Additionally, the viscoelastic properties of the Earth's mantle lead to a delay of about 20–30 thousand years between a change in load and achieving the equilibrium vertical displacement due to the new load. Because of this delay, the predicted deflection due to ice load must also account for recent changes in the ice load, or ice sheet retreat since the LGM.

[33] We assume that any subsidence that cannot be modeled by ice loading is tectonic, driven primarily by thermal contraction from cooling since the last episode of lithosphere extension with accompanying heating. Estimating how much subsidence has occurred due to increase in ice load since a warm climate interval in the middle Miocene prior to 14 Ma requires two separate estimates: the

equilibrium deflection due to the difference between the Miocene ice load and the present ice load, plus the effect of the load of ice removed over the last several thousand years since the LGM that still depresses the current surface due to viscoelastic memory in the Earth's mantle. An upper bound on the subsidence caused by the present ice load can be safely calculated by assuming that no ice was present in the past, with moderate uncertainties resulting from the poorly known effective elastic thickness of the West Antarctic lithosphere. Uncertainties due to viscoelastic rebound are difficult to evaluate. Accepting the recent interpretations discussed above that ice thickness near Siple Dome has only changed about 200 m since LGM limits the remaining rebound from this change in ice thickness to a fairly minor effect.

[34] We illustrate the combined effect of extrapolated rebound and removal of present ice in Figure 13. For viscoelastic rebound, we use the ice sheet model that is most consistent with the recent estimates of the ice cover during the LGM. The D91 model of *James and Ivins* [1998], calculated from the ice-volume history of *Denton et al.* [1991] is most appropriate, but still may overestimate the amount of ice removal near Siple Dome. We illustrate the effect of lithosphere thickness on determining the rebound resulting from ice still present by adding the rebound resulting from removing all West Antarctic ice to the D91 viscoelastic rebound model for effective elastic thicknesses of 25, 40, and 80 km. The calculations are based on a 2-D grid using the BEDMAP ice thickness, applying an upward force where ice is removed and a downward force where ice is replaced by seawater. Iteration is used to account for the modification of the water load by rebound. The combined effects of both rebound mechanisms bring the platforms north of 80°S close to sea level, but leave the Siple Dome plateau at 130–170 mbsl.

[35] Ice loading can only account for some of the elevation drop of part of the platform to its present depth, and thermal contraction is the most viable candidate for the remaining part. *McKenzie* [1978] has outlined the important principles for subsidence resulting from postextensional cooling, which is by far the most effective mechanism for producing subsidence in a recently extended area such as the West Antarctic Rift System [e.g., *Behrendt et al.*, 1991a]. During distributed extension, the average surface elevation subsides due to the isostatic response to crustal thinning. After extension, subsidence continues for tens of million years due to cooling of the hot asthenosphere that wells up during extension. *McKenzie*'s [1978] simple numerical model for subsidence resulting from instantaneous, pure-shear extension emphasizes the importance of extension age and stretching factor in predicting basin subsidence, as other parameters are well constrained by matching the subsidence rate of oceanic lithosphere [*Parsons and Sclater*, 1977] for the case of an infinite stretching factor.

[36] According to this model, cooling after Cretaceous rifting would not produce an adequate elevation drop because, if extension ended at that time, the rate of subsidence would be too slow during Miocene and later time to produce the required subsidence. However, early Cenozoic seafloor spreading in the Adare Trough [*Cande et al.*, 2000] north of the western Ross Sea (Figure 1) may be related to continental subsidence. If seafloor spreading in the Adare trough was accompanied by distributed continental extension in the West Antarctic Rift System including the western Ross Sea and Siple Coast region, the 43–26 Ma age of the spreading would be recent enough to explain the post-15-Ma subsidence of the platform. Using the McKenzie model, an extension age of 35 Ma predicts subsidence of 110 m for 15–0 Ma assuming a stretching factor of 1.25, and 200 m for a stretching factor of 1.5. The current width of the Siple Coast is about 600 km from the Edward VII Peninsula to the TAM, and the Adare Trough extension measured by *Cande et al.* [2000] is 180 km, implying a stretching factor of 1.4 if comparable extension were distributed across the Siple Coast region. Cooling in response to the same extension event observed at Adare trough is therefore sufficient to explain the ~150 m component of subsidence of the Siple Dome plateau that is not explained by ice loading. Cooling in response to extension necessarily is associated with heat flow above the background value. The stretching factors of 1.25–1.5 mentioned above would correspond to additional heat flow of 8–15 mWm^{-2} [*McKenzie*, 1978], which may be a factor in providing the heat needed for water to lubricate the base of the modern ice streams.

3.5. Implications of Crustal Subsidence for Climate History

[37] *DeConto and Pollard* [2003] have pointed out that high paleoelevations on Antarctica were a crucial factor influencing the growth of Cenozoic ice sheets. Because of potentially widely different

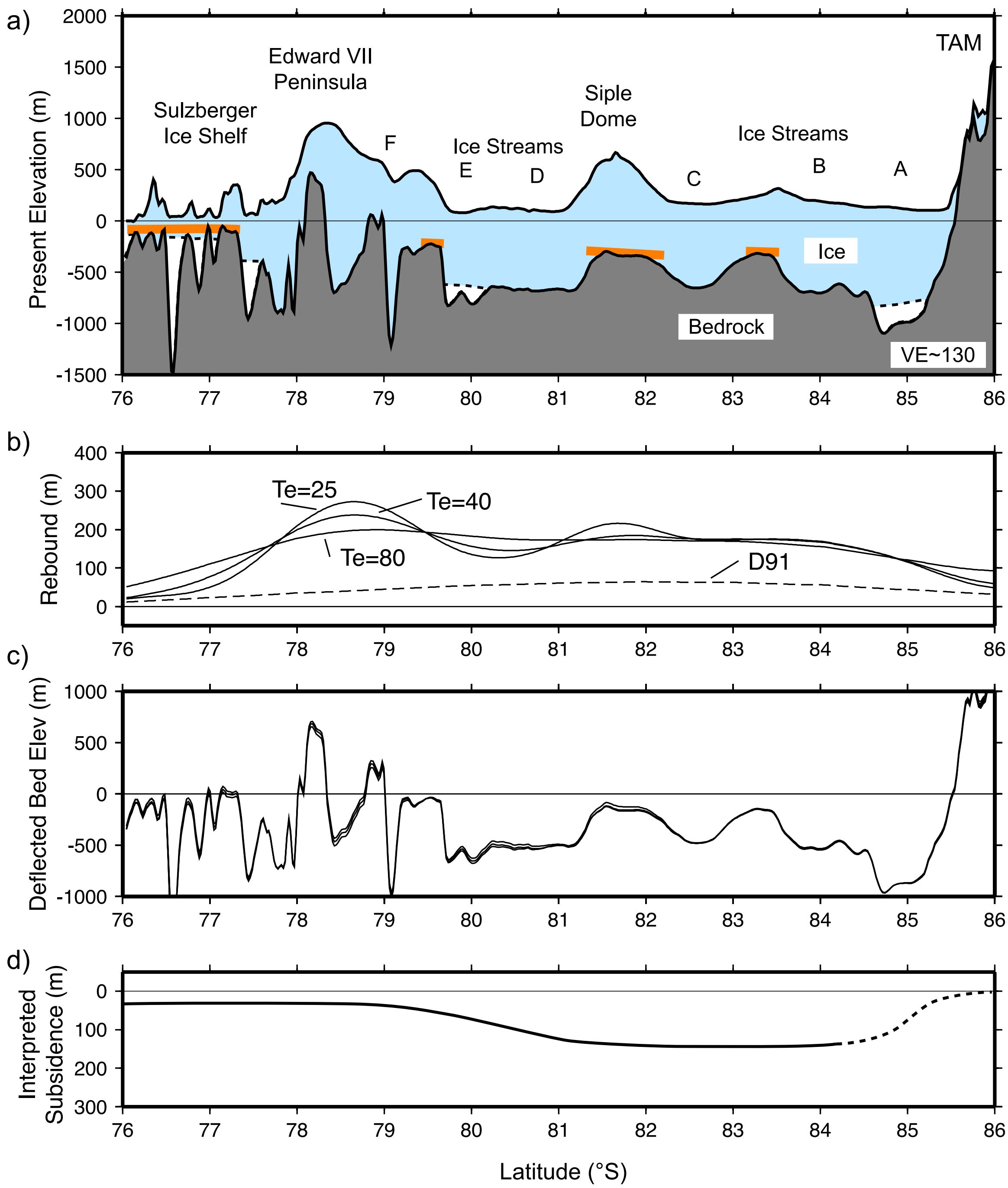

Figure 13. Profile following ~149°W (Figure 2) showing (a) current ice and bed surfaces, (b) rebound models, (c) bed surface deflected by the rebound models, and (d) interpreted postplateau thermal and tectonic subsidence. Bold orange lines highlight observed flat bed surfaces. Rebound models (Figure 13b) show extrapolated viscoelastic component (dashed) of the D91 model [*Denton et al.*, 1991; *James and Ivins*, 1998] and the combined effects of the D91 model and the removal of all West Antarctic ice for effective elastic thicknesses (Te) of 25, 40, and 80 km. Deflected bed elevations (Figure 13c) show the range of predictions for the viscoelastic model and the three elastic thicknesses. Deflected elevation of the 340-mbsl plateau surface at Siple Dome ranges between 130 and 170 mbsl, depending on the rebound model. The maximum interpreted subsidence of the plateaus at 81–84°S corresponds to the deepest part of the profile, demonstrating the importance of postextension cooling as a factor in creating the deep Ross Embayment.

predictions for paleoelevations, the issue of whether West Antarctica has been subsiding due to conductive cooling as we conclude here along with *Luyendyk et al.* [2001], or rising due to mantle plume activity [*LeMasurier and Landis*, 1996] could have global significance for paleoclimate. Studies of Late Cretaceous sea level histories show oscillations interpreted as controlled by growth and retreat of glaciers, despite a globally warm climate [*Gale et al.*, 2002; *Miller et al.*, 2004]. Estimated maximum ice volume variation from *Gale et al.* [2002] and *Miller et al.* [2004] based on 20-m amplitude of sea level oscillation is about 10^7 km^3, or roughly one fourth of the current Antarctic ice volume. The models of *DeConto and Pollard* [2003] suggest that CO_2 concentrations between 2 and 3 times recent preindustrial values are necessary to allow that much ice to form during the Eocene on Antarctic topography inferred from adjusting present topography for glacial loads. Records of $\delta^{18}O$ and limited CO_2 proxies imply that the Late Cretaceous climate was warmer, with higher CO_2 concentrations than the Eocene [e.g., *Retallack*, 2002]. A possible factor that may be necessary for forming large volumes of ice in the Cretaceous is that much of West Antarctica may have been at high elevation then [*Luyendyk et al.*, 2001]. West Antarctica was a convergent orogen in the Early Cretaceous, contiguous with the Andes and perhaps similar to current morphology there. If the continental extension we interpret as synchronous with Eocene-Oligocene Adare Trough spreading is only the last of several extension episodes in the Late Cretaceous and early Tertiary [*Behrendt et al.*, 1991a; *Cande and Stock*, 2004; *Studinger et al.*, 2004], all of which involved crustal thinning and subsequent cooling, extensive surfaces at elevations of 2–4 km in the Late Cretaceous would be expected, adequate to host mountain glaciers that sometimes may have coalesced to form an ice cap that significantly affected sea level.

4. Conclusions

[38] Glacial erosion seems implausible as a cause of the marine and sub ice plateaus we mapped around the eastern and southern borders of the Ross Embayment. Wave erosion, however, can explain the flatness and position of plateau remnants around margins of coastal wMBL. Formation of the plateaus in the middle Miocene is consistent with gentle tilt of a lower Miocene substrate, allows time for subsidence of the surface after formation, and probably is consistent with ice-free conditions at sea level, thereby permitting wave erosion. Glaciers and ice streams probably have since cut into the surface, dissecting it and causing continuing erosion. The present depths of the plateaus, up to 350 meters below sea level, require that they have been subsiding since they were created during middle Miocene. Some subsidence beyond that due to Pleistocene ice loading is needed to explain their present depths. Lithosphere cooling and contraction since Late Cretaceous Ross Sea rift extension is not adequate. Tertiary extension and subsequent thermal subsidence within the Ross Embayment, is required to explain their present depths. This argument in turn strengthens the role of Oligocene and younger extension as a significant influence on vertical tectonics in this region.

[39] The ice-free coastal conditions we postulate as present when the plateaus formed may have been similar to the current exposed coastal lowlands of Greenland today. The plateaus we describe are wave cut platforms similar to uplifted strand flats seen fringing the coasts of Norway. If sometime in the future the Ross Embayment plateaus were to emerge above sea level the Siple coast of West Antarctica would comprise strand flats and fjords and resemble that of Scandinavia.

Acknowledgments

[40] We thank the SOAR project of the University of Texas Institute for Geophysics for collection and processing of the airborne radar data, and the captain, crew, and technical staff of the RVIB *N.B. Palmer* for collecting seismic and bathymetric data. We thank Tom James for providing gridded rebound models, Chris Sorlien and Robbie Decesari for assistance with offshore seismic data, Jim Kennett and Amelia Shevenell for discussions, and John Behrendt and two anonymous reviewers for constructive comments. Supported by U.S. National Science Foundation grants OPP9615281 and OPP0088143. Contribution of the Institute for Crustal Studies 0793.

References

Ackert, R. P., Jr., D. J. Barclay, J. H. W. Borns, P. E. Calkin, M. D. Kurz, J. L. Fastook, and E. J. Steig (1999), Measurements of past ice sheet elevations in interior West Antarctica, *Science*, *286*, 276–280.

Anderson, J. B. (1999), *Antarctic Marine Geology*, 289 pp., Cambridge Univ. Press, New York.

Anderson, J. B., and L. R. Bartek (1992), Cenozoic glacial history of the Ross Sea revealed by intermediate resolution seismic reflection data combined with drill site information, in *The Antarctic Paleoenvironment: A Perspective on Global Change, Part One*, *Antarct. Res. Ser.*, vol. 56, edited by J. P. Kennett and D. A. Warnke, pp. 231–263, AGU, Washington, D. C.

ANTOSTRAT (1995), Seismic stratigraphic atlas of the Ross Sea, in *Geology and Seismic Stratigraphy of the Antarctic Margin*, *Antarct. Res. Ser.*, vol. 68, edited by A. K. Cooper, P. F. Barker, and G. Brancolini, 22 plates, AGU, Washington, D. C.

Barker, P. F., and A. Camerlenghi (2002), Glacial history of the Antarctic Peninsula from Pacific margin sediments [online], *Proc. Ocean Drill. Program Sci. Results*, *178*, 40 pp. (Available at http://www-odp.tamu.edu/publications/178_SR/VOLUME/SYNTH/SYNTH.PDF)

Bart, P. J. (2003), Were West Antarctic ice sheet grounding events in Ross Sea a consequence of East Antarctic ice sheet expansion during the middle Miocene?, *Earth Planet. Sci. Lett.*, *216*, 93–107.

Behrendt, J. C., W. E. LeMasurier, A. K. Cooper, F. Tessensohn, A. Trehu, and D. Damaske (1991a), Geophysical studies of the West Antarctic Rift System, *Tectonics*, *10*, 1257–1273.

Behrendt, J. C., W. E. LeMasurier, A. K. Cooper, F. Tessensohn, A. Trehu, and D. Damaske (1991b), The West Antarctic Rift System: A review of geophysical investigations, in *Contributions to Antarctic Research II*, *Antarct. Res. Ser.*, vol. 53, edited by D. H. Elliot, pp. 67–112, AGU, Washington, D. C.

Behrendt, J. C., D. D. Blankenship, D. L. Morse, and R. E. Bell (2004), Shallow-source aeromagnetic anomalies observed over the West Antarctic Ice Sheet compared with coincident bed topography from radar ice sounding—New evidence for glacial "removal" of subglacially erupted late Cenozoic rift-related volcanic edifices, *Global Planet. Change*, *42*, 177–193.

Bell, R. E., V. A. Childers, R. A. Arko, D. D. Blankenship, and J. M. Brozena (1999), Airborne gravity and precise positioning for geological applications, *J. Geophys. Res.*, *104*(B7), 15,281–15,292.

Cande, S. C., and J. M. Stock (2004), Constraints on Late Cretaceous and Cenozoic extension in the Ross Sea from the southwest Pacific Plate Circuit, *Eos Trans. AGU*, *85*(47), Fall Meet. Suppl., Abstract T14A-03.

Cande, S. C., J. M. Stock, D. Müller, and T. Ishihara (2000), Cenozoic Motion between East and West Antarctica, *Nature*, *404*, 145–150.

Conway, H., B. L. Hall, G. H. Denton, A. M. Gades, and E. D. Waddington (1999), Past and future grounding-line retreat of the West Antarctic Ice Sheet, *Science*, *286*, 280–283.

Davey, F. J. (2004), Ross Sea Bathymetry, 1:2,000,000, version 1.0, Inst. of Geol. and Nucl. Sci. Ltd., Lower Hutt, New Zealand.

Davey, F. J., and G. Brancolini (1995), The Late Mesozoic and Cenozoic structural setting of the Ross Sea region, in *Geology and Seismic Stratigraphy of the Antarctic Margin*, *Antarct. Res. Ser.*, vol. 68, edited by A. K. Cooper, P. F. Barker, and G. Brancolini, pp. 167–182, AGU, Washington, D. C.

DeConto, R. M., and D. Pollard (2003), A coupled climate-ice sheet modeling approach to the Early Cenozoic history of the Antarctic ice sheet, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *198*, 39–52.

Denton, G. H., M. L. Prentice, and L. H. Burckle (1991), Cainozoic history of the Antarctic ice sheet, in *Geology of Antarctica*, edited by R. J. Tingey, pp. 365–433, Oxford Univ. Press, New York.

DeSantis, L., S. Prato, G. Brancolini, M. Lovo, and L. Torelli (1999), The Eastern Ross Sea continental shelf during the Cenozoic: Implications for the West Antarctic ice sheet development, *Global Planet. Change*, *23*, 173–196.

Embleton, C., and C. A. M. King (1975), *Glacial Geomorphology*, 2nd ed., 573 pp., Edward Arnold, London.

Faleide, J. I., A. Solheim, A. Fiedler, B. O. Hjelstuen, E. S. Andersen, and K. Vanneste (1996), Late Cenozoic evolution of the western Barents Sea-Svalbard continental margin, *Global Planet. Change*, *12*, 53–74.

Gades, A. M., C. F. Raymond, H. Conway, and R. W. Jacobel (2000), Bed properties of Siple Dome and adjacent ice streams, West Antarctica, inferred from radio-echo sounding measurements, *J. Glaciol.*, *46*(152), 88–94.

Gale, A. S., J. Hardenbol, B. Hathway, W. J. Kennedy, J. R. Young, and V. Phansalkar (2002), Global correlation of Cenomanian (Upper Cretaceous) sequences: Evidence for Milankovitch control on sea level, *Geology*, *30*, 291–294.

Greischar, L. L., C. R. Bentley, and L. R. Whiting (1992), An analysis of gravity measurements on the Ross Ice Shelf, Antarctica, in *Contributions to Antarctic Research III*, *Antarct. Res. Ser.*, vol. 57, edited by D. H. Elliot, pp. 105–155, AGU, Washington, D. C.

Harwood, D. M., R. P. Scherer, and P.-N. Webb (1989), Multiple Miocene marine productivity events in West Antarctica as recorded in upper Miocene sediments beneath the Ross Ice Shelf (Site J9), *Mar. Micropaleontol.*, *15*, 91–115.

Hayes, D. E., and L. A. Frakes (1975), General synthesis: Deep Sea Drilling Project 28, *Initial Rep. Deep Sea Drill. Proj.*, *28*, 919–942.

Hinz, K., and M. Block (1983), Results of geophysical investigations in the Weddell Sea and in the Ross Sea, Antarctica, paper presented at 11th World Petroleum Congress, Wiley, London.

Jacobel, R. W., A. E. Robinson, and R. A. Binschadler (1994), Studies of the grounding-line location on Ice Streams D and E, Antarctica, *Ann. Glaciol.*, *20*, 39–42.

Jakobsson, M., N. Z. Cherkis, J. Woodward, R. Macnab, and B. Coakley (2000), New grid of Arctic bathymetry aids scientists and mapmakers, *Eos Trans. AGU*, *81*(9), 89, 93, 96.

James, T. S., and E. R. Ivins (1998), Predictions of Antarctic crustal motions driven by present-day ice sheet evolution and by isostatic memory of the last glacial maximum, *J. Geophys. Res.*, *103*(B3), 4933–5017.

John, B. S., and D. E. Sugden (1975), Coastal geomorphology of high latitudes, *Prog. Geogr.*, *7*, 53–112.

Joughin, I., L. Gray, R. Binschadler, S. Price, D. L. Morse, C. Hulbe, K. Mattar, and C. Werner (1998), Tributaries of West Antarctic ice streams revealed by RADARSAT interferometry, *Science*, *286*, 283–286.

Kennett, J. P., and P. F. Barker (1990), Latest Cretaceous to Cenozoic climate and oceanographic developments in the Weddell Sea, Antarctica: An ocean-drilling perspective, *Proc. Ocean Drill. Program Sci. Results*, *113*, 937–960.

Klemsdal, T. (1982), Coastal classification and the coast of Norway, *Norsk Geogr. Tidsskr.*, *36*, 129–152.

LeMasurier, W. E., and C. A. Landis (1996), Mantle plume activity recorded by low relief erosion surfaces in West Antarctica and New Zealand, *Geol. Soc. Am. Bull.*, *108*, 1450–1466.

Liu, H., K. C. Jezek, and B. Li (1999), Development of an Antarctic digital elevation model by integrating cartographic and remotely sensed data: A geographic information system based approach, *J. Geophys. Res.*, *104*(B10), 23,199–23,213.

Luyendyk, B. P., C. C. Sorlien, D. S. Wilson, L. R. Bartek, and C. H. Siddoway (2001), Structural and tectonic evolution of the Ross Sea rift in the Cape Colbeck region, Eastern Ross Sea, Antarctica, *Tectonics*, *20*, 933–958.

Luyendyk, B. P., D. S. Wilson, and R. Decesari (2002), New maps of gravity and bedrock-bathymetry of the Ross Sea sector of Antarctica, *Eos Trans. AGU*, *83*(47), Fall Meet. Suppl., Abstract T12D-1339.

Luyendyk, B. P., D. S. Wilson, and C. S. Siddoway (2003), Eastern margin of the Ross Sea Rift in western Marie Byrd Land, Antarctica: Crustal structure and tectonic development, *Geochem. Geophys. Geosyst.*, *4*(10), 1090, doi:10.1029/2002GC000462.

Lythe, M. B., et al. (2001), BEDMAP: A new ice thickness and subglacial topographic model of Antarctica, *J. Geophys. Res.*, *106*(6), 11,335–11,351.

McKenzie, D. (1978), Some remarks on the development of sedimentary basins, *Earth Planet. Sci. Lett.*, *40*(7), 25–32.

Miller, K. G., J. D. Wright, and R. G. Faribanks (1991), Unlocking the ice house: Oligocene-Miocene oxygen isotopes, eustasy, and margin erosion, *J. Geophys. Res.*, *96*(B4), 6829–6848.

Miller, K. G., P. J. Sugarman, J. V. Browning, M. A. Kominz, R. K. Olsson, M. D. Feigenson, and J. C. Hernández (2004), Upper Cretaceous sequences and sea-level history, New Jersey Coastal Plain, *Geol. Soc. Am. Bull.*, *116*, 368–393.

Miller, K. G., M. A. Kominz, J. V. Browning, J. D. Wright, G. S. Mountain, M. E. Katz, P. J. Sugarman, B. S. Cramer, N. Christie-Blick, and S. F. Pekar (2005), The Phanerozoic record of global sea-level change, *Science*, *310*(25), 1293–1298.

Nansen, F. (1922), *The Strandflat and Isostasy*, 313 pp., I Komm. hos Jacob Dybwad, Kristiania, Oslo.

Nerenson, N. A., C. F. Raymond, R. W. Jacobel, and E. D. Waddington (2000), The accumulation pattern across Siple Dome, West Antarctica, inferred from radar-detected internal layers, *J. Glaciol.*, *46*, 75–87.

Parizek, B. R., and R. B. Alley (2004), Ice-thickness and isostatic imbalances in the Ross Embayment, West Antarctica: Model results, *Global Planet. Change*, *42*, 265–278.

Parsons, B., and J. G. Sclater (1977), An analysis of the variation of ocean floor bathymetry and heat flow with age, *J. Geophys. Res.*, *78*, 5128–5137.

Rasmussen, E., and W. Fjeldskaar (1996), Quantification of the Pliocene-Pleistocene erosion of the Barents Sea from present-day bathymetry, *Global Planet. Change*, *12*(1–4), 119–133.

Retallack, G. J. (2002), Carbon dioxide and climate over the past 300 Myr, *Philos. Trans. R. Soc. London, Ser. A*, *360*, 659–673.

Rose, K. E. (1979), Characteristics of ice flow in Marie Byrd Land, Antarctica, *J. Glaciol.*, *24*(90), 63–75.

Shevenell, A. E., J. P. Kennett, and D. W. Lea (2004), Middle Miocene Southern Ocean cooling and Antarctic cryosphere expansion, *Science*, *305*, 1766–1770.

Siegert, M. J., and J. A. Dowdeswell (1996), Topographic control on the dynamics of the Svalbard-Barents Sea Ice Sheet, *Global Planet. Change*, *12*, 27–39.

Solheim, A., F. Riis, A. Elverhøi, J. I. Faleide, L. N. Jensen, and S. Cloetingh (1996), Impact of glaciations on basin evolution: Data and models from the Norwegian margin and adjacent areas—Introduction and summary, *Global Planet. Change*, *12*(1–4), 1–9.

Steig, E. J., J. L. Fastook, C. Zweck, I. D. Goodwin, K. J. Licht, J. W. C. White, and J. R. P. Ackert (2001), West Antarctic Ice Sheet elevation changes, in *The West Antarctic Ice Sheet: Behavior and Environment*, *Antarct. Res. Ser.*, vol. 77, edited by R. E. Alley and R. Bindschadler, pp. 75–90, AGU, Washington D. C.

Stern, T. A., and A. K. Baxter (2002), Glacial erosion, rock, and peak uplift within the Central Transantarctic Mountains, in *Antarctica at the Close of a Millennium: Proceedings of the 8th International Symposium on Antarctic Earth Sciences, Wellington, 1999*, edited by J. A. Gamble et al., pp. 471–478, R. Soc. of N. Z., Wellington, New Zealand.

Studinger, M., R. E. Bell, W. R. Buck, G. Karner, and D. Blankenship (2004), Sub-ice geology inland of the Transantarctic Mountains in light of new aerogeophysical data, *Earth Planet. Sci. Lett.*, *220*, 391–408.

Sugden, D. E., and B. S. John (1976), *Glaciers and Landscape: A Geomorphological Approach*, 310 pp., John Wiley, Hoboken, N. J.

Taylor, K. C., et al. (2004), Abrupt climate change around 22 ka on the Siple Coast of Antarctica, *Quat. Sci. Rev.*, *23*, 7–15.

Tessensohn, F., and G. Worner (1991), The Ross Sea rift system, Antarctica: Structure, evolution, and analogs, in *Geological Evolution of Antarctica*, edited by M. R. A. Thompson et al., pp. 273–278, Cambridge Univ. Press, Cambridge, Mass.

van der Linden, W. J., and S. P. Srivastava (1975), The crustal structure of the continental margin off central Labrador, in *Offshore Geology of Western Canada*, *Pap. 74–30*, edited by W. J. van der Linden and J. A. Wade, pp. 233–245, Geol. Surv. of Can., Ottawa, Ontario, Canada.

van der Linden, W. J., R. H. Fillon, and D. Monahan (1976), Hamilton Bank, Labrador Margin, *Pap. 75–40*, 31 pp., Geol. Surv. of Can., Ottawa, Ontario, Canada.

Waddington, E. D., H. Conway, E. J. Steig, R. B. Alley, E. J. Brook, K. C. Taylor, and J. W. C. White (2005), Decoding the dipstick: Thickness of Siple Dome, West Antarctica, at the Last Glacial Maximum, *Geology*, *33*(4), 281–284, doi:210.1130/G21165.21161.

Worsley, D., T. Agdestein, J. G. Gjelberg, K. Kirkemo, A. Mork, I. Nilsson, S. Olaussen, R. J. Steel, and L. Stemmerik (2001), The geological evolution of Bjornoya, Arctic Norway: Implications for the Barents Shelf, *Norsk Geol. Tidsskr.*, *81*(3), 195–234.

Zachos, J., M. Pagani, L. Sloan, E. Thomas, and K. Billups (2001), Trends, rhythms, and aberrations in global climate 65 Ma to present, *Science*, *292*, 686–693.