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The response of the East Antarctic ice-sheet to the evolving tectonic configuration of the Transantarctic Mountains

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Abstract

The landscape of the Transantarctic Mountains is the result of the coupled evolution of the West Antarctic rift system and the East Antarctic ice-sheet. Studies of this glacial–tectonic system generally assume that the evolving surface elevation of the Transantarctic Mountains is a key determinant of the changing East Antarctic ice-sheet dynamics between the Miocene and today. Here, we extend previous work [Huybrechts, Ph., 1993. Glaciological modelling of the Late Cenozoic East Antarctic ice-sheet: stability or dynamism? Geografiska Annaler Stockholm, 75A (4) 221–238.] by using numerical models of the ice-sheet and lithosphere to examine the impact of different bedrock surface elevations of the Transantarctic Mountains on ice-sheet dynamics. There are widely different interpretations of the evolution of the Transantarctic Mountains from the available data, so we explore bedrock surface elevations suggested by empirical evidence in recent papers about the sensitivity of the Late Cenozoic ice-sheet.

The results show that the surface elevation of individual mountain blocks has only a very local effect on ice-sheet dynamics. The existing mountain blocks of the Transantarctic Mountains, which force inland ice to drain through troughs adjacent to the mountain blocks, were overriden by inland ice when bedrock elevations were 1 km lower. When the troughs through the mountains were less well developed, in the Pliocene or Miocene, inland ice was thicker and ice-surface gradients and ice-velocities across the mountains were higher. This led to more active and erosive outlet glaciers through the mountains and the further development of these troughs. From these results, the key determinant of East Antarctic ice dynamics appears to be the interplay between the development of major troughs through the Transantarctic Mountains and rising mountain elevations.

The glacial history of the central Transantarctic Mountain ranges was very different to that of more peripheral mountain ranges, such as the Dry Valleys and Victoria Land. The development of independent ice centres in the latter regions and the overriding of these ice centres by the main ice-sheet is very sensitive to the timing of surface uplift and the particular climate profile of the period. Conversely, the ice-surface profile across the central ranges is similar under widely different climates.

The limitations of such a study stem from the necessarily schematic bedrock elevations input to the model and simplifications within the models. At present, insufficiently detailed modelling of the impact of troughs on ice-sheet dynamics means this paper is necessarily speculative. However, this work points to the importance of the outlet troughs on

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ice-sheet dynamics, rather than simply the rising surface elevations of the Transantarctic Mountains along the rift margin upwarp.

Keywords: Antarctica; Cenozoic; ice-sheet; glacial; tectonic; paleotopography

1. Introduction

The coupled evolution of the rift flank of the West Antarctic rift system and the East Antarctic ice-sheet has resulted in the characteristic landscape of the Transantarctic Mountains, bordering the Ross Sea Basin. The links between the tectonics, ice and landscape of this region have been examined intensively in recent years (Stern and ten Brink, 1989; Behrendt and Cooper, 1991; Denton et al., 1993; Webb, 1994; Sugden et al., 1995). Much of this work implicitly assumes that the evolving surface elevation of the Transantarctic Mountains is a key determinant of East Antarctic ice-sheet dynamics. For example, Webb et al. (1984) suggested that the rising mountain elevations blocked the flow of the ice-sheet and, thus, affected ice-sheet dynamics. Behrendt and Cooper (1991) asserted that the rising mountain elevations could have contributed to the change in basal thermal character of the ice-sheet from temperate (basal melting) to polar (frozen base), and hence a change from a dynamic to a more stable

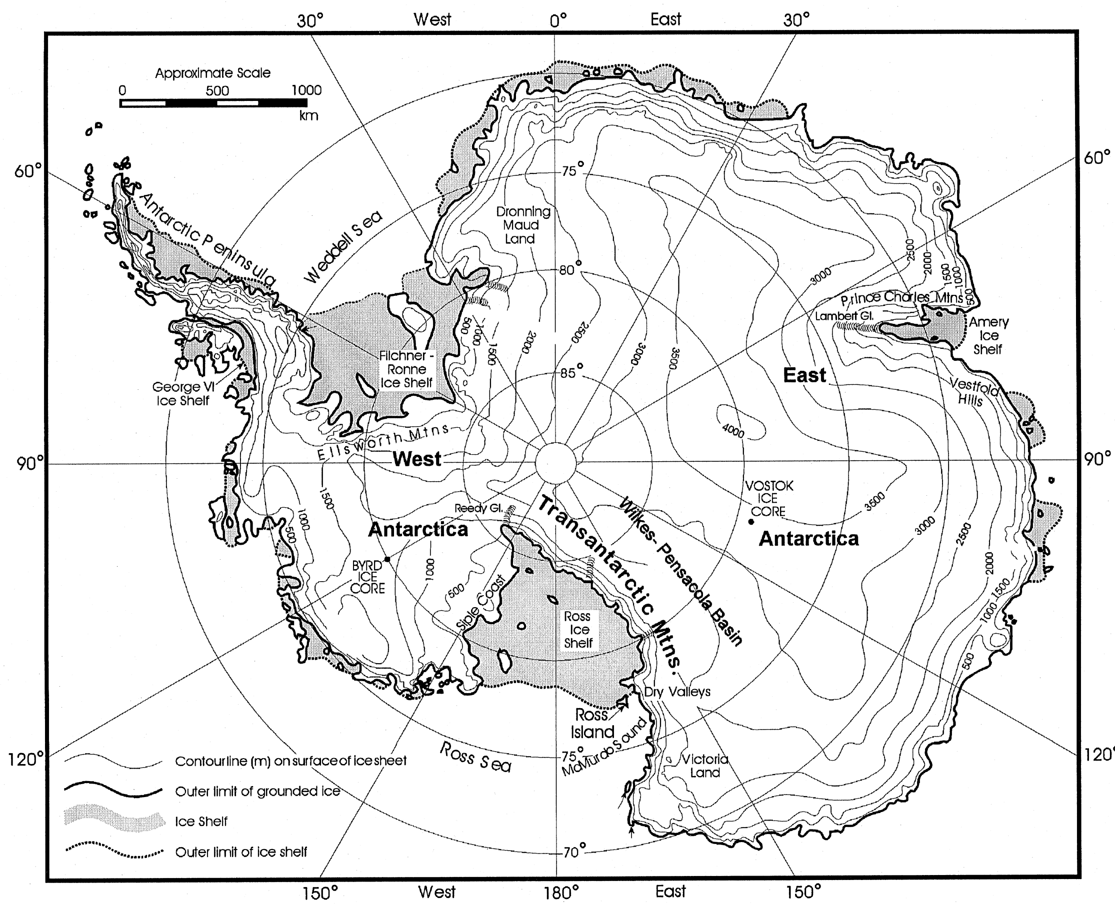

Fig. 1. The present ice surface of Antarctica, showing the major regions mentioned in the paper, including Victoria Land, the Dry Valleys, and the Wilkes–Pensacola Basin. The central ranges of the Transantarctic Mountains, comprising the Queen Alexandria, Queen Elizabeth and Queen Maud Ranges, are situated under the 'Transantarctic' of the Transantarctic Mountains caption.

ice sheet. Fig. 1 shows the present East Antarctic ice-sheet.

The Transantarctic Mountains sweep in an arc across Antarctica from the Pacific coast to the junction of the East and West Antarctic ice-sheets, and continue as a coherent topographic feature as far as the South Atlantic Ocean (Fig. 2). They consist of a large-amplitude (~ 2000–4000 m), short-wavelength (50–200 km) upwarp, that forms the high rim of an extensive plateau that rises gradually from the interior of East Antarctica. The outer flank of this plateau is characterised by a major escarpment, which marks the hinge between the East Antarctic craton and rifted continental lithosphere underlying the Ross Sea.

Identifying the history of changes in surface elevation of the Transantarctic Mountains remains elusive. The available geophysical, geochemical and geochronological data provide us with an understanding of the tectonic structure and nature of the Transantarctic Mountains and their hinterland (Behrendt et al., 1991; Kerr et al., 1999), but do not

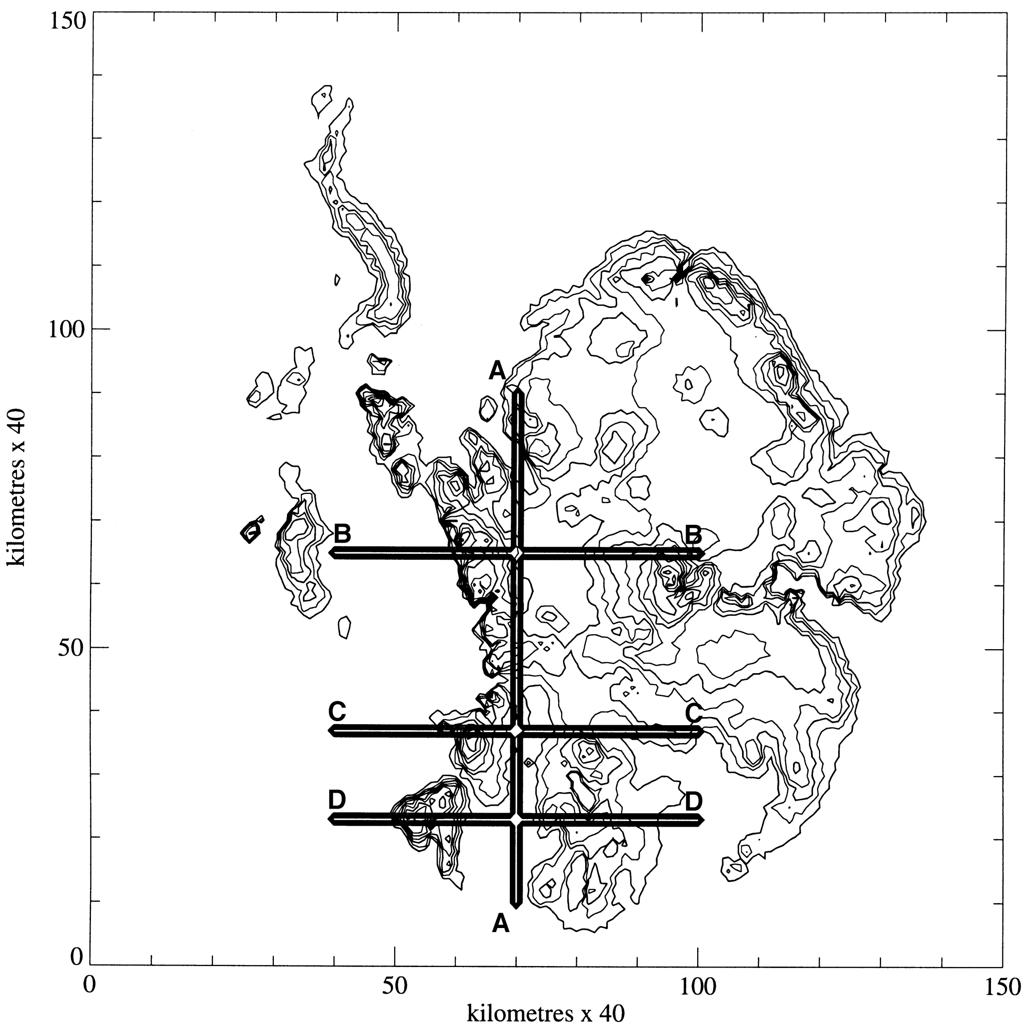

Fig. 2. The Antarctic bedrock, showing the position of the bedrock that was raised or lowered in the experiments and the transects shown in Figs. 3–10. Contours are every 500 m and the bold lines are every 1000 m.

provide an unambiguous history of the surface configuration of this escarpment. Apatite fission track methods have been used to provide an estimate of the spatial variation of denudation across the tectonic margin (Fitzgerald, 1992, 1994). Models that use these data provide theoretical estimates of possible surface uplift, which suggest substantial surface uplift, of the order of 1–2 km, is possible over the period of the Cenozoic (ten Brink and Stern, 1992; van der Beek et al., 1994; Kerr and Gilchrist, 1996). However, the only absolute constraints come from volcanic rocks, such as those in the Dry Valleys dated from 3.89 to 1.50 Ma, which constrain net surface uplift in this area to less than 300 m (Wilch et al., 1993). Similarly, Brook et al. (1995), in the same area, published cosmogenic surface exposure data, which suggested that erosion rates were very low, of the order of 0.06–0.27 m Ma^{-1}, and that rock surfaces were situated close to sea level implying small net surface uplift. Since there are few representative data for the entire continental margin, and differential movement of tectonic blocks is apparent, there have been widely differing interpretations of the evolution of the Transantarctic Mountains.

In this paper we use a numerical model of the ice-sheet and lithosphere to examine the impact of different Transantarctic Mountain bedrock elevations on ice-sheet dynamics. We extend previous work, which examined the impact on ice dynamics of arbitrarily lowered bed-topography in East Antarctica (Huybrechts, 1993), and explore whether the surface elevation of the Transantarctic Mountains is, in fact, a key determinant of ice-sheet dynamics. Since empirical evidence of the tectonic and climatic history of this region remains ambiguous, our approach is to explore the response of the East Antarctic ice-sheet to different bedrock elevations of the Transantarctic Mountains suggested in recent papers (Behrendt and Cooper, 1991; van der Wateren et al., 1994; Webb, 1994; Sugden et al., 1995).

2. The model

The ice-sheet model was developed by Huybrechts (1990) and adapted for the problem of pre-Quaternary East Antarctic ice-sheet dynamics (Huybrechts, 1993). In these experiments only grounded ice flow is taken into account. The model

Table 1
A list of the model parameters and units

Equation parameter	Description	Units
H	Ice thickness	m
v	Depth-averaged horizontal velocity	yr^{-1}
M	Surface mass balance	yr^{-1}
t	time	years
T	Ice temperature	K
k	Temp-dependent ice thermal conductivity	$J\ m^{-1}\ K^{-1}\ yr^{-1}$
ρ_i	Ice density	$910\ kg\ m^{-3}$
c_p	Temp-dependent specific heat capacity	$J\ kg^{-1}\ K^{-1}$
V	3-dimensional velocity vector	$m\ yr^{-1}$
Φ	Internal frictional heating	$J\ m^{-3}\ yr^{-1}$
ν	Poisson’s ratio	0.25
T_e	Effective elastic thickness	60 km
ρ_m	Mantle density	$3300\ kg\ m^{-3}$
$L(x)$	Ice load	kg
E	Young’s modulus	100×10^9
g	gravity	$9.81\ m\ s^{-2}$
x	distance	m
w	Lithospheric deflection	m
D	Flexural rigidity	N m

treats ice-flow by solving the full time-dependent set of thermo-mechanical equations in three dimensions, with the only restriction that the horizontal velocity vector cannot change direction with depth. The following ice flow assumptions apply: ice deformation is assumed to result from shearing in horizontal planes and longitudinal deviatoric stresses are disregarded (Paterson, 1993). This approximation implies that longitudinal strain-rate components are negligible. This assumption breaks down at the margin and at the centre of an ice-sheet, but does not significantly alter the profile of ice-sheets larger than about 30 km (Weertman, 1961). To close the set of equations specifying ice deformation, the temperature distribution needs to be known simultaneously since the material properties of ice are nonlinear with respect to temperature. In addition to deformation, ice moves by 'basal sliding', which is delimited to areas where the calculated temperature indicates basal melting of ice. This is modelled to depend on the shear stress at the bed, on bed roughness, and on the 'effective normal load', which generates a frictional force resisting the sliding motion.

The basic conservation equations for mass and heat solved by the model are:

$$\frac{\partial H}{\partial t} = -\nabla(\bar{v}H) + M \qquad (1)$$

$$\frac{\partial T}{\partial t} = \frac{k}{\rho c_p}\nabla^2 T - \bar{V}\nabla T + \frac{\Phi}{\rho c_p} \qquad (2)$$

The solution of Eq. 1 follows from reformulating the velocity field in terms of the ice sheet geometry (ice thickness and surface gradient) so only one unknown remains. The resulting equation is then a non-linear differential equation and has to be solved by numerical methods. H is ice thickness, v the depth-averaged horizontal velocity field (including deformation and sliding), M the mass balance, t is time, and the ∇-operator is two-dimensional in the xy-plane. Table 1 states the parameter values and units used in the experiments. Eq. 2 represents respectively vertical heat conduction, three-dimensional advection of heat, and heat generation by internal deformation. Here T is temperature, ρ is ice density, V is three-dimensional ice velocity, Φ is internal frictional heating caused by ice-deformation, and k and c_p are the temperature-dependent thermal conductivity and specific heat capacity respectively.

The equations are solved on a numerical grid using the Alternating-Direction-Implicit finite difference method, with a grid point spacing of 40 km. There are 11 layers in the vertical in a stretched co-ordinate system. This means that the vertical grid size is scaled to specific vertical intervals relative to the local ice thickness (Jenssen, 1977). Iterations of the model represent 10 model-years and the model is run for 100,000 model-years to approach an equilibrium solution.

The lithosphere model determines the response of the bedrock to ice loading and unloading. This comprises two parts: the time-dependent response and the final steady-state deflection from the original state. The time-dependent bedrock response is modelled as a damped return to its equilibrium state with a characteristic time-scale of 3000 years. Le Meur and Huybrechts (1996) tested different models and found that, apart from using a full self-gravitating visco-elastic spherical Earth model, this model best re-produced available Antarctic data. The steady-state deflection of the bedrock depends on the lithospheric flexure to loading, which can be modelled as representing the loading on a thin elastic plate of constant thickness overlying an inviscid fluid according to:

$$\frac{\partial^2}{\partial x^2}\left(D\frac{\partial^2 w}{\partial x^2}\right) + (\rho_m - \rho_i)gw = l(x) \qquad (3)$$

Here, w is the vertical flexure, D is the flexural rigidity, ρ_m is the mantle density, ρ_i is the flexure infill density, g is the acceleration due to gravity, and $l(x)$ is the ice load as a function of x (Nadai, 1963). The flexural rigidity, D, is given by:

$$D = \frac{\mathrm{ET}_e^3}{12(1-\nu^2)} \qquad (4)$$

E is Young's modulus, T_e is the effective elastic thickness of the lithosphere and ν is Poisson's ratio (Table 1). The lithosphere model is coupled to the ice-sheet model by re-calculating the bedrock response to the changing ice cover every 250 model-years.

3. Model inputs

The model inputs required for the experiments are the bed topography, the climate, which provides the mass balance of ice in each grid-cell, and a temperature gradient at the lower surface incorporating the effects of geothermal heating and heat dissipation by sliding.

The bedrock data was obtained from an updated and digitised version of the Drewry (1983) map folio. To obtain an undisturbed bedrock elevation with which to start the model experiments, the present ice-sheet was unloaded from the present deflected bedrock using an effective lithospheric elastic thickness of 60 km. This represents a flexurally rigid lithosphere, reflecting the old cratonic lithosphere underlying the East Antarctic ice-sheet. The rifted lithosphere of the West Antarctic rift system underlying the Ross Sea is likely to have a flexural rigidity lower than the value we use, but experiments show that the far-field effect of this rifted lithosphere is negligible.

The model freely evolves to an excellent representation of the present Antarctic ice-sheet when using the ice-sheet model parameters developed by Huybrechts (1990, 1993) and forced with the present climatic regime. Ice mass balance in each grid-cell is determined by the present climate, which is strongly dependent on the mean annual air temperature at the upper surface. Changes in climate are modelled by perturbing the present climatic profile over the ice sheet with different mean annual temperatures. This approach to climate change assumes that the present mass balance regime remains the same in different climates. This assumption appears reasonable for the period that Antarctica has remained geographically and climatically isolated by the Southern Ocean.

At its base, the ice sheet gains heat from both sliding friction and the geothermal heat flux, which are expressed as a geothermal gradient. A geothermal heat flux of 54.6 mW m^{-2} is used, which is representative of old Precambrian shields and is considered as an average value for the entire Antarctic continent (Sclater et al., 1980). Higher geothermal heat fluxes are to be expected in rifted continental regions, such as under the Ross Sea, but this does not greatly affect the area of grounded ice in East Antarctica. Increasing the geothermal heat flux increases basal melting, and hence basal sliding, but the value of the geothermal heat flux is not critical to the sensitivity of the modelled ice-sheet thickness (Huybrechts, 1990).

4. Limitations of the models

The limitations of the ice-sheet and lithosphere models reflect the assumptions used to simplify the thermo-mechanical equations representing ice and the lithosphere. The key limitations are:

(1) The Antarctic bedrock is represented on a grid by cells with dimensions of 40 km. As a result, regions of high relief are poorly represented in the model and only the larger troughs through the Transantarctic Mountains are resolved.
(2) Ice streams are only crudely represented and ice-shelf physics is not included. Instead, ice is treated as grounded ice until it reaches the grounding line, where it is lost to the ocean.
(3) Perturbing the present mass-balance field using the mean annual temperature to explore changes in climate implicitly assumes that the climatic regime of Antarctica will remain effectively unchanged through time.
(4) The lithospheric model uses a single effective elastic thickness to characterise the response of the lithosphere to loading and a single parameter to reflect the time-scale of its time-dependent response.

5. Linking tectonic and glacial models

Large-scale tectonic processes typically operate on time-scales of millions of years, while large ice-sheets have response times to forcing that are an order of magnitude smaller. We can therefore reasonably assume that the ice-sheet is in quasi-equilibrium with the tectonic processes. This indicates that we can explore the glacial–tectonic coupling by means of a series of snap-shots of ice-sheet dynamics, at specific points in the evolution of the Transantarctic Mountains. In reality, the process of glacial erosion provides a link between ice-sheet dynamics and regional tectonics since the erosion, transport, and deposition of material by glaciers reorganises lithospheric loading, and hence the configuration of the bedrock.

6. Experiments: rationale and design

Recent debate about the dynamics of the Late Cenozoic East Antarctic ice sheet has focused attention on the timing and mechanisms of surface uplift of the rift upwarp comprising the Transantarctic Mountains. Behrendt and Cooper (1991) and Webb (1994) argued that substantial surface uplift, of the order of 1 km, has taken place in the last 1–2 Ma while others such as Wilson (1995) and Sugden et al (1999) point to evidence of surface uplift prior to the Miocene. Tectonic modelling of the West Antarctic rift system has cited numerous mechanisms that could contribute to the formation of the upwarped margin. Recent work suggests that the flexural uplift results from one or all of a combination of lithospheric necking (van der Beek et al, 1994), thermal buoyancy (ten Brink et al., 1993) and differential denudation across the margin (Kerr and Gilchrist, 1996). While the tenor of these papers does not support significant uplift of the rift upwarp in the Late Cenozoic, the data are insufficiently constrained to rule it out.

If there has been recent surface uplift of 1 km in the last 2 Ma, then climate forcing on the East Antarctic ice-sheet will be little changed from today. Thus we can explore ice-sheet dynamics by lowering surface elevations along the rift upwarp using the present climate profile. If the surface uplift occurred much earlier, the climate is likely to have been different, and we need to explore ice sheet dynamics under different climate profiles.

A complication to the experimental design is that empirical evidence suggests different blocks along the upwarped margin have experienced different surface uplift histories. In northern Victoria Land, van der Wateren et al. (1994) explain the landscape as the glacial dissection of the original plateau, which has been tectonically uplifted since the Pliocene. Conversely, further south along the rift upwarp in the Dry Valleys and Royal Society Range, the surface uplift is constrained by sub-aerially erupted volcanic rocks to a maximum of around 300 m since about 2.5 Ma BP (Wilch et al., 1993). In the central Transantarctic Mountains, Fitzgerald (1994) provides evidence of a complex thermal history and rapid denudation of up to 10 km near the coast. While this cannot constrain the timing of uplift, the evidence and high existing elevations of the mountains of 4000–4500 m suggests substantial uplift.

Geomorphic evidence indicates that many glaciers draping the upwarped margin are more protective than erosive of the landscape (Marchant et al., 1994). Meanwhile, Huybrechts (1990) showed that major outlet glaciers of the ice-sheet have basal ice temperatures at pressure–melting point, indicating their capability of eroding the glacier bed, and Powell et al. (1996) demonstrated the erosive capability of selected outlet glaciers under the present climate. This suggests that troughs are deepening relative to mountain summit elevations under the current climate. The actual depth attained by troughs depends on the flexural rigidity of the lithosphere, since the erosion of major troughs will unload the lithosphere, leading to the flexural compensation of surface uplift of the surrounding region (Wellman and Tingey, 1981).

In summary, the timing and extent of surface uplift of the mountains and the extent to which separate blocks of the upwarped margin have different surface elevation histories remains debated. Our modelling experiments are therefore constrained to exploring the impact on the East Antarctic ice-sheet of different bedrock elevations of the rift upwarp, suggested in the papers identified above. Since surface uplift may have occured when the climate profile was different from today, we must also explore the impact of different climate profiles on the ice-sheet. To this end, we identify climate profiles that reflect the likely range of ice-sheet volume and areal extent since the Miocene.

Inevitably, there is debate about the climate history of the region. Sugden et al. (1993) argue for the classical view, which states that the existing polar climate dates from the opening of the Drake Passage between South America and Antarctica during the Miocene. A second view holds that the ice sheet fluctuated extensively until the Pliocene ($\sim$ 3 Ma), since when a full polar climate became established (Webb et al., 1984). Since there is little direct evidence for the climate profile in Antarctica in the period back to the Miocene, we attempt to encompass the possible climate range by using three climate profiles ranging from the present-day to a profile represented by mean annual temperatures 10°C higher than today. This latter profile provides a means of exploring the dynamics of a smaller, more responsive East Antarctic ice-sheet. Higher mean annual temperatures than this appear unlikely since the mid-Miocene in view of the lack of evidence of surficial melt features around glaciers at low elevations in the Dry Valleys (Marchant et al., 1994). Furthermore, there is little evidence for a large change in the Antarctic oceanic convergence or the equator–pole temperature gradient over this period, which might be expected if there were very different climate profiles from today.

To explore the impact of different bedrock elevations on the East Antarctic ice-sheet, we split the Transantarctic Mountains upwarp into five elements based on suggestions by papers identified in the above discussion. These comprise three mountain blocks: northern Victoria Land, the Dry Valleys and Royal Society Range, and the central section around the Queen Alexandra, Queen Elizabeth and Queen Maud mountains. For the purposes of this exercise the lower altitude ranges adjacent to the Weddell Sea, which are not strictly part of the West Antarctic rift system, are assumed to be unchanged. The fourth element consists of five major outlet troughs, which divide the Transantarctic Mountains into discrete massifs, through which outlet glaciers drain the East Antarctic ice-sheet. Finally, two large sub-glacial troughs in the Wilkes–Pensacola Basin, in the hinterlands of and parallel to the upwarp, are identified. One drains ice into the Weddell Sea and the other drains ice towards northern Victoria Land.

The experiments consist of running the coupled ice-tectonic model to equilibrium using different

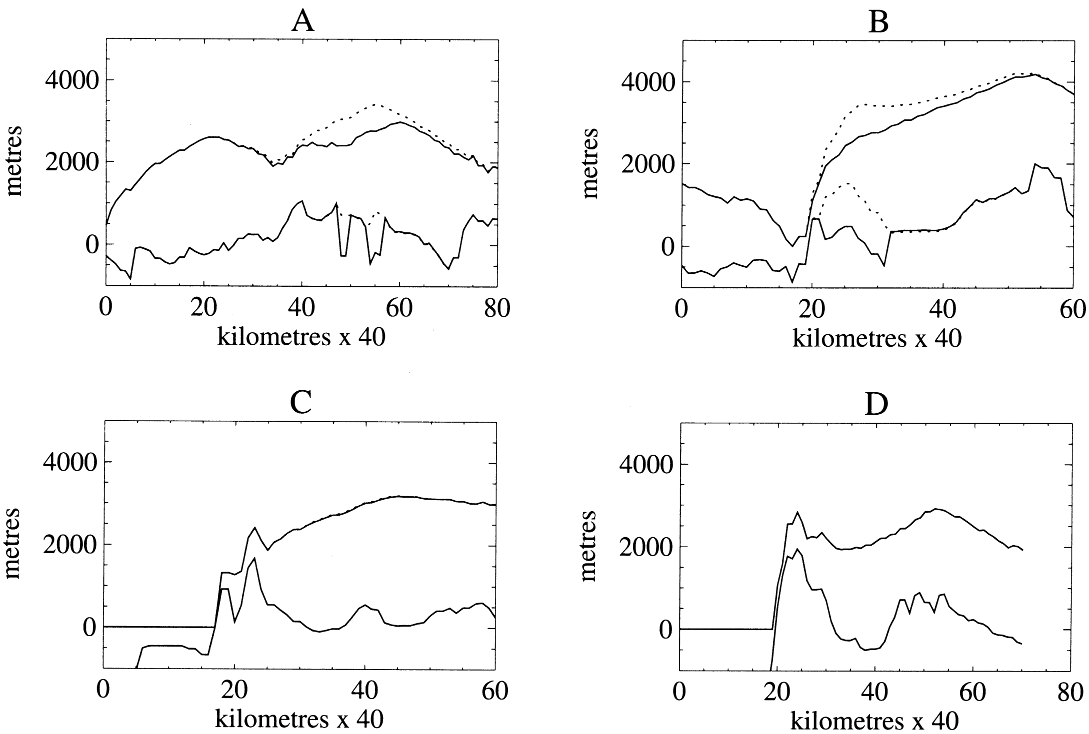

Fig. 3. The impact on ice-surface and bedrock elevations of lowering the central Transantarctic Mountain ranges by 1 km, using four transects cross-cutting the mountains and their hinterlands. (A) The profile along the Wilkes–Pensacola Basin showing the present-day ice-surface and bedrock (dotted lines) and the impact on the ice-surface and bedrock elevations resulting from lowering the bedrock of the central ranges by 1 km (continuous line). (B) As in (A), but the transect runs from the Ross Sea across the central ranges to the centre of the ice sheet. The lower dotted line shows clearly the effect of removing 1 km of bedrock from the rift margin upwarp. The upper two lines show the impact of the bedrock change on ice-surface elevations. (C) As in (A), but the transect runs from the Ross Sea across the Dry Valleys and Royal Society Range region and through the Wilkes–Pensacola Basin. (D) As (A), but the transect runs from the Ross Sea across Victoria Land and the Wilkes–Pensacola Basin.

bedrock configurations. We explore the following two key permutations of bedrock elevation and climate in this paper:

(1) If we assume that there was surface uplift of the order of 1 km since the Pliocene, we can use the present climate profile and examine the impact on the ice-sheet of lower-elevation mountain blocks in the Pliocene. Similarly, we can explore the response of the ice-sheet to shallower troughs (with respect to mountain summit elevations) through the Transantarctic Mountains and in the Wilkes–Pensacola Basin using the present climate profile.

(2) If the surface uplift is Miocene or older, we need to explore the impact on the ice-sheet of different climate profiles in addition to the changed bedrock elevations. We run the same experiments as before and compare the effect of the three different climate profiles.

For the former permutation, we show the impact on the ice-sheet as a series of four transects cutting through the Transantarctic Mountains and the adjacent Wilkes–Pensacola Basin. These show the bedrock and ice-surface elevations relative to a control run using the present-day bedrock elevation and climate. For the latter, we display one transect showing the impact of the different climate profiles relative to the present ice-surface for each different bedrock surface.

7. Results

7.1. Central ranges of the Transantarctic Mountains: Queen Alexandra, Queen Elizabeth and Queen Maud ranges

Fig. 3A–D shows the impact on the ice-sheet surface of lowering bedrock surface elevations in the central ranges of the Transantarctic Mountains by 1 km. The results are shown as three transects through the upwarped mountains (Fig. 3B–D) and one transect parallel to and inland from the upwarp, along the Wilkes–Pensacola Basin (Fig. 3A). The dotted lines mark the present bedrock and ice surface eleva-

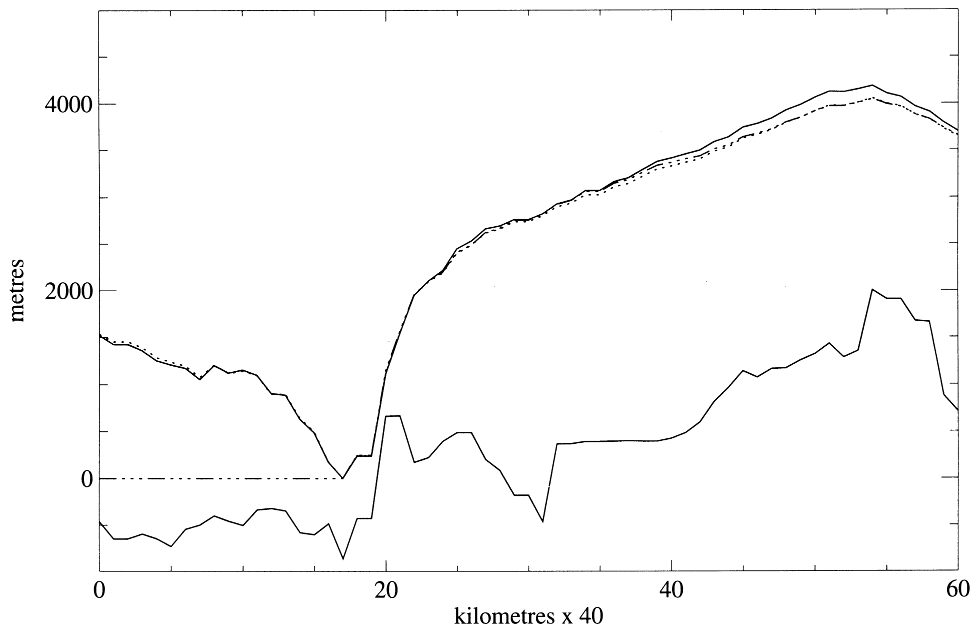

Fig. 4. The transect is the same as Fig. 3B, but this time showing the change in ice-surface elevation resulting from using different climate profiles across the central ranges: the present day (continous line), +5°C (dotted line), and +10°C (dash/dotted line). The impact of the changing climate on the ice-surface over the lowered central ranges of the Transantarctic Mountains, in the centre of the profile, is negligible.

tion, while the continuous lines represent the changed bedrock and ice elevations. It is apparent that changing the bedrock elevation of the central ranges by 1 km does not affect ice surface elevations in the region of the Dry Valleys and northern Victoria Land but that ice surface elevations across the central ranges are lowered by up to 750 m. This latter impact changes the ice-sheet profile, which currently has a steep margin rising from sea level to 3500 m over 300 km and then a very shallow gradient rising towards the centre of the ice-sheet at 4200 m nearly 1300 km inland. With lower bedrock, the ice profile rises to only 2750 m over the first 300 km and then rises progressively to an elevation of 4200 m at the centre. This change in surface elevation can also be seen on the transect along the Wilkes–Pensacola Basin, which crosses the former transect approximately 300 km inland. Ice surface elevations here are lowered by between 50 and 750 m along 800 km of the Basin.

Since ice velocity is driven by ice-surface gradient, these changes enhance ice-flow between the ice-sheet centre and the Wilkes–Pensacola Basin but reduces the net ice-flow across the Transantarctic Mountains under current climatic conditions. More ice overrides the mountain blocks but there is a smaller ice surface gradient and thus lower ice velocities. These lower ice velocities result in less basal ice melting, which further reduces ice-sliding velocities through the mountains.

The impact of different climate profiles is shown in Fig. 4. This displays the same transect as Fig. 3B shows the change in ice surface elevations for present, +5°C, and +10°C mean annual temperatures. The largest difference is at the centre of the ice-sheet, which is about 150 m lower for both warmer climate

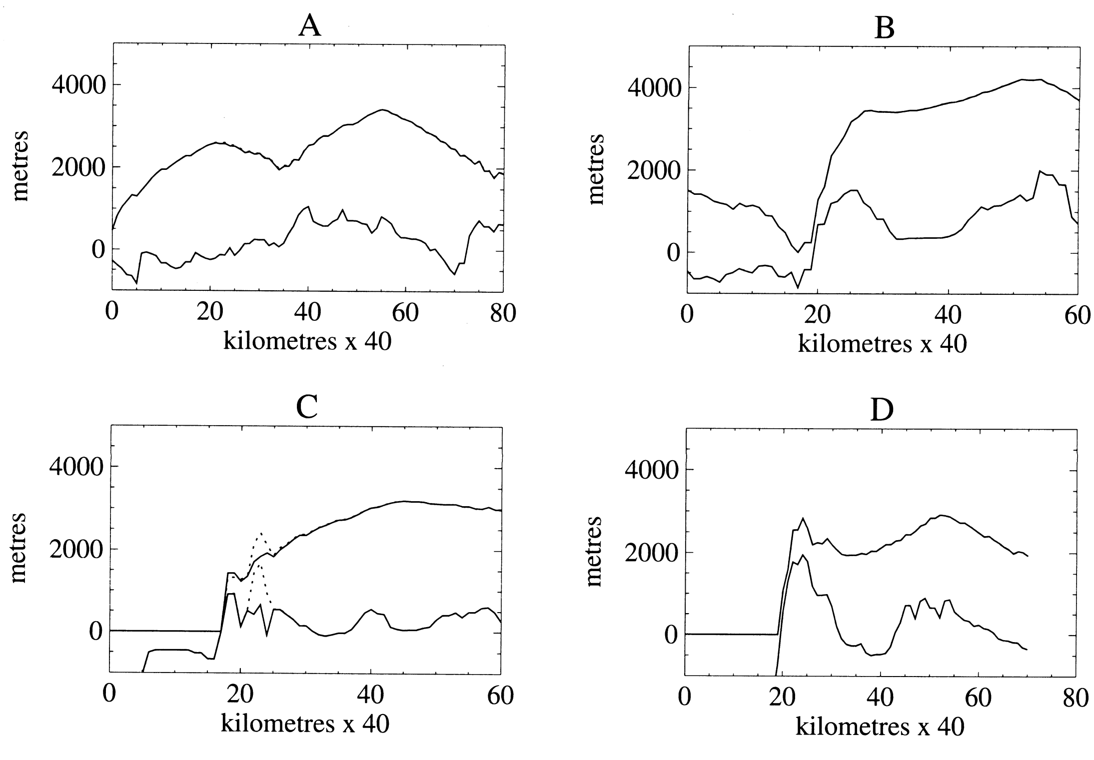

Fig. 5. The impact on the ice surface and bedrock elevations of lowering the Dry Valleys region by 1 km, shown using the same four transects used in Fig. 3. The removal of Dry Valleys bedrock is displayed in Fig. 5C, which also shows the resultant loss of the present ice centre that overlies the present region.

profiles. The impact on ice-surface elevations in the vicinity of the central ranges is negligible. This result suggests that ice dynamics in this mountain region are not critically dependent on the existence of the present climate profile. The left end of the profile shows the margin of the West Antarctic ice-sheet, which disappears under the warmest climate profile.

7.2. Dry valleys and the royal society range area

Fig. 5A–D shows the impact on the ice-surface of a lowered rift margin upwarp in the Dry Valleys and the Royal Society Range region. It is quickly apparent that lowering this section of the margin by 1 km has no effect on ice-surfaces over the ice-sheet apart from the loss of the local ice-centre immediately over these mountains, which lowers local ice-surfaces by 500 m. There is negligible change in the elevation of the ice in the adjacent Wilkes–Pensacola Basin.

The impact of different climate profiles can be seen in Fig. 6, which shows the bedrock and ice-surface profile across the Dry Valleys region. Raising mean annual temperatures by 5°C raises the surface elevation of the interior of the ice-sheet by 100 m but lowers it a little elsewhere. This result reflects the increased accumulation of snow, and hence mass-balance of ice, inland compared with increased melting in coastal regions. For a temperature increase of +10°C, the ice-sheet margin lies inland of the rift margin upwarp in the Dry Valleys area, which leaves a small independent ice-centre.

These results suggest that lower bedrock elevations in this area have only a local effect on ice dynamics, by allowing ice to overflow the rift margin upwarp more easily. Warmer climate profiles lead to the establishment of a small independent ice centre over the region, in isolation from the main ice sheet, which retreats to the Wilkes–Pensacola Basin.

7.3. Victoria land

Fig. 7A–D shows bedrock and ice-surface elevation profiles after the Victoria Land range is lowered by 1 km. The impact is similar to lowering the

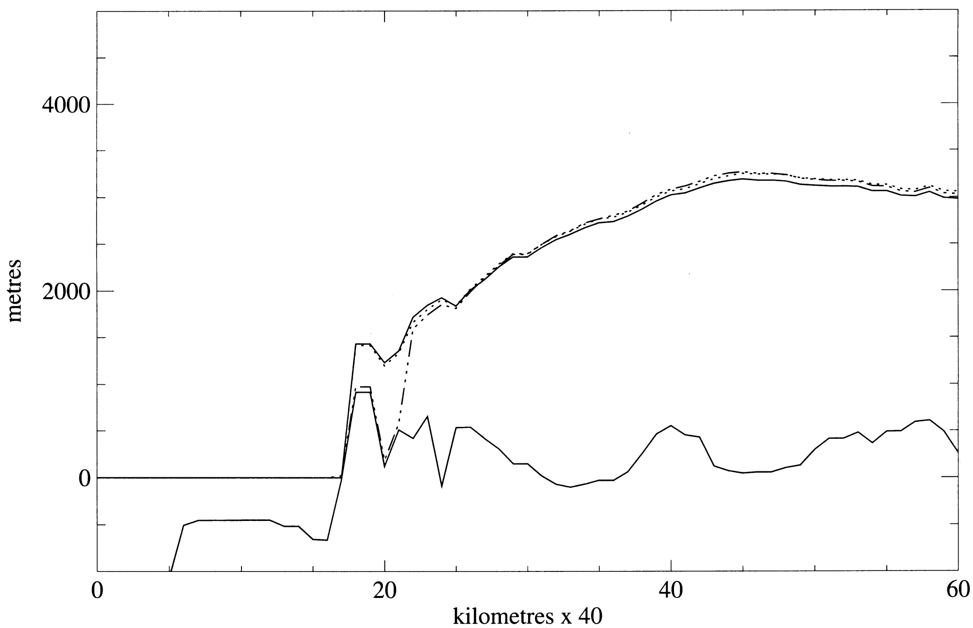

Fig. 6. As in Fig. 5C, but showing the change in ice-surface elevations across the Dry Valleys region resulting from applying different climate profiles. Again, present day ice elevations are displayed with a continous line, ice surfaces resulting from a +5°C temperature profile is shown with a dotted line, and the ice surface resulting from a +10°C temperature profile is shown with a dash/dot line. When the latter climate is applied, the ice-sheet margin retreats to the Wilkes–Pensacola Basin away from the lowered Dry Valleys block.

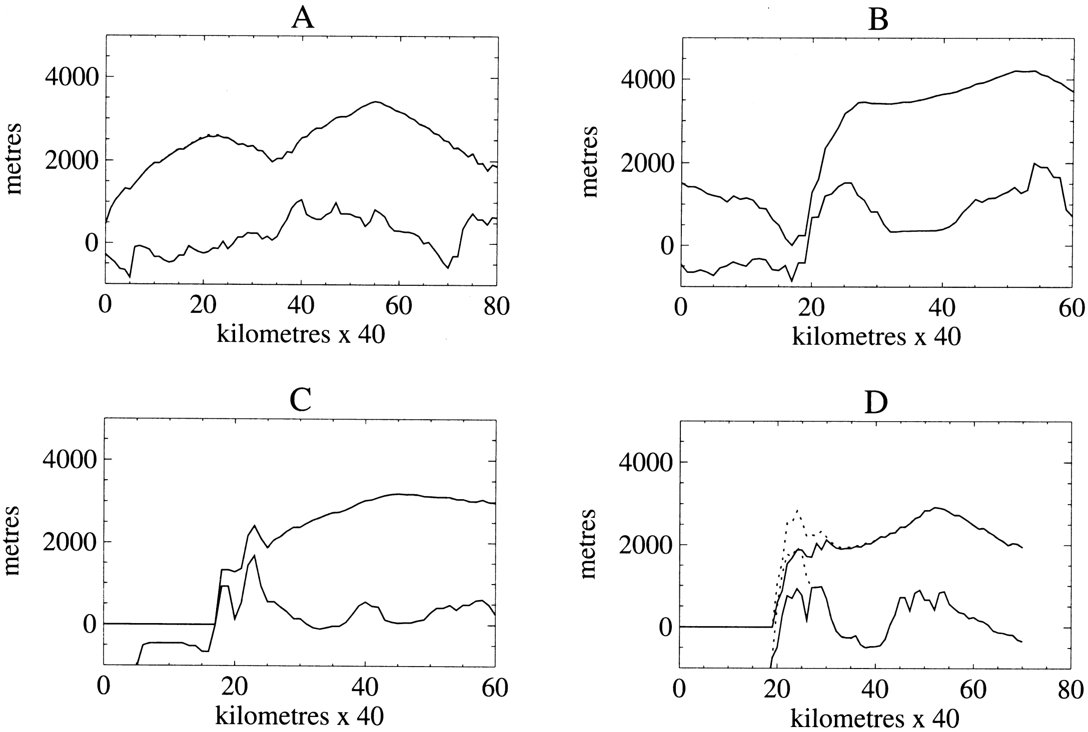

Fig. 7. The impact on the ice surface and bedrock elevations of lowering Victoria Land by 1 km, shown using the same four transects as Fig. 3. The change in bed elevations can be seen in (D), which shows the transect across the Victoria Land region. Lowering the bedrock elevation leads to the loss of the present ice centre over that region.

bedrock of the Dry Valleys region; the effect on ice-sheet elevations is very localised, with negligible impacts inland over the Wilkes–Pensacola Basin. Over the Victoria Land mountains, ice surfaces are lowered by approximately 1 km in response to a 1 km lowering of the bedrock.

The impact of different climate profiles on the ice-surface over Victoria Land is again similar to the to the Dry Valleys region (Fig. 8). For a climate profile 5°C warmer than present there is little change, while a 10°C warming causes the ice-sheet margin to shrink inland to the Wilkes–Pensacola Basin, leaving an independent ice centre over the Victoria Land mountains.

These results suggest that the main ice-sheet could overflow the region when the bedrock is lower under the present climate regime but that under a warmer climate, representative of a more dynamic ice-sheet, the margin of the ice-sheet does not reach Victoria Land.

7.4. Sub-glacial troughs in the Wilkes–Pensacola Basin

The impact on ice-surface elevations of infilling two large sub-glacial troughs in the Wilkes–Pensacola Basin, running parallel to the upwarped Transantarctic Mountains, is shown in Fig. 9A–D. Bedrock elevations in the Basin are raised by up to 750 m and leveled at an elevation of approximately 500 m over its 3000 km length. Unsurprisingly, where bedrock surfaces have been raised, ice-surface elevations also increase, by up to 500 m. This effect is particularly pronounced at the ice-margin closest to the Weddell Sea (on the left of the profile in Fig. 9a).

Inland from the central mountain ranges, there is a slight increase in both bedrock and ice-surface elevations by approximately 100 m. Adjacent to and inland from the Dry Valleys region there is a higher ice surface elevation, of approximately 150 m, but

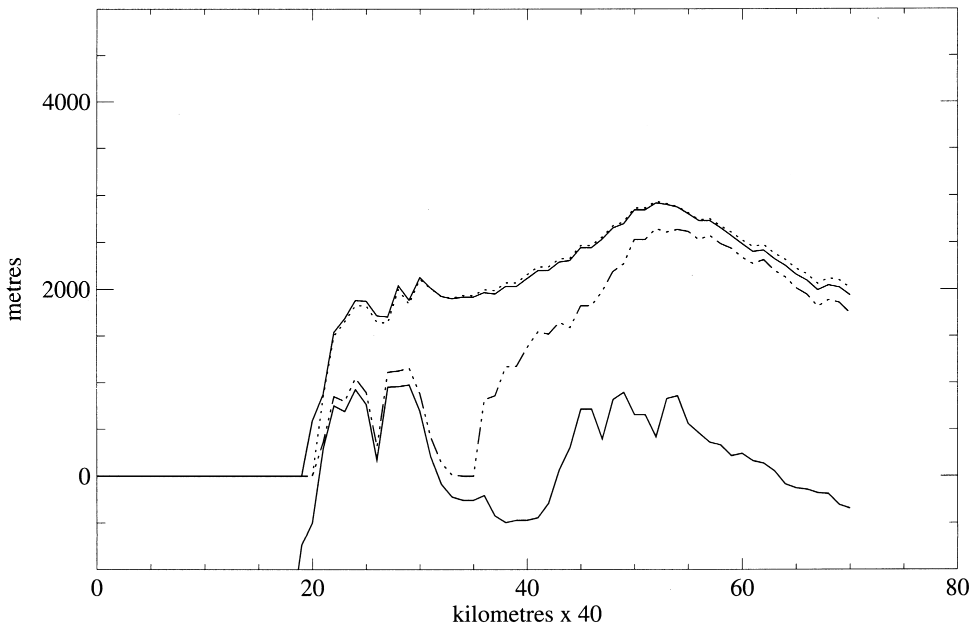

Fig. 8. As in Fig. 7C, but showing the change in ice surface elevations over Victoria Land resulting from applying different climate profiles: present-day (continuous line), +5°C (dotted line), +10°C (dash/dot line).

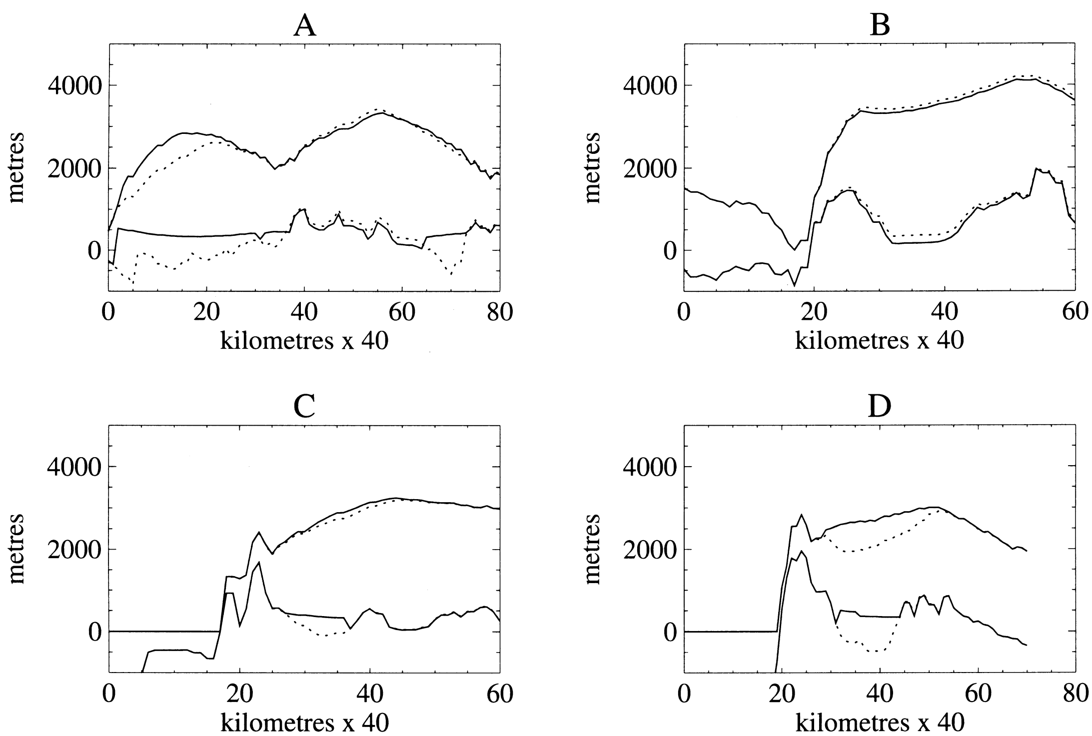

Fig. 9. The impact on ice surface and bedrock elevations of lowering the sub-glacial troughs in the Wilkes–Pensacola Basin. Bedrock elevations changes can be seen in all the transects. Transects A–D as in Fig. 3.

the ice centre over the mountain block remains, forcing ice to flow to either side. Adjacent to Victoria Land, ice surface elevations over the Wilkes–Pensacola Basin increase by up to 750 m in response to higher bedrock elevations of up to 750 m.

Higher bedrock elevations in the Wilkes–Pensacola Basin leads to thicker ice in the Basin. The resulting ice surface profile, and hence ice velocities, suggests more ice would flow across the Wilkes–Pensacola Basin towards the Transantarctic Mountains than down the Basin towards the Pacific and Weddell Sea as at present. However, the independent ice centres over Victoria Land and the Dry Valleys continue to stop ice from overflowing the respective mountain blocks and instead drive flow into troughs between them. Changing the climate profiles does not change this picture qualitatively.

7.5. Troughs through the upwarped Transantarctic Mountains

Fig. 10A–D displays the impact on ice-surface elevations of infilling five troughs through the upwarped Transantarctic Mountains between the central ranges and the Dry Valleys region. Inland from the central ranges, ice surfaces are up to 150 m higher over the entire 1200 km profile to the highest point on the ice-sheet. Similarly, inland of the Dry Valleys region, ice-surface elevations are raised by up to 300 m along 800 km of the profile. There is little change to ice-surfaces adjacent to Victoria Land. The profile along the Wilkes–Pensacola Basin (Fig. 10a) shows that blocking these outlet glaciers raises ice-surfaces for 1500 km parallel to the upwarp, by up to 750 m.

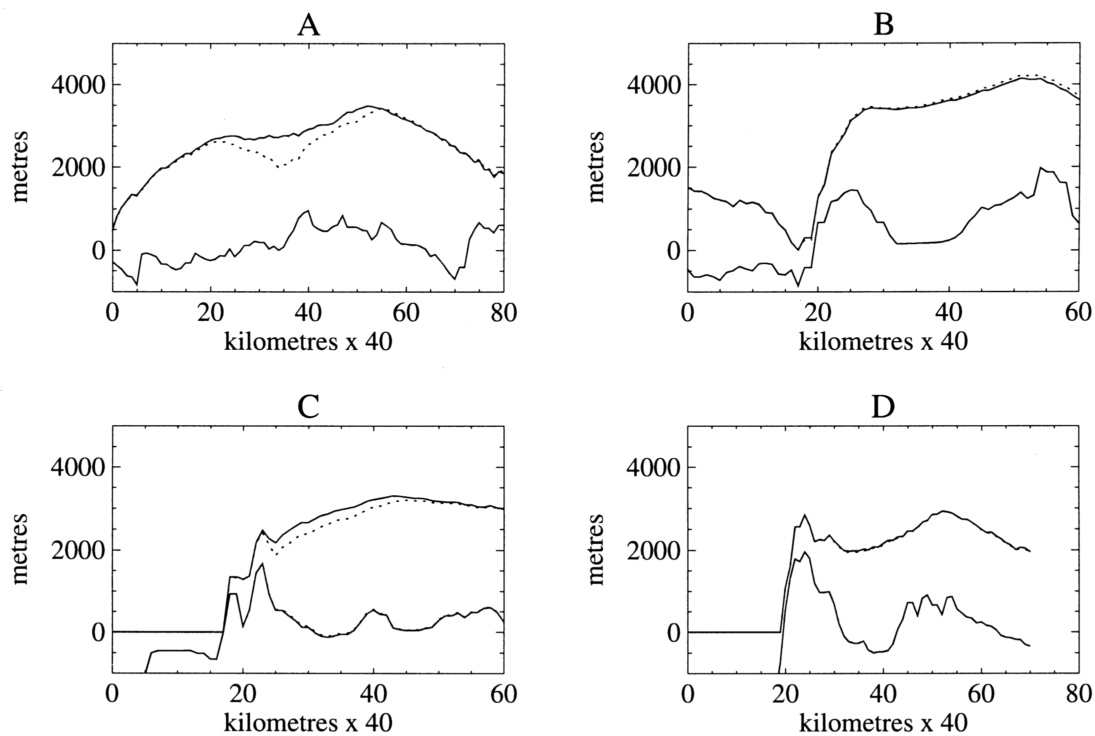

Fig. 10. The impact on ice surface and bedrock elevations of lowering five troughs through the Transantarctic Mountains between the central ranges and the Dry Valleys region. The transects are the same as those used in Fig. 3. These transects do not cross-cut the mountains by the infilled troughs, so the change in bedrock is not apparent, but do show the change in ice surface elevations resulting from the infilling of the troughs.

These results suggest that the progressive erosion of troughs through the marginal upwarp is leading to progressively lower ice-sheet surface elevations inland across a large proportion of the ice-sheet. Through time, this effect is making it increasingly difficult for the ice-sheet to override the Transantarctic Mountains. However, in the experiment displayed here, the ice-surface elevations over the Transantarctic Mountains are high enough to stop inland ice from overflowing the mountain summits, and instead the inland ice flows through other smaller troughs between and within the mountain blocks. Changing the climate by 5°C or 10°C does not qualitatively change this picture.

8. Discussion and conclusions

This paper explores the response of the East Antarctic ice-sheet to different bedrock topography under different climatic profiles. We show that tectonics, which change the surface elevation of the West Antarctic rift margin upwarp that comprises the Transantarctic Mountains, have a complex impact on the East Antarctic ice-sheet surface.

If Pliocene bedrock surface elevations in any of the Transantarctic Mountain blocks were 1 km lower than their present elevation, these results suggest that ice from the main ice-sheet overrode these blocks during that time. In the case of the central Transantarctic mountain ranges, there would be lower ice velocities across the mountain range, since the ice-sheet surface gradient is lower, but more ice would override the mountain summits than at present. Currently, ice is forced through a number of troughs between mountain massives, leading to high local ice velocities and the development of ice-streams. During Pliocene times, these features would be less well developed. In Victoria Land, inland ice would override the mountain block prior to the development of a local ice centre as bedrock elevations rose.

If during the Pliocene, sub-glacial troughs in the Wilkes–Pensacola Basin that are being eroded by the ice-sheet were less developed, ice surface elevations would be thicker along the length of the Basin. This would force more ice across the Transantarctic Mountains but, in the case displayed here, would not be sufficient to force ice to override the mountain blocks. Instead, steeper ice surface gradients would occur across the mountain upwarp, leading to more active glaciers and greater erosion in the troughs between mountain blocks. Similarly if troughs in the Transantarctic Mountains were less well developed during the Pliocene, ice surface elevations would be higher inland and lead to steeper surface gradients across the Transantarctic Mountains. Although not sufficient to override the mountain summits, it would again generate faster flowing and more erosive glaciers between the mountains.

If the surface uplift of bedrock elevations occurred much earlier than the Pliocene, when different climates were likely, then the picture is more complex. For the regions around the Dry Valleys and Victoria Land, a slightly warmer climate profile does not affect regional ice dynamics. However, when the climate profile is 10°C warmer, reflecting a smaller and more responsive ice-sheet, the main ice-sheet fails to reach the mountain blocks, which instead are covered by local ice centres. From this, we can suggest one possible sequence of events: during the Miocene, a warm climate and lower bedrock elevations resulted in small independent ice centres covering each mountain block. As the polar ice sheet developed, inland ice overrode the mountain blocks. As the bedrock elevations continued to rise through the Miocene and with climate increasingly similar to the present, ice centres again formed over the mountain blocks, diverting the inland ice to troughs adjacent to the blocks.

In the central ranges, unlike those more peripheral to the ice-sheet, warmer climates appears not to change the ice dynamics. As with the Pliocene scenario, ice overrides the mountains but there is a lower ice-surface gradient across the mountains, suggesting lower ice velocities and less developed ice streaming through the glacial troughs. Similarly, the impact of changing the climate when sub-glacial troughs are less developed is the same as in the Pliocene scenario. Higher inland ice elevations leads to higher ice-surface gradients across the Transantarctic Mountains, leading to the further development of troughs across this marginal upwarp.

These conclusions differ from those of Huybrechts (1993) because in that paper a uniform algorithm was applied to lower the whole of East Antarctica. The effect was to produce a very flat continent, which is susceptible, climatically, to large fluctuations in ice volume. In our experiments, mountain blocks are lowered more discretely along the West Antarctic rift margin but not elsewhere. This reflects more closely tectonic changes likely in East Antarctica since the inception of the main ice-sheet in the Miocene. In summary, raising or lowering individual Transantarctic Mountain blocks plays a relatively local role in changing ice-surface profiles and thus ice-dynamics, since local ice centres already force most inland ice to drain through the major troughs. In past times, when these troughs were less eroded, there was more ice in the Wilkes–Pensacola Basin, which was forced to drain through the Transantarctic Mountains with faster flowing and more erosive glaciers. This thicker ice was also forced to drain through lower-lying regions of the marginal upwarp, such as in the vicinity of the Dry Valleys or Victoria Land.

The limitations of this work stem from the necessarily schematic bedrock elevations used as inputs to the models and simplifications within the ice-lithosphere model. The most important limitation is the use of model grid-cells with dimensions of 40 km. These prevent the detailed demonstration of increased ice-surface gradients and hence ice velocities through the Transantarctic Mountains. Thus we do not assert that the surface elevations of the Transantarctic Mountain upwarp play no role in determining ice-sheet dynamics, but simply note that the modelling evidence points to the importance of the development of sub-glacial outlet troughs on ice-sheet dynamics. At present, insufficiently detailed modelling, and data, of the role of the major troughs on ice-sheet dynamics means this paper is necessarily speculative. However, the troughs are critical in determining ice drainage and thus ice dynamics in East Antarctica.

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