



## Early East Antarctic Ice Sheet growth recorded in the landscape of the Gamburtsev Subglacial Mountains <sup>☆</sup>

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### ABSTRACT

The Gamburtsev Subglacial Mountains are regarded as a key nucleation site for the Antarctic Ice Sheet and they retain a unique long-term record of pre-glacial and early glacial landscape evolution. Here, we use a range of morphometric analyses to constrain the nature of early glaciation and subsequent ice sheet evolution in the interior of East Antarctica, using a new digital elevation model of the Gamburtsev Subglacial Mountains, derived from an extensive airborne radar survey. We find that an inherited fluvial landscape confirms the existence of the Gamburtsev Subglacial Mountains prior to the onset of glaciation at the Eocene–Oligocene climate boundary (ca. 34 Ma). Features characteristic of glaciation, at a range of scales, are evident across the mountains. High elevation alpine valley heads, akin to cirques, identified throughout the mountains, are interpreted as evidence for early phases of glaciation in East Antarctica. The equilibrium line altitudes associated with these features, combined with information from fossil plant assemblages, suggest that they formed at, or prior to, 34 Ma. It cannot be ruled out that they may have been eroded by ephemeral ice between the Late Cretaceous and the Eocene (100–34 Ma). Hanging valleys, overdeepenings, truncated spurs and steep-sided, linear valley networks are indicative of a more widespread alpine glaciation in this region. These features represent ice growth at, or before, 33.7 Ma and provide a minimum estimate for the scale of the East Antarctic Ice Sheet between ca. 34 and 14 Ma, when dynamic fluctuations in ice extent are recorded at the coast of Antarctica. The implications are that the early East Antarctic Ice Sheet grew rapidly and developed a cold-based core that preserved the alpine landscape. The patterns of landscape evolution identified provide the earliest evidence for the development of the East Antarctic Ice Sheet and can be used to test coupled ice–climate evolution models.

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### 1. Introduction

East Antarctica hosts the largest and longest-lived ice sheet on Earth. Despite the importance of the East Antarctic Ice Sheet (EAIS) as both a responder to, and potential driver of global environmental and sea-level change, there are significant uncertainties about its early history and the scale and duration of its subsequent fluctuations (Barrett, 1996, 1999; Denton et al., 1984; Miller et al., 2008; Naish et al., 2008; Siegert and Florindo, 2009; Wise et al., 1991; Zachos et al., 1992, 2001). The EAIS is thought to have grown rapidly (Coxall et al., 2005) in response to a major decline in CO<sub>2</sub> levels at ca. 34 Ma

(DeConto and Pollard, 2003) and nucleated around the major highlands of East Antarctica, including the Gamburtsev Subglacial Mountains (hereafter, Gamburtsevs). Efforts to constrain the nature of early ice sheet growth in central East Antarctica have relied upon largely coastal records of EAIS behaviour, because of the paucity of data from the continental interior. However, geophysical and geological data from the interior are precisely what is required if the early patterns of glaciation in East Antarctica are to be constrained. Evidence of past EAIS behaviour is most likely to be preserved at sites where long-term erosion rates are extremely low and/or where cold-based ice preserves the landscape at the base of the ice sheet (Fabel et al., 2002; Naslund, 1997; Sugden et al., 1993, 1999; Summerfield et al., 1999). These conditions are found in central East Antarctica, where low ice velocities at Dome A (Rignot et al., 2011) are coupled with extensive areas of cold-based ice (Llubes et al., 2006), thus hindering subglacial erosion, as revealed in offshore deposits (Cox et al., 2010). Located in this region, the Gamburtsevs therefore

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represent a key site where a long-term record of both pre-glacial and glacial landscape evolution is likely to be found (Bo et al., 2009; Jamieson and Sugden, 2008; Jamieson et al., 2010).

Our aim is to understand the long-term landscape evolution of the Gamburtsevs, and in doing so to elucidate the dynamics of the early EAIS. In order to achieve this, we analyse the geomorphology of the Gamburtsevs at a regional-scale, using a new detailed and extensive airborne radar dataset, collected during the International Polar Year, as part of the Antarctica's Gamburtsev Province (AGAP) project (Bell et al., 2011; Ferraccioli et al., 2011). The AGAP survey provided a high resolution, regional, digital elevation model (DEM) of subglacial topography that greatly improves on the detail of previous continental-scale bedrock topography compilations (Le Brocq et al., 2010; Lythe et al., 2001) and the coverage of local surveys (Bo et al., 2009). We employ a series of morphometric techniques to quantify the geometry of the landscape and thereby map geomorphic features indicative of specific erosion processes, including fluvial erosion, warm-based glacial erosion and subglacial preservation. Our results are used to interpret the processes and patterns of landscape evolution in the Gamburtsevs, and to discuss their implications for the nature and timing of Antarctic Ice Sheet evolution.

## 2. Physiographic setting and origin of the Gamburtsev Subglacial Mountains

The Gamburtsevs lie beneath Dome A in the interior of East Antarctica (Fig. 1). They are bounded by the Pensacola Pole Basin to the south, Lake Vostok to the east and the Lambert Rift to the north. First discovered in 1958 by a Russian gravity and seismic survey (Sorokhtin et al., 1959), the Gamburtsevs are thought to be located in the middle of a Precambrian craton (Boger, 2011) and yet, unexpectedly, they retain a high elevation and significant relief. Prior to the AGAP project, the morphology, subglacial geology and deeper crustal structure of the Gamburtsevs were all poorly constrained, so that several contrasting models describing their origin remained largely untested (Cox et al., 2010; Fitzsimons, 2003; Sleep, 2006; van de Fliedert et al., 2008; Veevers, 1994). These models invoke ages for the Gamburtsevs' formation from the Cambrian (ca. 500 Ma) to the Cenozoic (30 Ma). However, based on analysis of the AGAP data, Ferraccioli et al. (2011) suggested that continental rifting processes provided the tectonic trigger for uplift of the Gamburtsevs at ca. 100 Ma. This was followed by fluvial (65.5–34 Ma) and then glacial erosion (34–14 Ma), causing renewed peak uplift.

## 3. Climate and ice sheet evolution

Past EAIS behaviour and key phases in Antarctic climate, ice sheet and surface process evolution over the last 100 Myr are derived from a mix of stratigraphic, geomorphological, geophysical and proxy data, combined with numerical modelling approaches (Supplementary Fig. S1). In the near-tropical climate conditions of the Early Eocene 'Greenhouse World', fluvial surface processes dominated (Baroni et al., 2005; Cooper et al., 2001; Francis et al., 2008; Pross et al., 2012). During these warmer periods, small, dynamic, ephemeral ice sheets may have formed on high elevation areas in the interior of East Antarctica (Birkenmajer et al., 2005; Cramer et al., 2011; Miller et al., 2005, 2008; Tripathi et al., 2005). At the Eocene–Oligocene (E–O) boundary (ca. 34 Ma) a shift to a cool-temperate climate marked the onset of widespread East Antarctic glaciation (Liu et al., 2009; Zachos et al., 2001, 2008). On- and off-shore sedimentary sequences indicate that between ca. 34 and 14 Ma (Supplementary Fig. S1, grey box) the ice masses on East Antarctica were warm-based and dynamic (Baroni et al., 2008),

fluctuating in pace with the Earth's orbital cycles (Escutia et al., 2005; Naish et al., 2001; Zachos et al., 1997, 2001). Glacial erosion was therefore a dominant agent of landscape modification, until a further decline in temperatures at ca. 14 Ma (Anderson et al., 2011; Lewis et al., 2007) resulted in a polar desert climate (Miller et al., 2008; Sugden and Denton, 2004; Zachos et al., 2001). This established a more stable continental-scale ice sheet and caused a switch from largely warm-based glaciation to a polythermal system (Anderson et al., 2011), where cold-based ice covered significant portions of the continent, reducing erosion rates across these areas (Armienti and Baroni, 1999; Ehrmann, 2001; Lewis et al., 2007; Miller et al., 2008).

## 4. Methodology

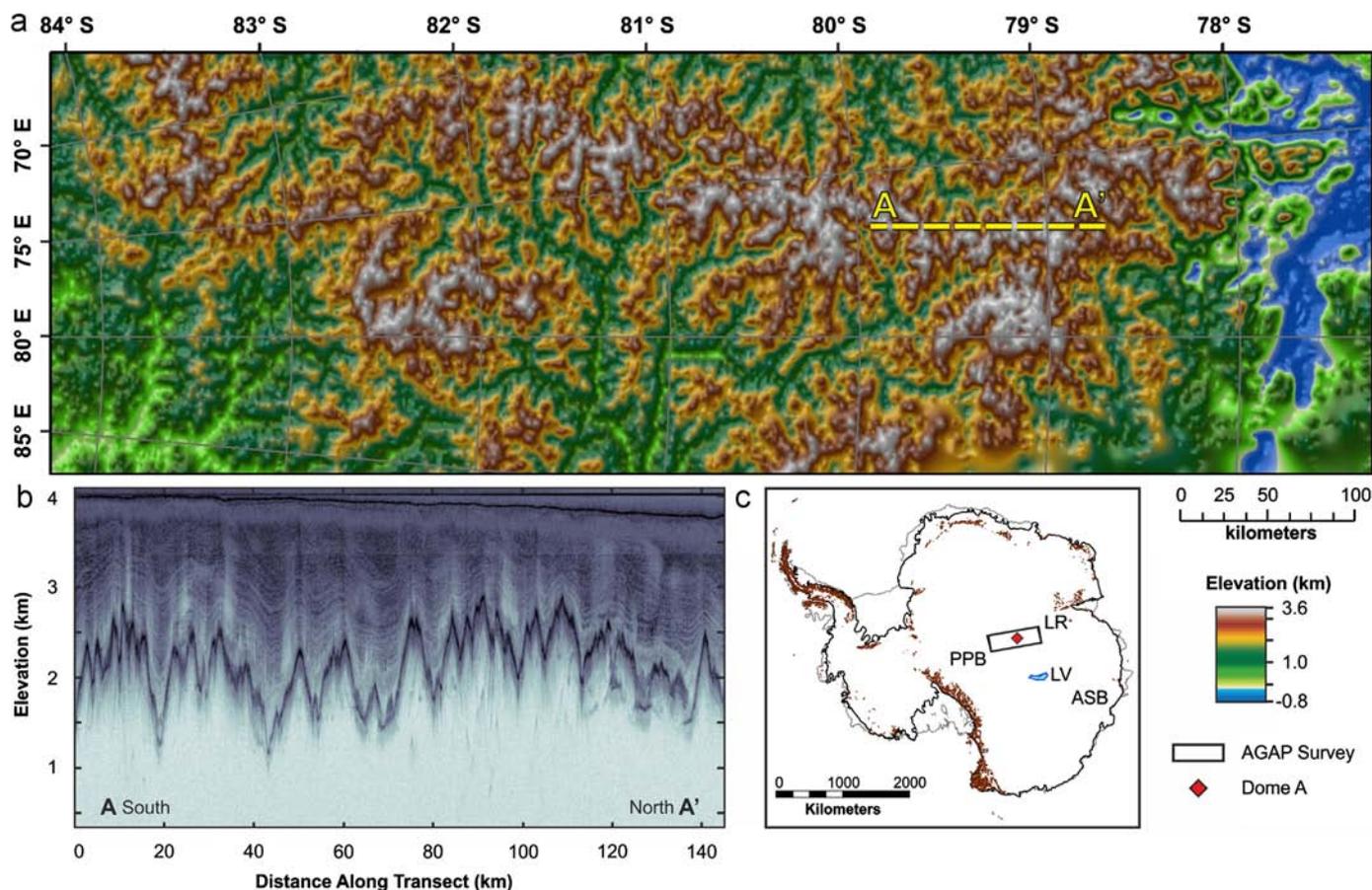
To understand the long-term patterns of landscape evolution in the Gamburtsevs, we analysed the geometry of an isostatically corrected DEM of subglacial topography generated from the AGAP airborne radar data. We then interpreted the morphometry in the context of former processes of landscape evolution and ice sheet dynamics. Our specific objectives were to

- generate a higher resolution DEM of the subglacial landscape of the Gamburtsevs;
- identify features that are representative of surface processes, which operate at local, regional and continental scales under warm and cold climates; and
- interpret patterns of long-term landscape evolution in the Gamburtsevs in the context of the interactions between topography, climate and ice sheet behaviour.

### 4.1. Data collection and DEM

The AGAP project completed a major aerogeophysical survey of the Gamburtsev Province during the 2008/09 field season, using two Twin Otter aircraft. 120,000 line-km of ice-penetrating radar, magnetic, gravity and laser measurements were collected in a detailed survey grid, with a line spacing of 5 km and tie lines 33 km apart (Bell et al., 2011; Ferraccioli et al., 2011). The study area covers 182,000 km<sup>2</sup>, encompassing most of the Gamburtsevs and extending into the southernmost margin of the eastern branch of the Lambert rift system (Fig. 1). Following earlier work (Siebert et al., 2005; Young et al., 2011), the airborne radar data are used as the basis from which subglacial landscapes and past ice sheet dynamics can be interpreted.

Cross-over analysis of the AGAP radar flight-line data indicates RMS errors in bedrock elevation of 64 m with a mean of 74 m. The bedrock elevation data were gridded onto a 1 km grid mesh using an iterative finite difference interpolation technique that employs a nested grid strategy to calculate successively finer grids until the user specified resolution is obtained (Hutchinson, 1988, 1989). Our grid was compared against previous DEMs of the Gamburtsevs area generated from the same data, using minimum curvature (Bell et al., 2011) and kriging algorithms (Ferraccioli et al., 2011). All three methods generate landscapes whose detailed structure is closely comparable. The gridded DEM was isostatically corrected to compensate for the removal of the modern ice sheet load (Ferraccioli et al., 2011) and to produce a topography that is hydrologically sensible (Fig. 1). The correction grid assumes a continuous plate with a uniform rigidity (average uplift of 500 m) and ignores the isostatic component of subsequent uplift related to erosion by incision (Wilson et al., 2012). The rebounded DEM was used as the basis for subsequent morphometric analyses, enabling us



**Fig. 1.** (a) Subglacial topography of the Gamburtsevs within the central AGAP survey grid (182,000 km<sup>2</sup>). A 1 km resolution grid, isostatically adjusted to account for ice sheet loading so that topographic elevations reflect ice free conditions. Yellow dashed line shows the location of (b) radar echo gram. (c) Inset shows the location of the central AGAP survey grid (black box) in Antarctica and the position of Dome A (red diamond), which marks the highest point on the East Antarctic Ice Sheet. Abbreviations: ASB—Aurora Subglacial Basin; LV—Lake Vostok; LR—Lambert Rift; PPB—Pensacola Pole Basin.

to assess the potential pre-glacial elevations of the region (Ferraccioli et al., 2011).

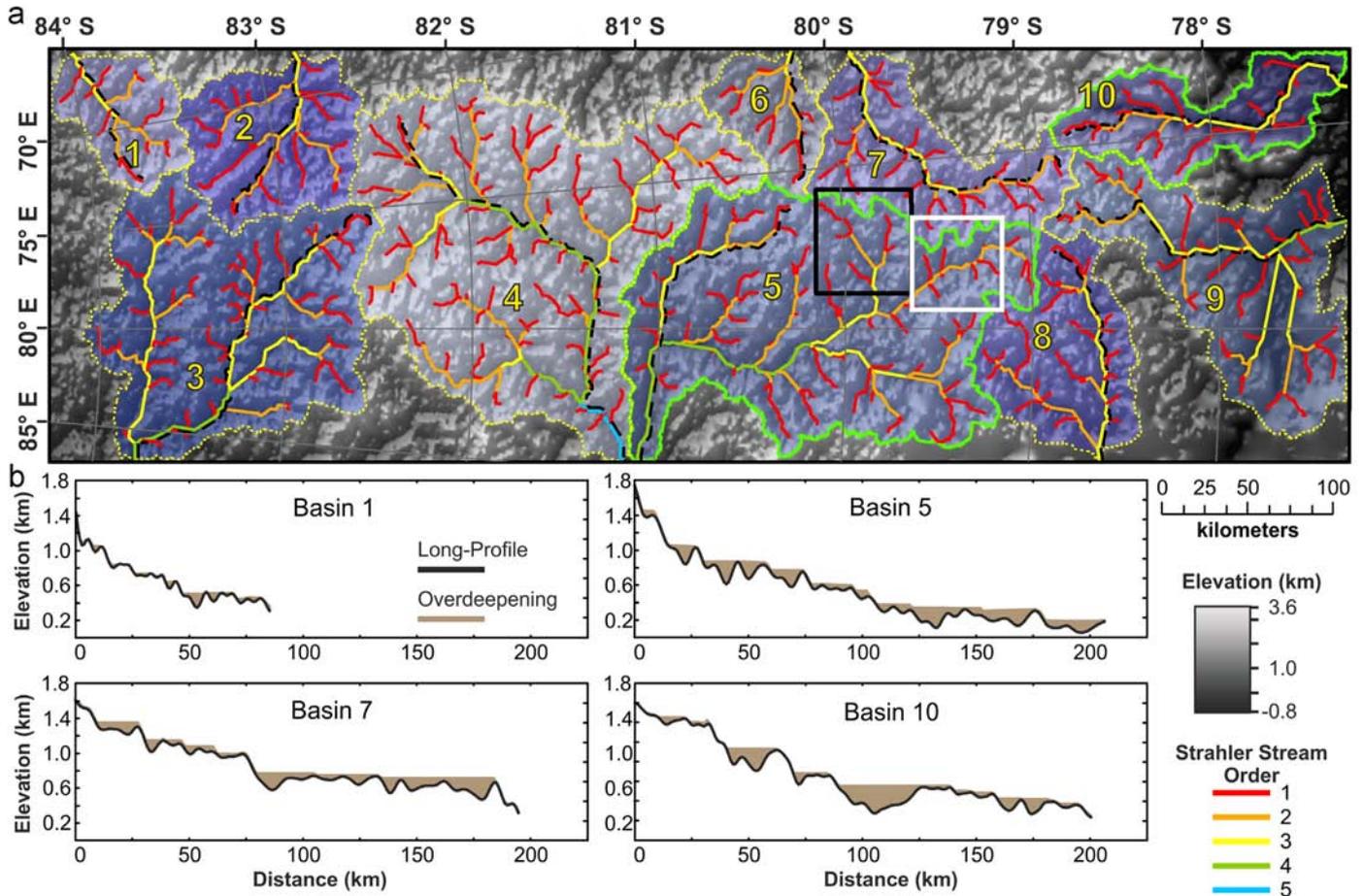
#### 4.2. Geomorphometry

We carried out standard multi-scaled analysis of geomorphic features in order to capture the potential variability in surface processes across the Gamburtsevs. The goal was to identify and quantify the scale and distribution of both pre-glacial (fluvial) and glacial landscape signals. A hydrological model, that assumes water flows down the steepest down-slope path, was used to identify flow paths and divide the landscape into a set of fluvial (ice free) drainage basins that reach from the central drainage divide on the mountain ridge to the edge of the survey area (Figs. 1 and 2). This division was used to analyse morphometry at a range of scales and allow the robust identification of geometric characteristics distinct to particular processes, which may have operated at particular scales (e.g. local vs. regional) within the mountain range. The morphometric analyses included drainage organisation, valley shape (long- and cross-profiles), and hypsometry (elevation–area distributions), all of which are outlined below.

Within the river networks (flow paths), the number of stream segments, their stream-order and bifurcation ratios were determined to quantify their planform geometry (Strahler, 1957; Supplementary Table S1). These data reflect the organisation of a river network, which can be perturbed by surface or tectonic processes. Stream order (Strahler, 1952, 1954) classifies rivers (and basins)

based on a hierarchy of stream segments and can indicate the size and drainage efficiency of the system. Globally, stream orders range between one (small headwater stream) and 12 (Amazon River). Bifurcation ratios reflect segment connectivity and range between three and five for natural stream networks on homogeneous lithologies. These statistics were compared against Horton's 'laws' for natural fluvial systems (Horton, 1945) to determine any significant overprint by non-fluvial processes. Drainage basin shape, order, relief statistics and area were also measured (Supplementary Table S1) and analysed spatially to identify geographic patterns that might be the result of distinctive processes of landscape evolution.

Valley long-profiles were extracted along the trunk of each drainage basin (Fig. 2). Graded river systems typically exhibit a smooth concave up form (Leopold and Bull, 1979; Strahler, 1954). Those modified by glacial erosion display a localised increase in tributary relief and a net reduction in profile concavity (Anderson et al., 2006; Brocklehurst and Whipple, 2006, 2007; MacGregor et al., 2000; Whipple et al., 1999). Profile steps, reflecting zones of erosional overdeepening, may also develop. These mark former regions of efficient glacial erosion around the equilibrium line altitude (ELA), where maximum ice discharge occurs (Brocklehurst and Whipple, 2006; MacGregor et al., 2000). A topographic lip at the downstream end of an overdeepening represents a former ice limit position (Benn and Evans, 2003; Sugden and John, 1976). The geometries of 294 valley cross-profiles, located perpendicular to the long-axis of valley networks, were then sampled directly from the 20 m horizontal resolution flight-line data (Supplementary Fig.



**Fig. 2.** Drainage characteristics for the Gamburtsevs under ice free conditions (Supplementary Table S1). (a) Drainage basins and river networks. There are 10 main drainage basins (yellow dashed outline and shaded blue), ranging in order from 3 to 5 (Strahler, 1954). Note: the southern Gamburtsevs relates to basins 1–3, the central Gamburtsevs to basins 4–8, and the northern Gamburtsevs to basins 9–10. Hypsometric distributions for basins 5 and 10 (additional green dashed basin outline) are shown in Supplementary Fig. S3a and b, respectively. River networks are coloured according to their stream order, which ranges from 1 to 5 (Strahler, 1954). All river networks obey Horton's (1945) 'laws' for natural fluvial systems. Black dashed lines mark the main valley long-profiles extracted from each drainage basin. The valley long-profiles have been modified from the smooth concave form of a graded river profile. Focused glacial erosion has formed steep valley headwalls, hanging valleys and overdeepenings (shaded in brown). The black and white boxes relate to the topography shown in Fig. 5a and b, respectively.

S2). Valleys widths were limited to the first lip where the valley wall either flattened off or showed a continuous decrease in elevation. Cross-profiles were classified according to a general power law coefficient  $b$  (Pattyn and van Huele, 1998), which relates valley form primarily to either fluvial or glacial erosion processes (Aniya and Welch, 1981; Harbor, 1990; Hirano and Aniya, 1988). Typically, cross-profile development resulting from glacial erosion is modelled to evolve from originally V-shaped valleys ( $b < 1$ ), reflecting little glacial modification, to steady-state, quasi-parabolic (or U-shaped) cross-profiles ( $b > 1$ ), dominated by glacial modification (Harbor, 1990, 1992).

Hypsometric analysis quantifies the distribution of land surface area with altitude (Strahler, 1952) and is commonly used to understand the relationship between local and regional tectonics and the spatial variability in fluvial and glacial surface processes (Montgomery et al., 2001; Pedersen et al., 2010). Hypsometry is generally scale-dependent (Brocklehurst and Whipple, 2004) and we therefore quantified it at three scales: 3rd-order basin, sub-range, and mountain range. The sub-range scale divided the mountain range into three sectors (southern, central and northern) and the 4th-order basins within these areas were compared (Supplementary Table S1). Range-scale hypsometry assessed the Gamburtsevs as a single unit, delineated by the extent of the central AGAP survey grid (Fig. 1a). These comparisons highlight whether the dominant signal of landscape erosion varies at different scales across the Gamburtsevs (Brocklehurst and Whipple, 2004).

In the final step of our analysis we identified and mapped geomorphic features that are indicative of specific modes and scale of erosion processes. In glacial environments, these features include cirques, hanging valleys, glacial troughs, and glacial breaches. These aid direct interpretation of former ice limits or, in the case of glacial breaches, which are generated when ice flow crosses a topographic drainage divide, a change in ice flow pattern and scale (Haynes, 1977; Sugden and John, 1976). Furthermore, the generation of glacial landforms is strongly tied to the basal thermal regime (i.e. warm or cold) of an ice mass and therefore provides an indication of evolving ice dynamics (Benn and Evans, 2003). We note that the dimensions of the glacial cirques identified in the Gamburtsevs ( $\leq 5$  km) are greater than those typically associated with cirques ( $\sim 2$  km) (Gordon, 1977; Krížek et al., 2012). However, they are comparable to the alpine valley heads described by Haynes (1995), who found that cirque widths of 5–6 km were not uncommon on the Antarctic Peninsula, and that overall, 87% of the features measured there were in the same range as those in northern Scotland (Gordon, 1977). We, therefore, retain the term 'cirque' for these alpine valley heads because these features are morphologically akin to the first stages of localised ice formation.

## 5. Results

The DEM shows that the rebounded topography has a linear form, with a central mountain ridge running NNE–SSW (Fig. 1).

Although the AGAP survey does not extend beyond the foothills of the Gamburtsevs, it is evident that the range is characterised by asymmetry from west to east. The topography is extremely rugged with relief averaging 2.25 km in the mountain core and 4 km on the northern flank, bordering the Lambert Rift (Supplementary Table S1). High elevation peaks along the mountain ridge range in height between 2.5 and 3.5 km above sea level. The mountain foothills are generally located between 0.8 and 1.3 km, but in the north, towards the Lambert Rift, valley floors reach almost a kilometre below sea level. Valley networks descend from either side of the mountain ridge. They are typically 10–25 km wide and 100–200 km long, and their planform geometries display a dendritic pattern. Amongst these networks, two significant, east–west trending, trunk valleys dominate the mountain topography (Fig. 1). Up to 350 km long, they extend from the mountain ridge towards Lake Vostok and the inferred rift basins that surround the Gamburtsevs (Ferraccioli et al., 2011). They drain the majority of the east-facing mountain flank through a topographic low that is ~22 km wide. These valleys are orientated parallel to each other in their lower reaches and their proximity at the edge of the AGAP survey area suggests that they converge down-slope. In the south-east sector of the Gamburtsevs foothills, another large network is orientated towards the present Byrd–Nimrod glaciers. Along the northern margin of the Gamburtsevs, a number of deeply incised linear valleys extend out from the mountain flank towards the eastern branch of the Lambert Rift. It is evident that the Gamburtsevs have sustained their high peaks and deep valleys (Fig. 1b), thereby retaining an alpine style landscape.

## 6. Signatures of long-term landscape evolution

Here, we discuss the fluvial and glacial geomorphic signatures identified within the landscape of the Gamburtsevs by a series of quantitative and qualitative analyses, as outlined in Section 4.

### 6.1. Drainage characteristics

Planform valley geometries and drainage characteristics are shown in Fig. 2 and Supplementary Table S1. These confirm that the majority of valley networks have a dendritic pattern, typical of fluvial systems overlying homogeneous bedrock without strong structural controls (Haynes, 1977; Summerfield, 1999). Mean bifurcation ratios of four demonstrate that networks are well organised in such a way as to obey Horton's 'laws' for fluvial systems (Horton, 1945). Ten primary drainage basins extend from the mountain ridge, with average areas of ~13,400 km<sup>2</sup>, average lengths of 172 km and stream orders (Strahler, 1954) of between 3 and 5 (Supplementary Table S1). Elongation and circularity ratios ( $R_e > 0.6$  and  $R_c \leq 0.4$ , respectively), indicate that, with the exception of basin 10, basins have a distinct pear shape, a form that is typically associated with systems that are in balance with the tectonic regime and have efficient run-off discharge (Christopher et al., 2010; Miller, 1953; Schumm, 1956; Strahler, 1964).

The long-profiles derived for each drainage basin are modified (to varying degrees) from the smooth concave form of a graded river profile (Fig. 2). The dominant pattern shown is a stepped geometry, where short steep topography is followed by a flattening of the profile. These steps are often accompanied by higher elevation downstream lips, reflecting focused erosional overdeepenings. Steps range in size, but are longer (~50–100 km) at lower elevations, than at higher elevations (~5–20 km). Additionally, long-profiles generally show an increase in relief in the upper channel reaches (steep headwater valleys) and an overall reduction in profile concavity (Fig. 2).

All cross-profile morphologies have a muted signal that indicates various stages of transition (Supplementary Fig. S2), either through widening or vertical down-cutting, towards a U-shape, the idealised best fit form for a glaciated valley (Harbor, 1990). The majority of cross-profiles sampled have an intermediate form or parabolic shape (47%), suggesting a degree of glacial modification throughout the Gamburtsevs (Aniya and Welch, 1981; Harbor, 1990; Hirano and Aniya, 1988). The proportional distribution of profile forms in each basin does not vary significantly with elevation. Spatially, however, V-shaped cross-profiles are most numerous in central basins 4–6 (> 47% per basin), but generally decrease in number in southern and northern basins (e.g. < 13% in basins 3, 9 and 10), whilst the proportions of parabolic and U-shaped cross-profiles increase (e.g. ≥59% parabolic in basins 3 and 10; 67% U-shape in basin 9).

### 6.2. Hypsometry

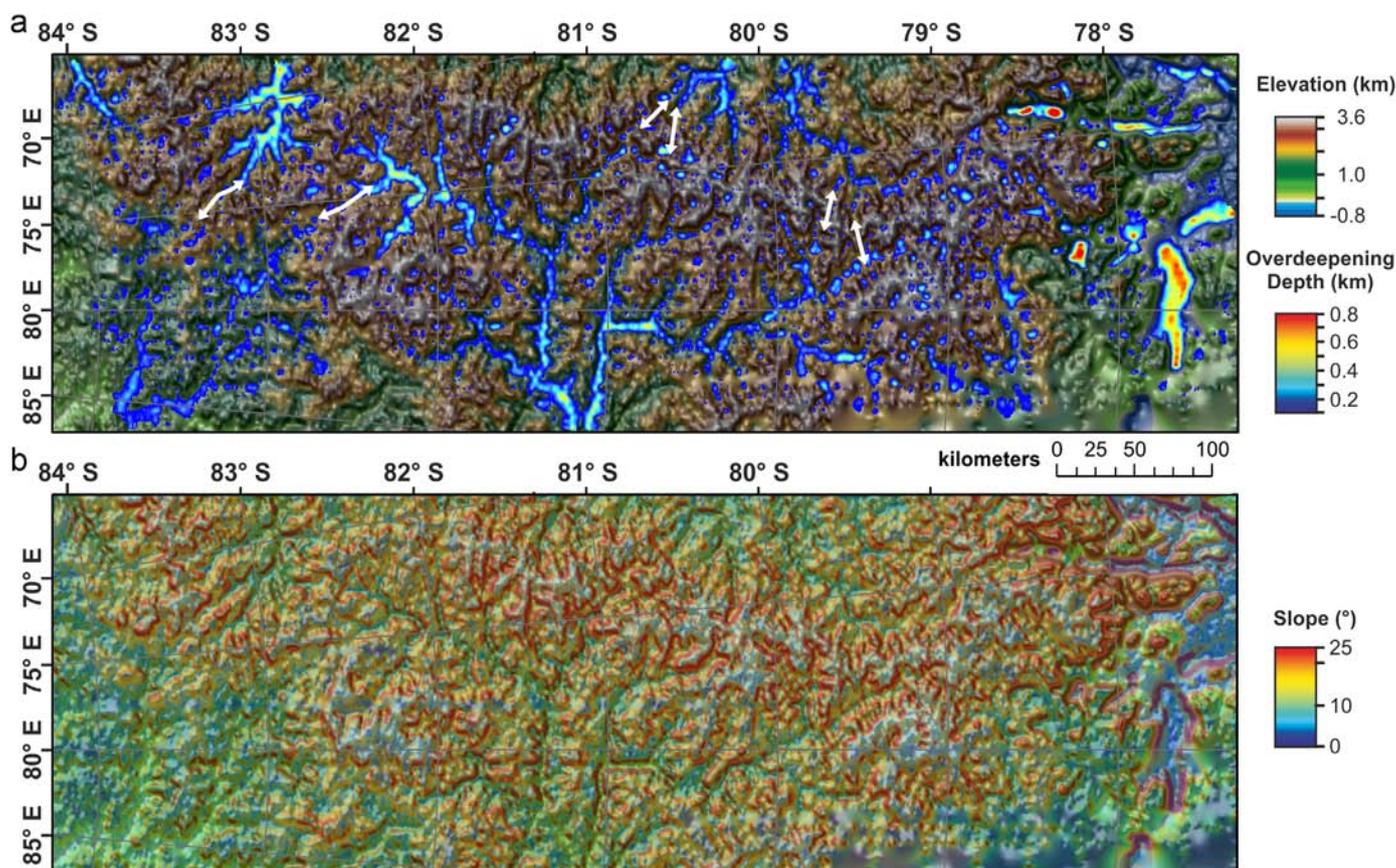
Two distinct styles of hypsometry exist across the mountain range, independent of scale (Supplementary Fig. S3). The first style, shown in basins 1–8 and the Gamburtsevs as a whole, displays a single maximum skewed upward ( $H_{\max}$  values in the range of 2.0–2.4 km), demonstrating that a significant portion of the landscape reaches mid to high elevations (Supplementary Fig. 3a and c). This is consistent with regions of alpine glaciation (Egholm et al., 2009; Supplementary Fig. S3c inset). The second style is found in basins 9 and 10, on the northern flank of the Gamburtsevs (Fig. 2). Here, the hypsometric distribution shows multiple maxima that correspond to the high elevation mountain peaks, the mountain foothills and the beds of the deeply incised valleys at the edge of the Lambert Rift (Fig. 1). The maxima are almost evenly distributed between high (2.6 km), intermediate (0.5 km) and low (–0.3 km) elevations (Supplementary Fig. S3b). This multi-modal signal indicates a more complex erosion history for these basins.

### 6.3. Glacial landforms

Glacially-formed features are widespread across the Gamburtsevs landscape (Figs. 3–5). Glacial cirques are particularly common along the central mountain ridge (Fig. 4) and are characterised as small, bowl-shaped, steep-sided valley heads (Fig. 5). Cirques have an even aspect distribution (frequency range of 6% in eight directions). The majority of cirque floor elevations fall into two bands at 2.1–2.2 km and 2.4–2.5 km (Fig. 4b). Cirques are more prevalent in the north, becoming fewer and more dispersed in the southern Gamburtsevs. In several areas, suites of hanging valleys were found directly above, and feeding into, broad, linear, U-shaped valleys, close to the central drainage divide (Figs. 3 and 5). U-shaped troughs, that correspond with valley long-profile overdeepenings (Fig. 2), are found at different scales throughout the study area and are largest towards the foothills of the Gamburtsevs (Figs. 1 and 3). Significantly streamlined topography dominates the networks feeding the Lambert Rift (Figs. 1 and 2). There are few, very minor, glacial breaches found between drainage basins that lie along the mountain ridge drainage divide (Fig. 3a).

## 7. Sequence of geomorphic development

Here, we interpret how the morphometric data fit together in a sequence of fluvial and glacial events that lead to the growth and expansion of the EAIS across the Gamburtsevs. We provide evidence for overprinting of the inherited fluvial landscape by two distinct scales of successively larger ice coverage, for which we present interpretative maps (Fig. 6).



**Fig. 3.** (a) Distribution of overdeepenings and potential small-scale high elevation glacial breaches (white arrows) overlain on subglacial topography and (b) surface slope, semi-transparent layer overlain on subglacial topography.

### 7.1. Inherited fluvial landscape

The analysis of hydrological flow networks in the Gamburtsevs provides evidence for a widespread, high-elevation, fluvially-dominated valley geometry preserved beneath the central EAIS (Fig. 2; Supplementary Table S1). The regional valley orientations are inconsistent with modern ice flow configurations that currently cross-cut the valleys that drain the west and east mountain flanks. Given that the Dome A ice divide is unlikely to have shifted significantly under polar ice sheet conditions, the valley networks were therefore formed before the present-day continental-scale EAIS was established (Bo et al., 2009; Ferraccioli et al., 2011). It is clear that fluvial erosion played a key role in incising the mountain range (Fig. 1). This pre-glacial signal has been re-enforced as it has driven subsequent erosion and deposition patterns via topographic steering (Sugden and John, 1976; Kessler et al., 2008). This is demonstrated where the planform geometry of the fluvial topography has been retained, but the morphology of valleys has been significantly altered by warm-based, erosive ice. The modified long-profiles (Fig. 2), for example, reflect the formation of glacial overdeepenings at a range of scales. At mountain-scale, the overall hypsometry of the Gamburtsevs (Supplementary Fig. S3c) also demonstrates a dominantly glacial, rather than fluvial, signal of erosion (Egholm et al., 2009), confirming the widespread overprint of glaciation on the inherited fluvial landscape.

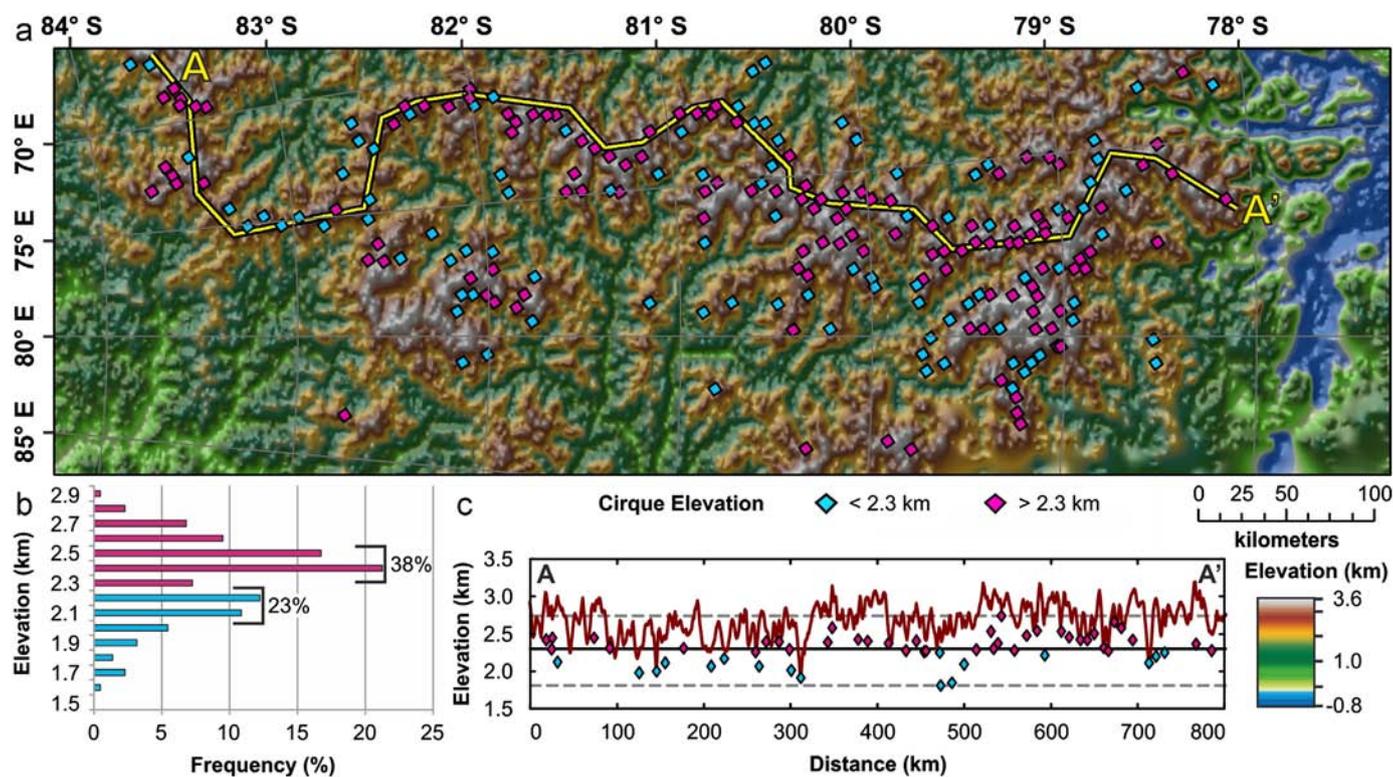
### 7.2. High elevation glaciation

The distributions of high elevation (> 2 km) cirques (Fig. 4) reflect the initiation points for glacier development. As such, they provide evidence for localised warm-based glaciation that marks

the onset of glaciation in East Antarctica (Fig. 6a). These small-scale ( $\leq 5$  km), bowl-shaped, overdeepenings are apparent in the upper reaches of the valley long-profiles (Figs. 2 and 5). The lack of pattern in cirque orientation may reflect the polar location of the Gamburtsevs and therefore the relatively even exposure of the mountains to the sun. Alternatively, it may indicate that there was no preferential moisture source direction during the growth of the cirque glaciers. In several basins, overdeepenings are positioned slightly further down-valley than in the majority of cirques, marking the position of larger, valley-confined, headwater glaciers. We infer these overdeepenings are associated with the formation of localised high elevation independent ice caps (Fig. 6a). These formed where the highest elevation cirques coalesced into larger ice bodies with enough mass to drive a limited expansion of glaciers into the headwaters of the existing fluvial valleys. Consistent with this, basin hypsometry shows a skew towards higher elevations (Supplementary Fig. S3a) that is interpreted as the development of cirque glaciers as a landscape transitions from non-glaciated to glaciated conditions (Brocklehurst and Whipple, 2004).

### 7.3. Mountain glaciation

The coalescence of localised ice caps across the Gamburtsevs, leading to a phase of widespread alpine glaciation, is evidenced by the distribution of overdeepenings, long- and cross-profile forms and hypsometry. A small number ( $\sim 7\%$ ) of cirques, located distal to the mountain ridge, are found at lower elevations (< 2 km) (Fig. 4). These features are more dispersed, indicating that they did not co-evolve with those on the ridge, but instead reflect a gradual lowering of the ELA and growth of East Antarctic glaciers.



**Fig. 4.** (a) Location of cirque overdeepenings (diamonds) overlain on subglacial topography with a transect of topography taken along the mountain ridge (yellow line). (b) Histogram of cirque floor elevations (deepest point of overdeepening). Note: the majority of cirques lie within two elevation bands between 2.1–2.2 km and 2.4–2.5 km. (c) Elevations of cirques (diamonds) located within 5 km of the mountain ridge line transect (yellow line panel 'a'). Lines show average (black) and maximum and minimum (grey hatched) cirque floor elevations.

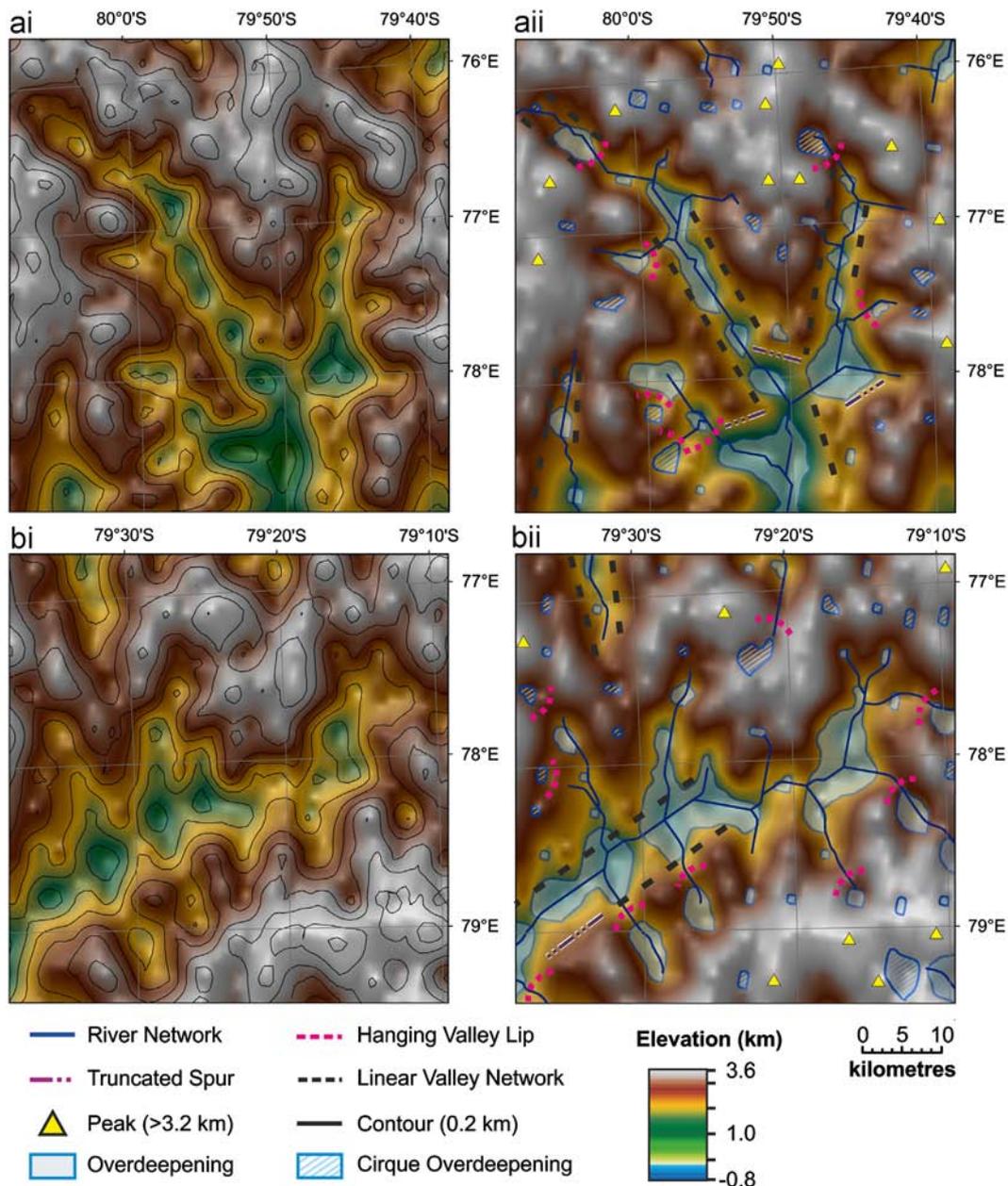
At all scales, hypsometric distributions show evidence for alpine glaciation (Supplementary Fig. S3a and c), where a large proportion of land at high elevation reflects the formation of deeply incised valleys that leave extensive areas at relatively high elevations (Summerfield, 1999). The widespread distribution of parabolic cross-profiles, across a range of elevations (Supplementary Fig. S2), reflects a degree of landscape modification that is typically imposed by warm-based outlet glaciers (Harbor et al., 1988; Hirano and Aniya, 1990; Sugden and John, 1976). Long-profiles show steep valley headwalls, cirque overdeepenings and hanging valleys, forming in the upper reaches of drainage basins (Figs. 2 and 5), whilst more significant overdeepenings at lower elevations are associated with steep-sided, broad linear valley networks and truncated spurs (Fig. 5). These overdeepenings are particularly evident in the main trunk valleys and are displayed as steps in the valley long-profiles (Figs. 2 and 3). The general increase in step size away from the mountain divide also suggests that gradually larger troughs are being eroded under larger ice masses, as is consistent with other mountain ranges as glaciers grow and coalesce (Anderson et al., 2006). This evidence is the clearest indication that selective linear erosion, by dynamic warm-based ice, has modified the landscape of the Gamburtsevs.

Overdeepenings of a common depth and the locations of their associated downstream topographic lips were used to define former ice limits for a transient phase of extensive valley glaciation (Fig. 6b). We associate this pattern with a dynamic, mountain-scale, ice sheet that formed as the ELA descended, causing the previously isolated mountain ridge ice caps (Fig. 6a) to coalesce and encompass the core of the Gamburtsevs. This is highlighted by the presence of a few small-scale glacial breeches (Fig. 3a), located along the mountain ridge in areas where the once localised ice caps joined together, resulting in a change of ice flow direction across drainage divides. Ice flow would then follow the geometry

of the topography, enabling rapid ice discharge and erosion at a regional-scale. We envisage this ice sheet would be similar to present day South Patagonia with a low-gradient, high elevation ice field being drained by significant, fast-flowing, glaciers that occupy the pre-existing fluvial valley networks (Rabassa and Clapperton, 1990).

#### 7.4. Continental-scale glaciation

The geomorphic features identified in the Gamburtsevs (basins 1–8) are all indicative of alpine-style glaciation, but signals associated with glacial erosion under continental-scale ice flow are not observed. The retention of the inherited dendritic valley system (Fig. 2), for example, provides strong evidence for the survival of a purely alpine system, where selective linear erosion by warm-based ice is controlled by topographic steering (Sugden and John, 1976). The fact that the geometry of valley networks cross-cuts modern ice flow configurations also demonstrates that they were produced under different palaeo-environmental conditions (e.g. Young et al., 2011). Conversely, active continental-scale ice sheets result in extensive breaching of multiple watersheds, enhancing basin and drainage inter-connectivity (Haynes, 1977; Sugden and John, 1976), and produce an overall smoothing and streamlining of the topography (Benn and Evans, 2003). In the Gamburtsevs, we find high relief topography that is too rough and steep to support the presence of fast flowing outlets, such as ice streams. Furthermore, hypsometric distributions are indicative of the development of cirque glaciation (Supplementary Fig. S3a and c). In contrast, extensive long-lived glacial modification (associated with a lowering of the ELA) tends to shift hypsometric maxima towards lower elevations (Brocklehurst and Whipple, 2004; Egholm et al., 2009). This confirms that the Gamburtsevs are a landscape of alpine glaciation, implying persistent preservation of



**Fig. 5.** Two examples of subglacial topography showing evidence for glacial geomorphic features indicative of mountain-scale alpine glaciation. Locations of panels (a) and (b) shown as black and white boxes, respectively, in Fig. 2. Interpretations are based on the style and distribution of overdeepenings, surface slope, river networks and bed elevation. (ai) Topography with 0.2 km contours (range: 1.4–3.4 km), (aii) glacial geomorphic features, (bi) topography with 0.2 km contours (range: 1.8–3.4 km), and (bii) glacial geomorphic features. The mountain-scale, alpine geomorphology highlights high elevation peaks, steep valley headwalls with small-scale ( $\leq 5$  km), bowl-shaped, cirque overdeepenings in the upper reaches of drainage basins with more significant overdeepenings at lower elevations, associated with steep-sided, broad linear valley networks and truncated spurs.

the bed under, long-term, minimally erosive conditions (Cox et al., 2010).

In contrast, the northern flank of the Gamburtsevs (basins 9–10) displays a different signal of glaciation that is consistent with persistent and significant ice flow speeds in a uniform direction. In this region, the eastern branch of the Lambert Rift abuts the northern flank of the mountains (Fig. 1). Here, we observe extensive linear erosion, in the form of well defined topographic channels, or a fjord-like landscape, resulting from large-scale, dynamic ice fluctuations (Bennett, 2003; Kessler et al., 2008). Relief is much greater (average 3.9 km) as valleys descend beyond the foothills of the mountains into the overdeepened (below sea level) troughs of the Lambert Rift (Supplementary Table S1). This change in topography is evidenced at basin-scale by

multi-modal hypsometries (Supplementary Fig. S3b). These may represent two different scales and/or stages of glacial erosion, as well as the influence of basin specific conditions, such as the proximity of the Lambert Rift. Long-profiles from basins 9 and 10 have the steepest gradients, as implied by the high basin relief (Fig. 2; Supplementary Table S1), and display the largest overdeepenings in both depth and width (Fig. 3a). The latter is also demonstrated by the increase in the number and size of U-shaped cross-profiles in the northern basins (e.g. 67%–basin 9; Supplementary Fig. S2), which reflect intense lateral glacial erosion of valley floors under a single ice sheet (Graf, 1970; Hirano and Aniya, 1988). The extensive linear erosion evident in this region is in keeping with continental-scale ice cover and might be expected given the inferred long-term consistency of ice flow direction

across this region (Arne, 1994; Jamieson et al., 2005), the associated net increase in ice velocities (Rignot et al., 2011) and basal melt rates (Llubes et al., 2006) north of the mountain flank, and the strong tectonic and topographic control exerted by the Lambert Rift (Kessler et al., 2008; Ravich and Fedorov, 1982).

## 8. Discussion

We consider that the morphometric data reflect landscape development from a Greenhouse climate dominated by fluvial processes, to an Icehouse climate, when glaciers and subsequently ice sheets of varying size become established in central East Antarctica. Traditionally, proxies for past EAIS behaviour are derived from the periphery of the continent, from global records or models and dominantly relate to variations following ice sheet formation (Section 3; Supplementary Fig. S1). More recently, however, new datasets are starting to emerge that have higher resolutions and/or encompass timescales prior to 34 Ma (e.g. Cramer et al., 2011). Here, we compare our geomorphic signatures against a range of proxies. We discuss what implications the patterns of landscape evolution identified might have for understanding climate and the timing of early glaciation and subsequent ice sheet evolution in East Antarctica. Our inferences are meant to represent a framework that can be adapted into the future as new and more detailed proxy evidence comes to the fore.

### 8.1. Cirque formation and palaeoclimate

We interpret the high elevation cirques as the inception points for glaciation in East Antarctica (Fig. 6a). However, the timing of their formation, either at the E–O boundary or earlier from the Late Cretaceous (Miller et al., 2005, 2008), remains difficult to constrain, due to the lack of proximal geological constraints (Supplementary Fig. 1). To assess the timing of widespread ice growth in the Gamburtsevs, we examined the basal elevations of cirques, assuming they reflect the approximate ELA of the earliest ice bodies (Ohmura et al., 1992). Cirque floor elevations are dominantly found at 2.4 km and 2.2 km (Fig. 4b), giving a potential upper and lower boundary for the earliest ELA. The mean summertime temperature (MST) at the ELA of temperate glaciers is typically  $4.3 \pm 3$  °C (Ohmura et al., 1992). A range of MST lapse rates, representative of high latitudes ( $-6.5$  °C km<sup>-1</sup>—Kerr and Sugden, 1994;  $-7.6$  °C km<sup>-1</sup>—Krinner and Genthon, 1999), coastal conditions ( $-5$  °C km<sup>-1</sup>—Krinner and Genthon, 1999), and the standard dry adiabatic lapse rate ( $-9.8$  °C km<sup>-1</sup>—King and Turner, 2007), were then used to determine a range of MSTs at the coast of Antarctica at the time of cirque formation. We found maximum, minimum and average coastal temperatures of 27.8 °C, 16.3 °C, and 21.6 °C at 2.4 km and, 25.9 °C, 15.3 °C, and 20.2 °C at 2.2 km cirque floor elevations, respectively. These values are within growth ranges expected for southern beech (*Nothofagus*), which is known to grow before and during early glaciation (Anderson et al., 2011; Francis et al., 2009).

We also compare these temperatures with estimated MSTs at sea level in Antarctica for the E–O boundary, as derived from biotic (pollen, spores, leaves) proxies. In the Early Oligocene (34 Ma) suggested MST values range between 4 and 12 °C (Cantrill, 2001; Prebble et al., 2006; Raine and Askin, 2001). In contrast, we find warm month mean temperature estimates of 24 ( $\pm 2.7$ )–27 °C (Francis et al., 2008; Pross et al., 2012) during the Middle Eocene (48–40 Ma),  $\sim 23$  °C (Poole et al., 2005) in the Early Eocene (55–48 Ma), and  $25.7 \pm 2.7$  °C in the Late Palaeocene (59–55 Ma) (Francis et al., 2008; Supplementary Fig. S1). Our mean MST values show closer agreement with inferred climate prior to 34 Ma, implying that such coastal temperatures are not mutually

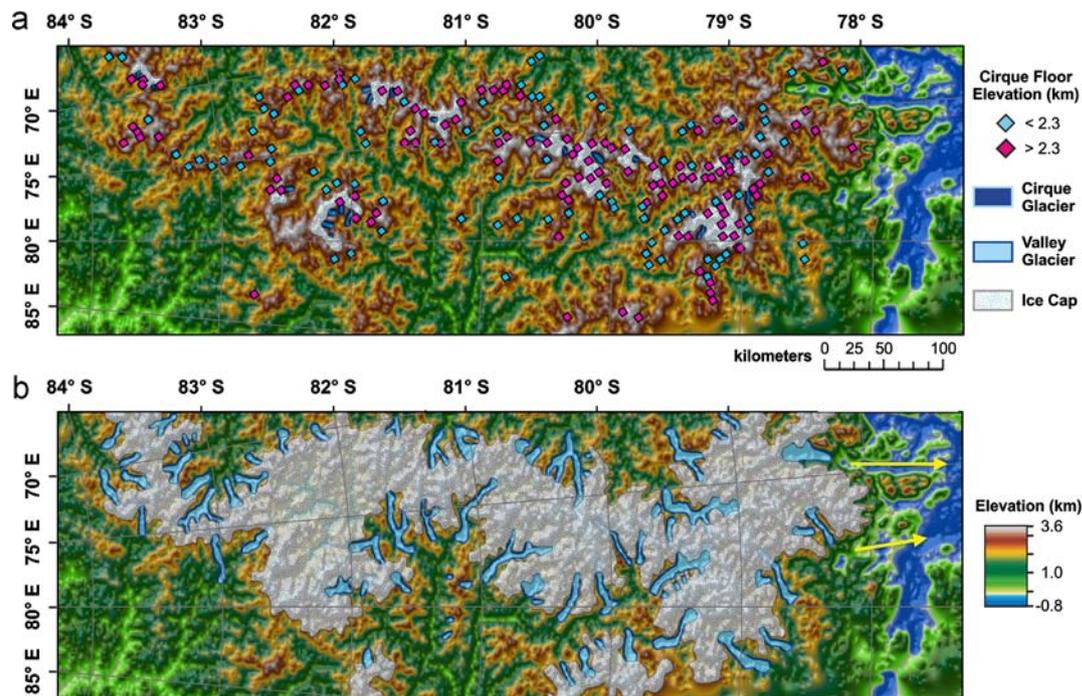
exclusive to cirque formation at high elevation in the Gamburtsevs at those times. Therefore, these early cirque glaciers could have formed at, or significantly before, 34 Ma. This differs from the interpretations of Bo et al. (2009) who suggest that the cirques they identified beneath Dome A were considerably younger, forming between 34 and 14 Ma (Supplementary Fig. S1, grey box). However, our findings are compatible with dated glacial sediments, biotic and geochemical proxies, and ice volume estimates, which show evidence for ephemeral glaciation in Antarctica during the Eocene (Anderson et al., 2011; Birkenmajer et al., 2005; Cramer et al., 2011; Miller et al., 2005, 2008; Tripati et al., 2005).

### 8.2. Ice sheet development

The preservation of glacial signals at a range of scales, within the landscape of the Gamburtsevs, has significant implications for interpreting the evolution of the early EAIS, following cirque glaciation. We suggest that once ice had expanded to a mountain-scale (Fig. 6b), the Gamburtsevs acted as a pinning point, whereby the margin of the EAIS consistently extended beyond the Gamburtsevs' boundary between ca. 34 and 14 Ma (Supplementary Fig. S1, grey box), when cyclical fluctuations in ice extent are identified at the coast of Antarctica (Naish et al., 2001). Given the length of time required for the development of 'glaciated' terrains (Harbor, 1992; Harbor et al., 1988; Jamieson et al., 2008), the overdeepenings preserved within the alpine landscape of the Gamburtsevs could be generated over a net period of between 0.3 and 3 Myr, if we assume erosion rates of 0.1–1 mm yr<sup>-1</sup> for temperate valley glaciers overlying crystalline or diverse bedrock (Hallet et al., 1996). We interpret that the larger-scale overdeepenings, distal to the mountain ridge (Fig. 3a), reflect a chronological continuation of ice expansion, rather than the product of a palimpsest of ice margin oscillations because there appears to be little overprinting of scale in the valleys, as might be expected if they were the result of interglacial retreat phases. These larger-scale glacial features may have been eroded into the fluvial topography during warm-based ice sheet expansion any time before the earliest Oligocene, when ice is recorded at the palaeo-shelf break (Hambrey et al., 1991; Zachos et al., 1992). Offshore geological records favour a rapid two step (33.9 and 33.7 Ma) expansion to a near continental-scale ice sheet within  $\sim 300$  Kyr of the E–O boundary (Coxall et al., 2005; Coxall and Wilson, 2011; Hambrey et al., 1991; Lear et al., 2000; Scher et al., 2011; Wilson et al., 2012). We, therefore, propose that the alpine landscape formed rapidly prior to 33.7 Ma and infer a minimum estimate for the scale of the EAIS between ca. 34 and 14 Ma (Fig. 6b).

Our interpretation is consistent with geochemical proxies (Miller et al., 1987) and sedimentary records (Barrett, 1989; Hambrey et al., 1991; Zachos et al., 1992) that indicate the presence of a continental-scale EAIS at the coast in the earliest Oligocene (Supplementary Fig. S1). Estimated ice volumes for East Antarctica record a significant ( $\sim 55$  m) drop in sea level (Zachos et al., 1992; Barrett, 1999) at 34 Ma that corresponds with the formation of this continental-scale ice sheet. Subsequently, during the EAIS fluctuations of the Oligocene–Miocene (Cramer et al., 2011), even minimum ice volume estimates (equivalent to  $\sim 24$  m sea level) ensure that high elevation areas, such as the Gamburtsevs, are likely to have remained fully glaciated. The retention of mountain-scale ice cover (Fig. 6b) also agrees with ice sheet models, which require a climate that is around 15 °C warmer than today to significantly deglaciate the Gamburtsevs (Huybrechts, 1993; Pollard and DeConto, 2005). Such peak warmth has not been documented since the E–O boundary (Zachos et al., 2001).

The survival of the alpine mountain landscape suggests that once ice expanded to encompass the mountains it formed a largely



**Fig. 6.** Maps detailing early ice sheet development atop of the Gamburtsevs, during two different phases of ice expansion. (a) Onset of glaciation at, or prior to, ca. 34 Ma, characterised by the formation of cirque glaciers and high elevation, localised, independent ice caps. (b) Mountain-scale glaciation, resulting from the coalescence of ice caps prior to 33.7 Ma (e.g. Coxall et al., 2005), characterised by a high elevation ice field drained by significant fast flowing glaciers that are steered along the pre-existing fluvial valley networks. Yellow arrows mark the flow direction of outlet glaciers on the northern flank of the Gamburtsevs, which reached the coast of Antarctica via the Lambert Rift at 33.7 Ma.

cold-based core that prevented significant erosion, thereby preserving the landscape since at least 33.7 Ma (Fig. 6b). The evidence we present for erosive continental-scale ice flow (Fig. 1; Supplementary Fig. S3b) is restricted to the northern margin of the Gamburtsevs in the region of the Lambert Rift, where exposed Neogene glacial marine sediments record dynamic EAIS behaviour (Hambrey and McKelvey, 2000a, 2000b). We propose that stable, cold-based ice cover across the Gamburtsevs coexisted with a warm-based fluctuating ice margin, in areas such as the Lambert Rift (Hambrey and McKelvey, 2000a, 2000b) and the Aurora Subglacial Basin (Young et al., 2011; Fig. 1c), consistent with ice sheet model predictions (Jamieson et al., 2008, 2010).

## 9. Conclusions

We used an extensive airborne radar dataset, derived from the AGAP survey, to compile a new DEM for the largely unexplored Gamburtsev Province. By combining quantitative morphometric analyses with geomorphic mapping we aimed to understand the long-term landscape evolution of the Gamburtsevs and to elucidate the nature of early East Antarctic ice dynamics. We show that (i) a remarkably well-preserved ancient fluvial landscape survives beneath the EAIS, confirming that the mountains were already elevated prior to the onset of widespread Antarctic glaciation at ca. 34 Ma; (ii) the fluvial landscape has been glacially modified at a range of scales and these glacial geomorphic features are used to constrain early ice growth and dynamics; (iii) high elevation cirques are interpreted to mark the presence of ephemeral ice masses at, or prior to, the onset of widespread East Antarctic glaciation at 34 Ma; and (iv) the alpine landscape of the Gamburtsevs may have formed rapidly, prior to 33.7 Ma, allowing us to infer a minimum estimate for the scale of the EAIS between ca. 34

and 14 Ma. Crucially, our analyses show that the Gamburtsevs record the earliest evidence for ice growth in East Antarctica.

Our interpretations, and the newly acquired topography of the Gamburtsevs, can be used as inputs to constrain numerical models of ice sheet and climate interaction that seek to understand the growth and stability of the EAIS. This work advances our understanding of the patterns of landscape evolution in the interior of East Antarctica, allowing us to reassess the key role of the Gamburtsevs in relation to long-term Antarctic climate and ice sheet evolution.

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## Appendix A. Supporting information

Supplementary data associated with this article can be found in the online version at <http://dx.doi.org/10.1016/j.epsl.2013.03.053>.

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