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RESEARCH ARTICLE

Landslides in the Transantarctic Mountains: lower Jurassic and older strata displaced in late Mesozoic to late Cenozoic time

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ABSTRACT
Eight landslides, seven certain and one possible, have been identified in the central Transantarctic Mountains and one at Carapace Nunatak, south Victoria Land. Four consist of Kirkpatrick Basalt lavas alone, two comprise Kirkpatrick lavas with associated pyroclastic rocks, one consists of Hanson Formation beds and Kirkpatrick lavas, and one involves Fremouw Formation strata. One possible block, of uncertain origin, consists only of Hanson Formation beds. All rocks comprising the displaced blocks, except one, are Early Jurassic in age. The exception is the inferred slide involving the Triassic Fremouw beds. The locations of some landslides are consistent with emplacement on present-day topography, which has been little modified since the middle Miocene, but the time of emplacement of others is either Oligocene to pre-middle Miocene or pre-dates the onset of glaciation in Eocene/Oligocene time. The older landslides reflect fortuitous preservation of an ancient landscape not unlike that of today, one dominated by horizontal beds consisting of resistant dolerite sills and quartz-rich sandstones alternating with intervals of weak fine-grained sedimentary beds, and capped by basalt lavas. The landslides are interpreted to document three stages in landscape evolution: a pre-glaciation semi-arid landscape, an early warm-based glacial environment, and a late cold-based glacial setting.

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Introduction

Landslides are widespread in mountainous regions involved in active tectonics and have been intensively investigated (e.g. Korup et al. 2007; Hewitt et al. 2008). They also occur in other regions unaffected by the on-going tectonism of mountain belts such as volcanic arcs and terrains, an example of which is the giant Markagunt gravity slide in Utah (Hacker et al. 2014; Biek et al. 2019) that resulted from the collapse of a volcanic edifice. Smaller slides are present where erosional processes have led to steep topography such as the Trotternish Escarpment on the Island of Skye, Scotland (Ballantyne 1990) and the Wasatch Range, Utah (Meyer and Harris 2020). Steep and locally vertical landforms occur across much of the Colorado Plateau and give rise to its characteristic 'mesa and butte' topography, and to distinctive, back-rotated, fallen blocks of coherent strata (Watkins et al. 2004; Watkins and Rogers 2007; Rogers and Watkins 2007, 2008) for which Reiche (1937) coined the term 'toreva block'. The obvious topographic resemblance of the Dry Valleys region of south Victoria Land (Transantarctic Mountains; Figure 1) to the Colorado Plateau of the southwestern United States has been noted by Denton et al. (1993) and others. Landslides occur in many different topographic settings and the drivers of the mass movements include over-steepened topography, bedrock characteristics, rock structure, and climate, in addition to triggering mechanisms such as seismicity and tectonism.

Landscape evolution and the history of glaciation in Antarctica are strongly linked. Investigations in the Dry Valleys region of south Victoria Land demonstrated the antiquity of the landscape (Sugden et al. 1995; Sugden, Summerfield, et al. 1999), as suggested by the age of volcanic cones in the valleys (Wilch et al. 1993) and palaeontological dating of beach and fjord deposits (Webb 1972). Studies in glacial geology have been the primary information source, but one long recognised impediment has been the difficulty in accurately dating deposits and surfaces. Nevertheless, radiometric dating of volcanic ash has established the basic framework that the landscape has been little modified since the mid-Miocene (Lewis et al. 2008). The advent of cosmogenic exposure-age dating of surfaces (e.g. Bruno et al. 1997; Ackert and Kurz 2004; Dickinson et al. 2012) has provided additional constraints but an understanding of the evolution of the Transantarctic Mountains via that technique is still limited to post-mid Miocene time. Earlier glacial history, apart from evidence in marine cores, has remained uncertain because of the lack of terrestrial geological data, although it has been addressed indirectly through interpretation of marine drill core sequences (McKay et al. 2016), and multi-proxy data

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This article has been corrected with minor changes. These changes do not impact the academic content of the article.

sets and modelling (Wilson et al. 2012; Paxman, Jamieson, Ferracioli, et al. 2019a; Paxman, Jamieson, Hochmuth, et al. 2019b). Nevertheless, the glacigenic Sirius Group strata (Barrett 2013) are demonstrably associated with warm-based ice, which is strongly erosional in comparison to ice frozen to its bed (cold-based ice), and are Oligocene/Miocene in age at Mt. Sirius (elevation 2299 m) west of the Queen Alexandra Range (Harwood 1986). Sirius Group deposits are widespread in the Transantarctic Mountains (Denton et al. 1993) and occur in a variety of settings, some of which are unrelated to present-day topography. Given the variety of topographic settings, they almost certainly represent deposition over a long time-span, with deposition beginning in the early stages of warm-based glaciation in Late Eocene/Early Oligocene time.

These lines of investigation address the glacial and tectonic history, but not the actual topography prior to the onset of cold-based glaciation in the mid-Miocene. Landslides, although seldom preserved in the geological column, are another potential avenue of research into terrain evolution. There are obvious challenges in directly relating any landslide, possibly tens of millions of years old, to the terrain at the time of its formation, but the relative simplicity of the inferred long-standing mesa and butte topography of the Transantarctic Mountains offers a possible opportunity. Six landslides (Table 1), two of which must pre-date glaciation, have been identified in the Marshall Mountains, Queen Alexandra Range, one at the Otway Massif, and one at Carapace Nunatak in southern Victoria Land (Figure 1). An additional one at the Otway Massif may have another explanation. These landslides are described briefly and the possible ages of emplacement assessed.

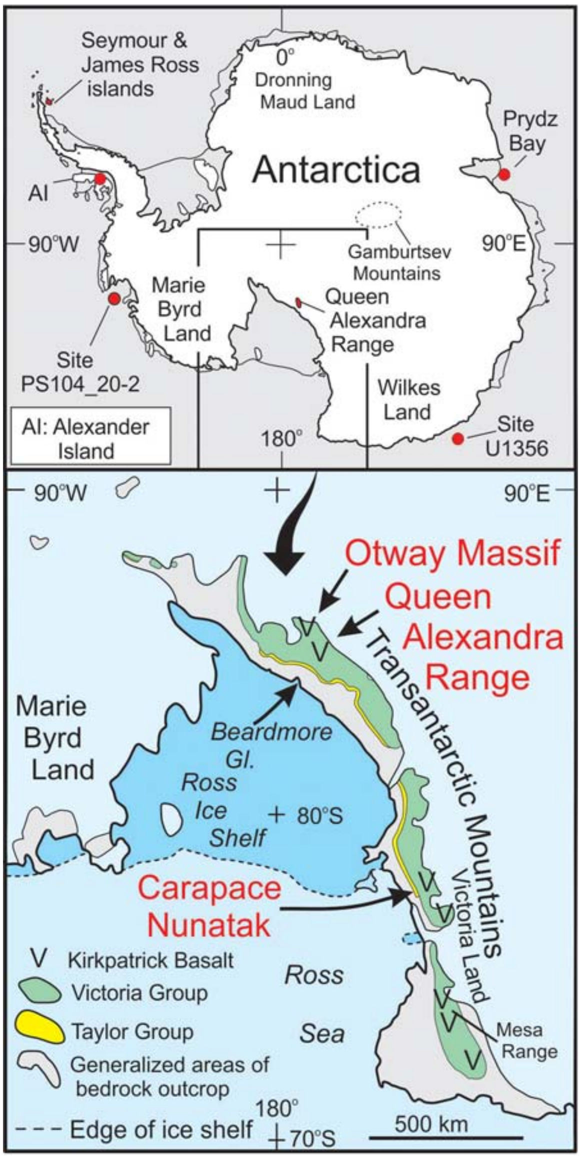

Figure 1. Location map for Antarctica. The Devonian Taylor Group and Permian-Triassic Victoria Group comprise the Beacon Supergroup.

Geological and landscape background

The central Transantarctic Mountains consist of a pre-Devonian basement complex of intrusive and extrusive igneous rocks and low-grade metasedimentary strata (Stump 1995). The basement is overlain by an essentially flat-lying Devonian to Triassic siliciclastic succession (Beacon Supergroup; Barrett et al. 1986; Barrett 1991) (Figure 2) and then by Lower Jurassic volcanic rocks which include silicic tuffaceous strata (Hanson Formation; Elliot et al. 2017) followed by basaltic phreatomagmatic deposits and flood lavas of the Ferrar Large Igneous Province (Elliot et al. 2021). Ferrar sills are widely distributed in the Beacon strata and form highly resistant intervals each commonly more than 100 m thick (Elliot and Fleming 2021).

Most of the landslide deposits consist of intact Lower Jurassic Kirkpatrick Basalt successions, but two include phreatomagmatic deposits which occur below the lavas (Prebble Formation) or within the lava sequence, one consists of Hanson Formation beds, and one involves Triassic Fremouw Formation strata. One landslide (Mt. Falla) has not been visited because, at the time of observation, landscape evolution was not a focus of the field research. Another possible landslide (Barnes Peak) was noted only on aerial photographs.

The uplift/denudation history of the Transantarctic Mountains is documented by fission-track dating of the pre-Devonian (upper Neoproterozoic to Lower Ordovician) granitoids of the Ross Orogen (Fitzgerald 1994, 2002; Fitzgerald and Stump 1997). These data suggest slow uplift/denudation in Cretaceous to Palaeocene time, and onset of rapid uplift in the early Eocene (55 Ma). There appears to have been only very slow uplift in the last 20 m.y. in the central Transantarctic Mountains (Miller et al. 2010), and in southern Victoria Land (Summerfield et al. 1999). It

Table 1. Locations and details of certain and possible landslides.

Location	Strat. thick.	Displacement[a] H	Displacement[a] V	Rock units[b]	Lower contact[c]	Attitude	Inferred age
Mt. Falla (QAR)	300+m	2 km	300 m	Jk	Trfa	~65°SE	Early Eocene or older
Buttress Peak (QAR)	245 m	6 km	400 m	Jk	Trfa	~40°SE	Early Eocene or older
Barnes Peak (QAR)	200+m	unknown	unknown	Trfr	Pb	10-15°S	Early Eocene or older
Mt. Stonehouse (QAR)	25+m?	7 km	1000 m	Jk	Trfa	~25°NW	Oligocene to mid-Miocene
E of Storm Peak (QAR)	80+m	2 km	300 m	Jh, Jk	Trfa	15°S	Oligocene to mid-Miocene
Elliot Peak (QAR)	25+m?	0.1 km	300 m	Jk	Jh/Jp	~20°N	Post-mid-Miocene
Mt. Spohn (OM)	100+m	0.2 km	100+m	Jp, Jk	Jp	near horizontal	Post-mid-Miocene
NW of Mt. Petlock (OM)	55+m	uncertain	uncertain	Jh	unknown	19°S	
Carapace N. (SVL)	10+m	0.1 km	100+m	Jm, Jk	Jc	18°S	Post-mid-Miocene

Notes: Strat. thick.: stratigraphic thickness of the beds in each landslide. N: Nunatak. OM: Otway Massif. QAR: Queen Alexandra Range. SVL: South Victoria Land. Pb: Permian Buckley Fm. Trfr: Triassic Fremouw Fm. Trfa: Triassic Falla Fm. Jh: Jurassic Hanson Fm. Jp: Jurassic Prebble Fm. Jm: Mawson Fm lithologies interbedded in lower part of Kirkpatrick lava succession (Bradshaw 1987). Jc: Carapace Sandstone. Jk: Jurassic Kirkpatrick Basalt. The uncertain landslide is in italics.

[a]Minimum horizontal (H) and vertical (V) displacement, given present day distribution of rock formations and assuming derivation from the closest outcrops.

[b]Rock units in landslide blocks.

[c]Rock units on which landslides rest.

should be noted that the timing and rates of uplift/denudation differed between the discrete blocks into which the range is divided (Fitzgerald 2002). The conclusions of a more recent evaluation of the data suggest that a thick sedimentary succession was deposited on the Kirkpatrick Basalt and the adjacent regions in the Ross Sea sector, forming the Mesozoic Victoria Basin of Lisker and Laufer (2013), which has additional implications for the evolution of the landscape. It should be noted that the notion of the Mesozoic Victoria Basin is at odds with the proposal that West Antarctica was an elevated plateau during the Cretaceous (Bialas et al. 2007; Deccesari et al. 2007).

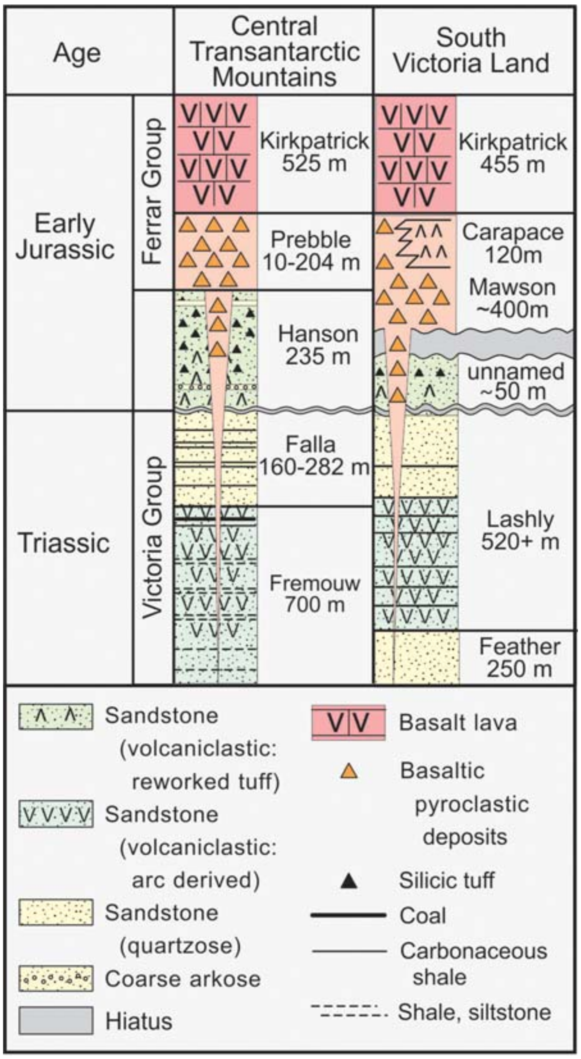

Figure 2. Simplified stratigraphic columns for Triassic and Jurassic strata in the central Transantarctic Mountains and south Victoria Land.

Major glaciation was initiated at about 34 Ma, close to the Eocene/Oligocene boundary, with warm-based conditions persisting until about 14 Ma when cold-based conditions were established (Barrett 1996; McKay et al. 2016). The warm-based glaciation must have modified pre-existing fluvially developed topography, which, given the layer-cake stratigraphy with highly resistant dolerite sills and less resistant sedimentary strata, had resulted in a mesa and butte landscape. This landscape was gently tilted toward the Polar Plateau by the Cenozoic uplift of the mountains. Rates of erosion, following the establishment of cold-based glaciation, must have decreased significantly. For the Dry Valleys region of south Victoria Land rates are estimated to be very slow, possibly 1–6 cm/m.y. (Dickinson et al. 2012). Thus, landslide blocks apparently resting on present-day topography may be any age between mid-Miocene and today.

The pre-glacial landscape of Antarctica is a matter of much interest because of its bearing on the development of the ice sheets. Although the earliest glaciation may have occurred in the Gamburtsev Mountains (Bo et al. 2009) and may be older than Eocene/Oligocene (Wilson et al. 2013), only the Transantarctic Mountains landscape is considered here. That landscape has been discussed by analogy with the geomorphological evolution of the adjacent Gondwana continents, with emphasis on the escarpments, coastal plains and continental drainage patterns, which resulted from the break-up of that supercontinent (Jamieson and Sugden 2008; Sugden and Jamieson 2018). Another avenue of research into the landscape, but limited to the Eocene-Oligocene boundary and younger times, has been the reconstruction of Antarctic bedrock topography using factors such as erosion, sedimentation, ice sheet loading, and isostatic responses (Wilson et al. 2012; Paxman, Jamieson, Ferracioli, et al. 2019a; Paxman, Jamieson, Hochmuth, et al. 2019b).

Site descriptions

Queen Alexandra Range

Mt. Falla. The landslide (observed but not examined in outcrop) occurs on a ridge about 4.5 km east-northeast of Mt. Falla were more than 300 m of basalt lavas, dipping at about 65° SE, crop out (Figure 3). There is no exposed contact with underlying strata on the west side of the ridge; the east side is not clearly visible in vertical photography (Polar Geospatial Centre/USGS photo TMA 2756 022 V), but appears to be covered by patterned ground and surficial debris. The nearest comparable outcrop of *in situ* basalt occurs at a small outcrop 2 km east of Mt. Falla, which now consists of no more than about 100 m of lavas; a full (475 m) basalt succession occurs at Mt. Falla itself. The strata trending into the landslide consist of Lower Jurassic Prebble Formation beds, Hanson Formation beds, and possibly the uppermost part of the Triassic Falla Formation (Figure 4). A minimum of at least 775 m of relief due to erosion through the Kirkpatrick Basalt down to the upper contact of the Falla Formation (475 m of lavas, 55 m of Prebble beds, 245 m of Hanson beds at Mt. Falla) had to have existed prior to landslide emplacement. It is possible that the original extent of the landslide was greater and a basal part, with unknown stratigraphic thickness, rested on eroded Falla Formation beds. Should that be the case, then that basal part of the Kirkpatrick lava landslide succession was removed after emplacement and during further landscape evolution.

Given the position of the slide, it is most probable that it occurred at a time when a steep slope or cliff consisting of those formations plus the Kirkpatrick Basalt was located adjacent to and approximately southwest of the slide, assuming the strike of the Kirkpatrick lavas in the slide block is perpendicular to the transport direction. Subsequently, the cliff or slope retreated as sub-aerial erosion processes created the ridge on which the extant slide is now located. Surficial debris was deposited later on the east slope of the ridge, and subsequently transformed into patterned ground. The northwest flank of today's ridge falls

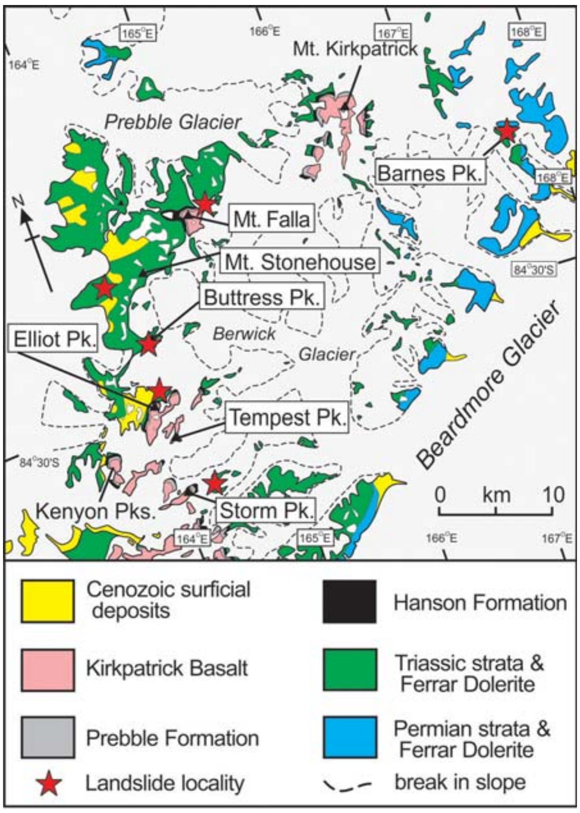

Figure 3. Simplified geologic map for the Marshall Mountains, Queen Alexandra Range.

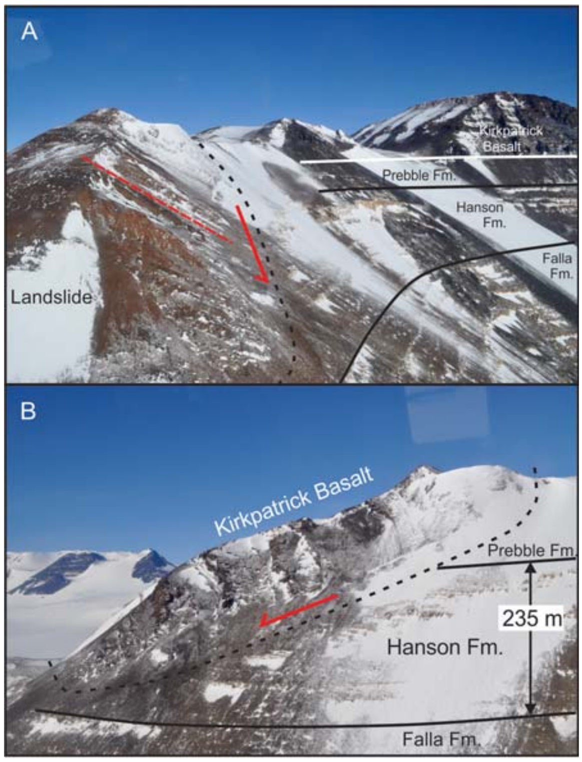

Figure 4. Coherent block of Kirkpatrick Basalt resting on a surface cut into Prebble, Hanson and possibly uppermost Falla Formation strata. A: View from the north to illustrate the relationship to Mt. Falla (summit at far right). B: View from the west along strike of the Kirkpatrick lava landslide. In A the red dashed line indicates bedding in the Kirkpatrick Basalt lavas forming the landslide. In A and B the short dashed lines mark the approximate contact of the landslide with bedrock. Red arrows indicate the sense of motion of the block. The landslide is located approximately 4.5 km east-northeast of Mt. Falla.

more than 1000 m to the Prebble Glacier, whereas the southeast flank merges into the high ground that connects Mt. Falla to Mt. Kirkpatrick.

Buttress Peak. The exposed rock on the north side of Buttress Peak (Figure 3; elevation ∼2950 m) consists of Triassic and Jurassic strata, whereas a 245-m-thick measured section of Kirkpatrick Basalt is exposed on the southeast side and on the southwest side only talus and surficial debris are present (Figures 5 and 6). The attitude (dip and strike) of the lavas is problematic; a dip measurement of ∼40° to the southeast was obtained from the uppermost lava outcrop, although there is uncertainty as to whether that attitude correctly represents the whole of that measured section. The nearest *in situ* Kirkpatrick Basalt, with a lower contact elevation at about 2835 m, occurs at Elliot Peak more than 5 km to the south. The lowest thick flow at Buttress Peak is correlated with the lowest flow at Elliot Peak, both of which are distinctive, thick, black glassy lavas, although the lower third at Buttress Peak is unexposed or absent (Barrett et al. 1986, Plate 1c; sections 40 and 36).

A sill with the upper contact at an elevation of about 2400 m (elevation obtained from the Reference Elevation Model of Antarctica or REMA; pgc.umn.edu), which separates the older Triassic Fremouw Formation from the younger Falla Formation, provides a somewhat imprecise elevation marker connecting the two localities (Figure 7). The stratigraphic succession at Buttress Peak is about 550 m thick. This unmeasured succession consists of Falla and Hanson beds, and a sample from 20 m below the summit is tuffaceous and confirms the stratigraphy.

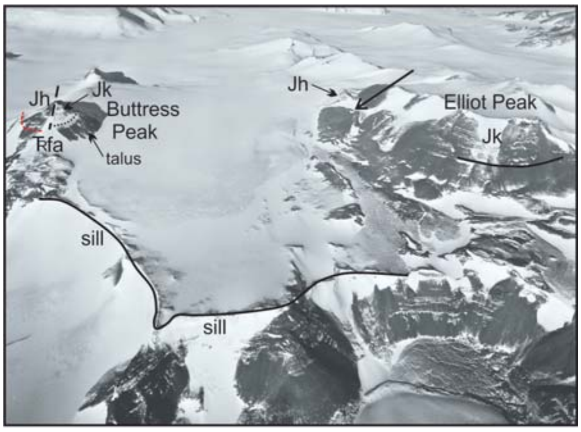

Figure 5. Aerial view (Polar Geospatial Centre/USGS air photo TMA 765 312 F31) of the Buttress Peak-Elliot Peak region. Trfa: Triassic Falla Formation; Jh: Jurassic Hanson Formation; Jk: Jurassic Kirkpatrick Basalt (in shadow at Buttress Peak). Dashed black line marks the approximate position of the contact between the Kirkpatrick Basalt landslide and Jurassic strata. Dashed red line marks the approximate position of the contact between the Falla and Hanson formations. Arrow near Elliot Peak points to the probable landslide illustrated in Fig. 10.

The measured section of the Kirkpatrick Basalt lavas at Elliot Peak is 435 m (Barrett et al. 1986), and the base of the lavas lies at 2835 m (obtained from the REMA data for Elliot Peak). On the west and northwest faces of Elliot Peak exposures of the Prebble and Hanson beds are very sparse (Barrett and Elliot 1973; note that the volcanic upper part of the Falla Formation on this map was separated off as the Hanson Formation by Elliot 1996). In contrast to Buttress Peak, the combined thickness of the Falla, Hanson and Prebble formations at Elliot Peak is about 450 m. This consists of 45 m of Prebble beds measured at Elliot Peak, possibly 210 m of Hanson beds based on a measured section northeast of Elliot Peak (Elliot, unpublished data; Elliot et al. 2017, site 79; in error, it was identified in that text as station 39), and the remainder (ca. 195 m) therefore is assigned to the Falla Formation (the nearest measured section at Mt. Falla is 285 m thick). The reduction in the combined stratigraphic thickness of the Falla, Hanson and Prebble beds suggests that the Falla Formation thins from Buttress Peak to Elliot Peak.

There are two end-member scenarios for the emplacement of the landslide. First, if the slide occurred on present-day or recent topography, it moved about 6 km horizontally to the north and dropped about 200 m. Movement could have been over an ice surface if, at an earlier stage of glaciation,

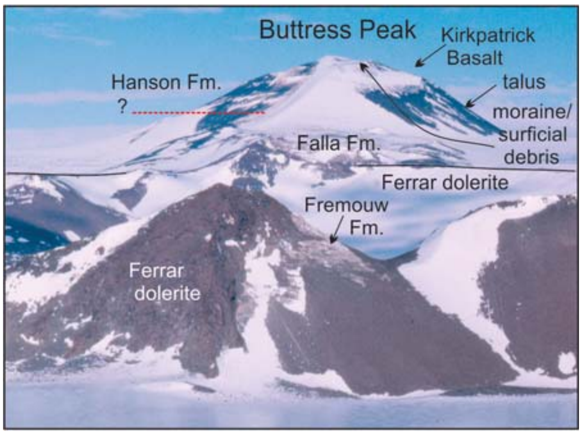

Figure 6. Buttress Peak (summit elevation 2950 m) viewed from the west. Falla, Hanson and possibly Prebble formation beds are inferred to crop out on the north flank of the peak, and, on the southeastern flank (hidden from view), exposures consist of an intact, measured, but poorly exposed 245-m-thick section of Kirkpatrick Basalt. The attitude of these lavas is uncertain; a dip of about 40° to the southeast was obtained on the top lava, but is suspect. Beds on the upper southwest flank appear to dip gently to the north, but examination of the vertical photography (Polar Geospatial Centre/USGS air photo CA275300VOO72) shows only surficial debris and patterned ground above the talus; on that vertical photo the lava section is in deep shadow. The dashed red line on the west flank of Buttress Peak marks the approximate location of the inferred contact between the Falla Formation and Jurassic beds (Hanson strata and possibly Prebble beds).

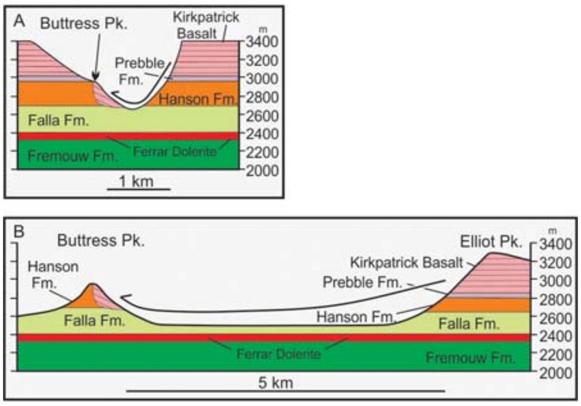

Figure 7. Schematic cross-sections for the slide at Buttress Peak. A. Early emplacement of the slide, after initial valley cutting and before glaciation in Eocene/Oligocene time. B. Late emplacement on post-mid Miocene topography. Arrow indicates the sense of motion of the Kirkpatrick Basalt block from Elliot Peak to Buttress Peak.

ice and snow covered the present-day low elevation terrain between Buttress Peak and Elliot Peak. The occurrence of an 87-m-thick till capping Mt. Sirius (elevation 2299 m) about 38 km north-northwest of Buttress Peak, which includes diatoms of Oligocene to early Miocene age (Harwood 1986, p. 124), makes this a not unreasonable inference. Further, Prentice et al. (1986) and Denton et al. (1991, 1993) reported Sirius-type warm-based till occurring between 3490 and 3825 m on Mt. Falla and between 3170 and 4015 m on Mt. Kirkpatrick. The Sirius-type till is older than mid-Miocene (Lewis et al. 2007, 2008) and demonstrates a warm-based ice-cover much thicker than that of today. In the other end-member scenario, erosion produced relief of ca. 750 m (500 m of lavas at the Elliot Peak-Tempest Peak massif plus, from Elliot Peak, ca. 45 m of Prebble beds and ca. 210 m of Hanson beds, and assuming erosion did not cut into Falla strata at Buttress Peak) and formed a narrow valley between what later became Buttress Peak and a steep slope or cliff to the immediate southeast. In this scenario, the landslide has the characteristics of a toreva block (Reiche 1937) in which strata are back-rotated and dip towards their source. A scenario between the two end-members is possible, but the second end-member is considered more probable because of the stratigraphic thickness (245 m) of the apparently intact slide block of Kirkpatrick lavas.

Barnes Peak. Observations from afar and from Polar Geospatial Centre/USGS air photos (TMA 776 0281 F33; TMA 775 0168 F31) suggest that there is an angular discordance between the Fremouw Formation and the underlying Permian Buckley Formation (Figure 8) at Barnes Peak (Figure 3). The layer-cake stratigraphy of the Beacon strata in the Queen Alexandra Range precludes the apparent discordance being an angular unconformity. Thus, the difference in attitude is tentatively interpreted as the Fremouw strata forming a large toreva block, down to the south, which moved on fine-grained beds in the uppermost Buckley Formation.

Mt. Stonehouse. Black Kirkpatrick Basalt lava, mainly rubble but locally with apparent bedding, overlies Fremouw Formation strata (Figure 9) (TMA 2755 0162 V; TMA 765 0307 F31) at Mt. Stonehouse (Figure 3). It is not known if surficial debris separates the lava rubble from the Fremouw beds. The debris consists of black glassy lava similar to the second flow at Storm Peak (Elliot and Fleming 2008), to the lowest flow on the northeast ridge of Mt. Falla and to the lowest flow at Elliot Peak (Barrett et al. 1986). Blocks of wood-bearing silicified sedimentary strata, similar to blocks found on the Kirkpatrick lavas on the east-facing slope of Kenyon Peaks (Figure 3), are also present. The nearest *in situ* black glassy lava outcrop is 8 km distant at the southern end of Mt. Falla where the base of the lavas lies at about 3400 m, more than 1000 m higher. This implies that topography, at least somewhat similar to that of today, had formed by the time of the landslide. Further, the apparent bedding (low dip to the southeast) suggests it is not a young (late Cenozoic) vent extruding a

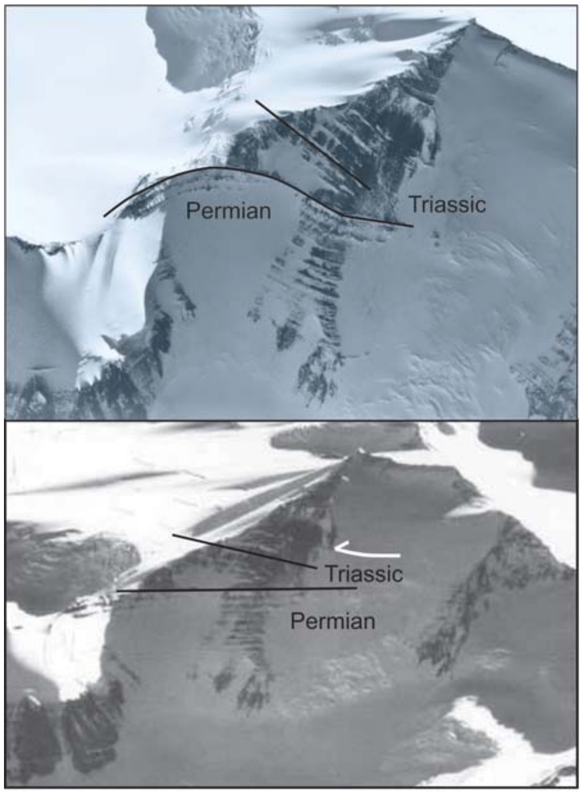

Figure 8. Polar Geospatial Centre/USGS air photos TMA 776 0281 F33 (upper photo) and TMA 775 0168 F31 (lower photo) cropped and enlarged to illustrate the Barnes Peak locality. Lines are parallel to bedding and illustrate apparent discordance between the Permian Buckley Formation and the Triassic Fremouw Formation. White arrow indicates the inferred sense of motion. View from the east.

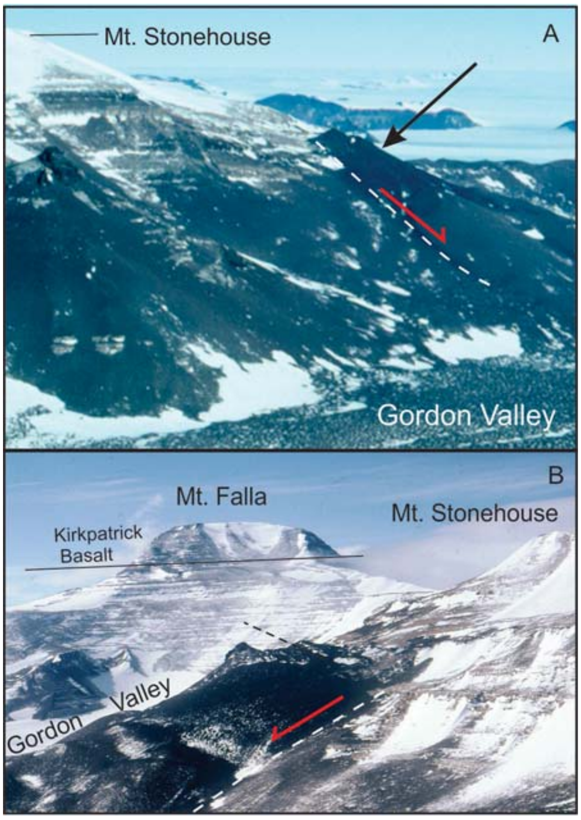

Figure 9. A possible landslide consisting mainly of black Kirkpatrick Basalt rubble on the slope below Mt. Stonehouse. The dashed white lines mark the approximate position of the base of the slide. The red arrows indicate the sense of motion. A: view from northeast, with arrow indicating the landslide. B: view from the west-southwest, with black dashed line indicating apparent bedding (dip ca. 25° to the southeast) in the uppermost part of the landslide. The nearest *in situ* Kirkpatrick Basalt occurs at Mt. Falla 8 km to the east-northeast. See also Polar Geospatial Centre/USGS air photos TMA 765 0307 F31 and TMA 2755 162 V.

glassy flow that fractured, broke apart, and spread rubble down slope.

Elliot Peak. On the northern slope of Elliot Peak (Figure 3) there is an outcrop of broken basalt lava

Figure 10. Kirkpatrick Basalt overlying talus northeast of Elliot Peak (summit is beyond the image to the right). Long dashed black line indicates the probable attitude of the landslide block. Short dashed line marks the base of the landslide. Red arrow indicates the sense of motion. Jh: Hanson Formation strata (see Fig. 5). View toward the northeast.

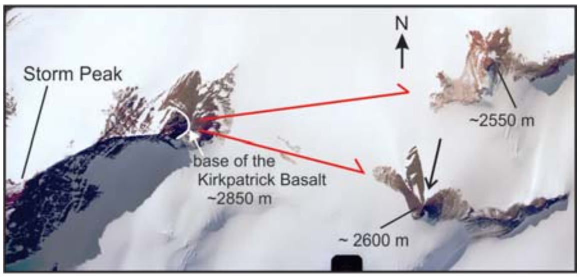

Figure 11. Polar Geospatial Centre/USGS air photo TMA 2750 215 V cropped and annotated to illustrate the landslide blocks east of Storm Peak. The estimated elevation of the base of the basalts at Storm Peak (2850 m) and the summits of the two displaced blocks (2550 and 2600 m) have been obtained from the REMA data. The large black arrow points to the block examined. Red arrows indicate the sense of displacement. Note that the beds at the base of Storm Peak mapped as the Falla Formation by Barrett and Elliot (1973) would now be mapped as the Hanson Formation (see Elliot et al. 2017).

overlying surficial material that itself lies on Hanson and Prebble beds (Figure 10; see also TMA 2752 046 V). There is a suggestion of stratification and a dip parallel to the land surface. Apparently resting on the present-day surface, this slide block is post-mid Miocene.

Storm Peak. Two small isolated nunataks occur about 2 km east of Storm Peak (Figure 3 and Figure 11; TMA 2750 215 V). The southerly one comprises Hanson and Kirkpatrick beds. The Hanson beds, on the north side of that nunatak, are tuffaceous sandstones with a thickness of 10s of m and dip about 15° to the south. The poorly exposed Kirkpatrick strata on the south flank include silicified wood, a silicified tree stump, and blocks of conchostracan-bearing sedimentary beds. The latter probably come from the interbed that lies above the second lava flow in the Storm Peak succession (see Elliot and Fleming 2008), rather than the interbed below the capping lava. The northerly outcrop consists of an uncertain but small thickness of Kirkpatrick Basalt. The landslide blocks have dropped a minimum of about 250 m and in the case of the fossiliferous beds a minimum of about 388 m. The setting of both blocks requires erosion down through the Hanson Formation and possibly into the upper Falla Formation (Elliot 1996; Elliot et al. 2017). The landslides could have been proximal at the time of emplacement and later cliff retreat has separated them from *in situ* Kirkpatrick Basalt. Less probably, they could have been emplaced more recently and were transported over snow to their current locations.

Otway Massif

Mt. Spohn. About 2.5 km south-southwest of Mt. Spohn (Figure 12), there are two isolated outcrops of

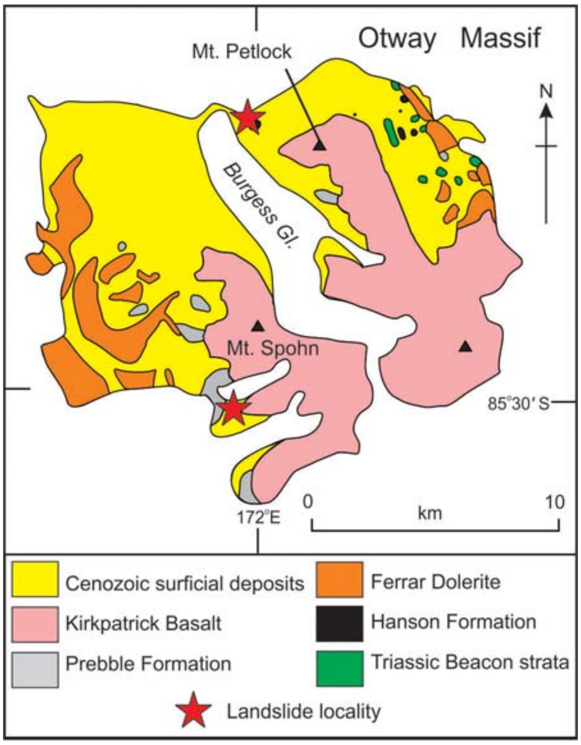

Figure 12. Simplified geologic map for the Otway Massif.

Kirkpatrick Basalt, both of which are surrounded by surficial debris (Figure 13; TMA 855 358 F33; TMA 2726 046V). One of the outcrops consists of about 100 m of Prebble Formation and Kirkpatrick Basalt. A small thickness of Kirkpatrick Basalt forms the other. Both overlie Prebble Formation tuff breccia, which at the Otway Massif is more than 300 m thick. The slide blocks, resting on present-day topography, must be post-mid Miocene.

Mt. Petlock. A 55-m-thick section of Hanson Formation strata, dipping 19° to the south, occurs about

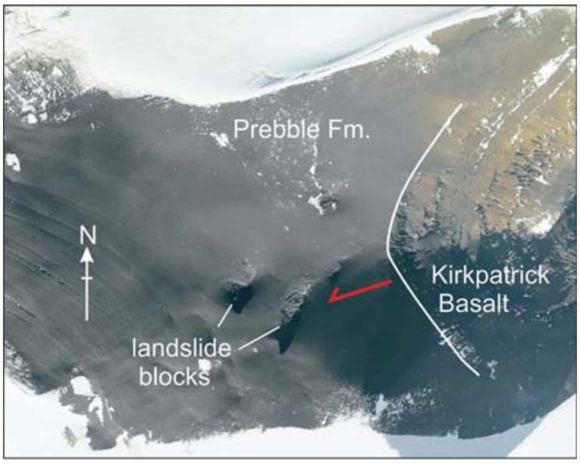

Figure 13. Polar Geospatial Centre/USGS air photo TMA 2726 046 V enlarged and annotated to show the two landslide blocks and the adjacent outcrop of the Prebble Formation. Landslide blocks are below the lowest *in situ* Kirkpatrick Basalt lava. Red arrow indicates the sense of motion. Location is south-southwest of Mt. Spohn (Fig.12). Air photo TMA 855 358 F33 provides a distant view of the site.

Figure 14. Polar Geospatial Centre/USGS air photo TMA 2730 013 V enlarged and annotated to illustrate the Hanson Formation locality west northwest of Mt. Petlock (Fig. 12). Air photo TMA 762 324 F31 provides a distant view of the locality.

2.5 km to the west of Mt. Petlock (Figure 12) and east of the terminus of the Burgess Glacier (Figure 14; TMA 2730 013V; TMA 762 324 F31) at an elevation of 2200–2300 m. The outcrop is entirely surrounded by surficial debris. The base of the Kirkpatrick Basalt is at about 2800 m, and the exposed underlying Prebble Formation most probably is part of a >300-m-thick phreatocauldron filling (Elliot and Hanson 2001), the boundaries of which are unknown because of surficial cover.

If the Hanson stratigraphic succession was originally outside the phreatocauldron, it could have been covered by up to 200 m of stratified Prebble beds (the greatest recorded thickness of Prebble strata in the central Transantarctic Mountains), and if this were correct, the Hanson Formation block was dropped several hundred metres. Relationships to older Triassic beds or the Prebble Formation can only be speculative.

If encompassed by the phreatocauldron, the Hanson beds are simply a megaclast resulting from incorporation into a phreatocauldron fill such as has been documented at Coombs Hills in south Victoria Land where megaclasts of Triassic Lashly Formation foundered into a vast volcanic vent (White and McClintock 2001). Several megaclasts (up to 60 m long and with steep attitudes) of Hanson and/or Falla beds embedded in Prebble pyroclastic debris occur northwest of Mt. Spohn (see Elliot and Hanson 2001, Figure 3A, location 96-24; Elliot and Fleming 2008) and support this possibility.

The age of emplacement of the 55-m-thick Hanson succession is also speculative: Early Jurassic in age and contemporaneous with the formation of the phreatocauldron, or a landslide that was pre-glacial (pre-Eocene/Oligocene), or pre-Miocene in age. The simplest, but not definitive, explanation is that the Hanson beds are embedded in the phreatocauldron and the age of emplacement is Early Jurassic.

South Victoria land

Carapace Nunatak. Outcrops of interbedded pyroclastic rocks and lavas on the northeast flank of the Carapace Nunatak (Figure 3) (Bradshaw 1987; Ross et al. 2008) have been mapped as surrounded by faults. Although crustal faults may cross the area, the attitude (dip of 18° S) suggests these beds could be a set of toreva blocks emplaced in post-mid Miocene time (Figure 15).

Late Jurassic to Eocene palaeoclimate

The palaeoclimate of the Queen Alexandra Range is difficult to assess because relevant data are confined to far distant localities (Figure 1). Cretaceous strata in the Antarctic Peninsula region crop out on Alexander Island, about 2600 km distant from the Queen Alexandra Range, and in the James Ross Island basin off the northern Antarctic Peninsula, 3450 km distant. The Alexander Island coal-bearing beds of mid-Cretaceous (late Albian) age indicate a temperate rainforest and a warm humid climate (Falcon-Lang et al. 2001). The James Ross Island Cretaceous strata are all marine, and the palaeoenvironments have been inferred from transported leaves, wood, pollen and spores, and to have been warm and wet (Ditchfield et al. 1994; Poole et al. 2005; Francis et al. 2008).

A marine core has been recovered from a closer site (~1900 km to the Queen Alexandra Range) off eastern Marie Byrd Land (Figure 1, Site PS 104_20-2; Klages et al. 2020). The mid-Cretaceous part of the core includes a 3-meter-thick section consisting of laminated carbonaceous mudstone with a network of fossil roots and containing abundant plant debris. Pollen and spores indicate a temperate rainforest and provide an age of Santonian-Turonian (92-83 Ma). The estimated palaeolatitude of this mid-Cretaceous site is 82°S. Pollen and spores in the lower part of a core recovered at Prydz Bay (Figure 1; Truswell and Macphail 2009) about 2500 km distant from the Queen Alexandra Range suggest a late Cretaceous (Turonian) age and a flora dominated by conifers. The vegetation points to a temperate climate with moderate humidity.

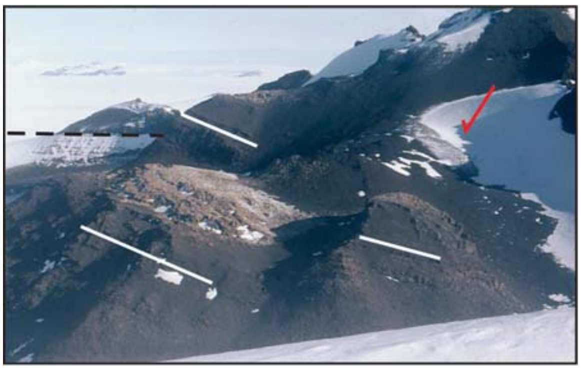

Figure 15. Carapace Nunatak, south Victoria Land. The attitude of the Kirkpatrick Basalt lavas and interbedded pyroclastic rocks (tuff breccias and lapilli tuffs) is indicated by solid white lines. The sense of motion is indicated by the red arrow. Near horizontal bedding in the Carapace Sandstone is shown by the dashed black line (centre left). Kirkpatrick lavas are conformable on the Carapace beds. View to the southeast. The summit of Carapace Nunatak lies to the right of the photo.

Eocene strata crop out on Seymour Island (Marenssi 2006), are present in a marine core recovered off the Wilkes Land coast (Figure 1, Site U1356; Pross et al. 2012) and also comprise a 120-m-thick section overlying the Cretaceous part of the core recovered at Prydz Bay. The Palaeocene and Eocene floras from Seymour Island are consistent with a cool temperate climate with woodlands dominated by conifers and angiosperms (Francis et al. 2008). Estimates of the palaeotemperatures point to significant cooling by late Eocene time. The Prydz Bay microfloras of middle to late Eocene age suggest a cool to cold temperate rainforest (Truswell and Macphail 2009).

In the Wilkes Land case, a 100-m-thick section of a core contains pollen and spores that demonstrate an early to mid-Eocene age. The early Eocene flora has two components which suggest a proximal low elevation near-tropical rainforest and a cooler temperate rainforest either at higher elevations or distally in the continental interior. The mid-Eocene flora is markedly different and is interpreted to suggest an expansion of the temperate rainforest into the coastal region. The mean annual temperature estimate for the near-tropical flora is 16 ± 5°C, and 14 ± 3°C for the temperate rainforest flora.

The Beardmore Glacier region was located about 1900 km away from the nearest Cretaceous site, which is on the Pacific margin, and about 2350 km away from the nearest Eocene site, one of which was on the developing seaway between Antarctica and Australia (Huber et al. 2004) and the other at Prydz Bay (~2500 km distant). The palaeofloras ringing Antarctica during both time intervals suggest abundant vegetation growing in relatively warm and wet environments. The Beardmore region was far distant in the continental interior at both those times, and its climate was probably typical of environments far removed from moisture sources. With a near polar location, stronger diurnal and seasonal temperature variations were probable, and precipitation was likely to have been much less than the >100 cm/year estimated for the Wilkes Coast Eocene site. Whenever the mesa and butte topography now observed was initiated, it must have been under a semi-arid climate and one very different from those documented coastal localities.

Interpretation

The Mesozoic landscape of Antarctica, and the Transantarctic Mountains in particular, is difficult to assess.

Direct comparison with either southern Africa or South America is problematic because the Transantarctic Mountains do not form a rift shoulder related to rifting and sea-floor spreading. It should be noted that this is the case for Dronning Maud Land (Sugden and Jamieson 2018). The Gondwana margin, which had been a tectonically active margin through much of the Palaeozoic and into the Jurassic, was broken up and reorganised between the start of rifting at about 180 Ma and about 110 Ma when the various continental blocks of West Antarctica achieved their present relative positions (Grunow et al. 1987, 1991). During that time, the Ross Sea sector of the Transantarctic Mountains would have been on the margin of stable East Antarctica and adjacent to the reorganised Palaeozoic and early Mesozoic magmatic belts (Elliot 2013; Elliot et al. 2017). They were not adjacent to a rifted margin that was evolving into an oceanic basin. Therefore, they were unlikely to have had significant elevation.

The youngest Gondwana rocks, the Kirkpatrick Basalt (182.7 Ma; Burgess et al. 2015), might have had an additional cover (up to 1.5 km) of now eroded Lower Jurassic lavas and/or, as suggested by Lisker and Laufer (2013), a Mesozoic sedimentary cover, up to 4 km thick, but now lost by erosion. In the region of interest, the Queen Alexandra Range, the supra-crustal rocks must have formed a layer-cake stratigraphic succession, with or without the proposed Mesozoic sedimentary cover. There is evidence for mid-Cretaceous events in both the fission-track dating of basement granitoids (Fitzgerald 1994) and in secondary mineralisation in the Kirkpatrick lavas (Fleming et al. 1999). A latest Cretaceous/earliest Palaeocene event is also recorded in secondary mineralisation in upper Permian Buckley Formation beds (Elliot et al. 2004). The fission-track data suggest slow uplift or denudation, and the mineralisation suggests disturbances to the groundwater systems. The alternative evolutionary scenarios, one with and the other without a Mesozoic sedimentary cover, differ and will be considered separately.

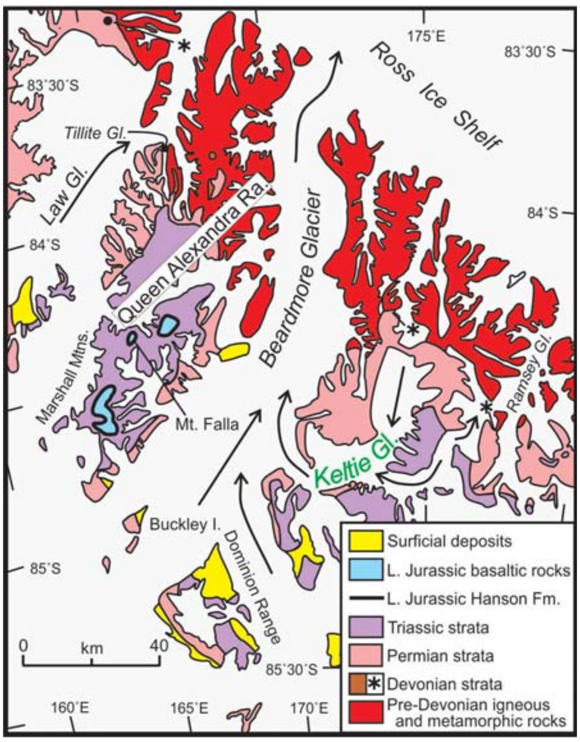

Figure 16. Map of the Beardmore Glacier region to show the regional geology and the location of the Keltie Glacier. Ice flow is indicated by the black arrows.

Assuming no Mesozoic sedimentary cover, slow uplift may have started in the middle to late Jurassic, but no later than the mid-Cretaceous, as suggested by the fission-track data. Erosion of the succession during Cretaceous to Eocene time is assumed to have occurred sub-aerially and by fluvial and other surface processes. The geomorphological effects of erosion would have been strongly influenced by the resistant units in the Gondwana succession, that is the lavas, dolerite sills and quartz-rich sandstones. The development of the characteristic vertical cliffs of resistant rock units and rectilinear slopes of weaker beds could have occurred any time before the late Eocene and must have created relief of at least several hundred metres (e.g. 775 m at Mt. Falla).

The proposal by Lisker and Laufer (2013) of an up to 4-km-thick Mesozoic sedimentary cover leads to a somewhat different interpretation. That cover necessitates a low-lying terrain crossed by rivers transporting debris from one or more elevated terrains, the most probable and proximal one being the sub-glacial Gamburtsev Mountains of possible early Paleozoic origin (Veevers 2018). Cretaceous secondary mineralisation events may have been related to the burial of the Beacon and Ferrar rocks rather than groundwater changes due to slow uplift. Further, the Mesozoic beds, which must have been at most weakly consolidated and therefore easily eroded, had to have been removed no later than during the early Eocene on the initiation of rapid uplift. The mesa and butte topography, therefore, must post-date removal of the Mesozoic beds, which then dictates that it was developed no later than in Early Eocene time.

Assuming a weak easily eroded Mesozoic succession, the Kirkpatrick Basalt formed the uppermost resistant rock unit in the stratigraphic column. The capping lava is a distinct and unique chemical type present at most outcrops (see Elliot et al. 2021; Elliot and Fleming 2021). If there had never been an additional 1.5 km of now eroded lavas, it is just possible that remnants of the Mesozoic sedimentary rocks might be present as residual pockets on the capping lavas, most probably in the extensive Mesa Range in north Victoria Land (Figure 1).

It is proposed that the mesa and butte topography was developed under a semi-arid climate in the interior of the continent in pre-glacial times and no later than the Early Eocene. Initially, it was cut in the lava succession and then progressively with time including the underlying Prebble and Hanson beds and then the fluvial Falla Formation of Triassic age and older strata. The start of rapid uplift/denudation at about 55 Ma (Fitzgerald 1994) would have initiated greater rates of erosion and incision, but the basic landform would not have changed. Perhaps the best evidence for an early episode of erosion, unrelated to the present-day ice drainage patterns, lies in the Keltie Glacier (Figure 16), which flows towards the continental interior before merging with the Beardmore Glacier and flowing out to the Ross Ice Shelf (Huerta 2007).

Present-day topography has been little modified since the advent of cold-based glaciation, which post-dates the middle Miocene (ca.14 Ma; Sugden et al. 1995; Sugden, Summerfield, et al. 1999; Barrett 1996; Lewis et al. 2008). Therefore, the age of the landslide blocks resting on what appears to be modern topography and adjacent to a source (Mt. Spohn, Carapace Nunatak, Elliot Peak) must lie between about 14 Ma and recent. On the east side of the Otway Massif, outsize Kirkpatrick Basalt blocks more than 2 m across were noted lying on the surficial cover (Elliot, unpublished data) and point to the possibility of rockfall and runout of 1 km or more under present-day conditions.

Landslide blocks of intact strata separated by some distance from the nearest potential source (Storm Peak, Mt. Stonehouse, Buttress Peak) indicate, if emplacement was proximal to source, significant cliff retreat since emplacement. If not emplaced proximally, then blocks could have been transported for some distance over snow and ice from cliff fronts at an earlier stage of cliff retreat. In this case, the landslides could have been emplaced during earlier stages of glaciation when ice levels were higher (Prentice et al. 1986; Denton et al. 1991) and the ice was warm-based and much thicker, as demonstrated by the glacial deposits of late Oligocene/early Miocene age at Mt. Sirius (Harwood 1986).

The Buttress Peak landslide could have occurred either before glaciation or during the warm-based stage between its onset at about 34 Ma (Eocene-Oligocene boundary) and the shift to cold-based glaciation at about 14 Ma in the middle Miocene (Shackleton and Kennett 1975; Lewis et al. 2007, 2008). Ice and snow cover with a surface elevation of about 2700 m could have provided a surface of transport. Because the Kirkpatrick Basalt block is intact, it is considered more probable that the slide occurred much earlier in the erosional history of the region when the source was proximal; therefore the landslide is interpreted as a toreva block (*sensu stricto*). If that is correct, the source would have lain to the southeast of Buttress Peak, not in the extant Tempest Peak-Elliot Peak massif. In this case the time of emplacement was after local erosion down to at least the Falla/Hanson boundary and at a time no later than the middle Eocene. The isolation of Buttress Peak on a ridge between Mt. Stonehouse and Elliot Peak (Figure 3) argues for considerable antiquity of the landslide at that site.

The present polar climate and the associated slow rates of denudation argue against a recent origin (>14 Ma in age) for the landslide at Mt. Stonehouse. That would imply the loss, under present polar conditions, of several hundred metres of the section from above the Triassic Falla beds that today form the summit of Mt. Stonehouse. The Stonehouse slide differs from all others in that it is the only one consisting mainly of rubble. The preferred interpretation is that the landslide was transported over snow and ice from a distant Kirkpatrick Basalt outcrop before the pre-middle Miocene at a time when the ice was much thicker and warm-based. Subsequently, the landslide debris was let down onto bedrock as a result of pre-middle Miocene wasting of that much thicker ice. This scenario supports the proposal that the present-day topography is middle Miocene or older.

The landslide east-northeast of Mt. Falla, like that at Buttress Peak, must represent an early stage in the evolution of the landscape. Considerable erosion and generation of relief must have occurred both before emplacement and subsequent to emplacement, leading to its preservation on a ridge shoulder. The age of this landslide lies somewhere in late Jurassic to Eocene time. The youngest extant pre-glaciation beds are the Kirkpatrick lavas of late Early Jurassic age (182.7 Ma; Burgess et al. 2015). It is possible the Kirkpatrick Basalt was overlain by up to 1.5 km of additional lavas, as has been suggested by zeolite zonation (Elliot 1970), and/or a thick Mesozoic sedimentary succession. If either or both of these suggestions are correct, then the landslide emplacement age was probably closer to Early Eocene rather than Cretaceous.

Incision of the Mesozoic landscape may have begun any time from the middle Jurassic onwards. Fission-track data (Fitzgerald 1994, 2002; Fitzgerald and Stump 1997; Lisker et al. 2014) and ages of apophyllite in the basalt successions (Fleming et al. 1999) suggest middle Cretaceous denudation or uplift and therefore initial development of the landscape at least by that time. The essentially flat-lying and layer-cake stratigraphy of the Beacon strata and Ferrar Large Igneous Province rocks suggests that after stripping off the possible Mesozoic sedimentary cover the landscape evolved principally as a mesa and butte topography. In the early stages of evolution, the landscape would have been dominated by the resistant Kirkpatrick Basalt lavas, at least 450 m thick, which overlie weak strata of the Prebble and Hanson formations and the alternating sandstones and fine-grained beds of the Triassic Falla and Fremouw formations. Ferrar dolerite sills occur within the sedimentary succession, but more prominently in the Permian beds than the Triassic strata. Such a terrain with steep topography, particularly if in a semi-arid climate, could yield large landslide blocks. The landslides may have resulted from stress-release consequent on over-steepening of cliff faces and/or the involvement of water to increase pore-pressures and as a lubricant during movement. It is possible but not a requirement that the landslides could have been triggered by earthquakes. If water was involved in the generation and movement of the landslides, it implies a warmer pre-glacial climate (Eocene or older) or an early warm-based glacial climate (Eocene/Oligocene to middle Miocene).

Given the evidence for possible late Cretaceous slow uplift, the Transantarctic Mountains during that time were most probably a slightly elevated terrain, which was located in the hinterland of the plate margin magmatic arc and separated from the Panthalassic Ocean by hundreds of km. Although the higher elevation would have been a focus for precipitation, the climate was probably dictated mainly by its continental interior location. The climate was unlikely to have been humid. With the onset of more rapid uplift in the Eocene, the climate was probably more rigorous, with possibly increased precipitation and more pronounced annual temperature variations.

The Mt. Falla and Buttress Peak landslides were emplaced after considerable erosion of the landscape and no later than the Early Eocene. Relief of hundreds of metres had to have been carved by that time. Those two and the Barnes Peak locality involve hundreds of metres of intact stratigraphic successions and probably were more extensive laterally at the time of emplacement. It suggests major processes, such as large magnitude earthquakes and/or unusual hydrological conditions, were necessary to facilitate those large landslides. Major earthquakes, in particular, were more probable during the early stages of rapid uplift of the Transantarctic Mountains.

Other landslides were most probably emplaced later in the denudation history of the mountains. They document localised collapse of steep cliffs, possibly driven by freeze and thaw in the case of the north-facing Elliot Peak slide. Further, because 8 km of cliff retreat under cold-based glaciation seems unreasonable, the inferred landslide near Mt. Stonehouse suggests today's topography was essentially created by mid-Miocene time.

Summary/conclusions

To the author's knowledge, aside from Barrett's (1968) PhD dissertation, in which other landslide blocks were noted, and a conference abstract (Elliot 2002), this is the first report of landslides in the Transantarctic Mountains within the peer-reviewed, published literature. The landslides include intact stratigraphic successions as much as 300 m thick, three of which occur in settings unrelated to present topography. Others are offset from the nearest potential source and suggest the possibility of transport over snow. Yet others are proximal to their sources and appear to be consistent with present-day topography. Only one landslide may be formed predominantly of rubble.

Two of the known landslides must date from early stages in the evolution of the Transantarctic Mountains, probably late Cretaceous to Eocene time. Others, date from the early stages of warm-based glaciation in early Oligocene through middle Miocene time. Yet others, proximal to their sources, must post-date the middle Miocene transition to cold-based conditions, by which time the present-day landscape had been established. Landslides, therefore, occurred in the three major stages in the evolution of the climate and landscape. What is remarkable and unusual is the preservation of landslides possibly more than 55 m.y. old.

The widespread occurrence of landslide blocks in the central Transantarctic Mountains suggests that a thorough search for similar features elsewhere might be rewarding. Further, the landslides described here deserve re-investigation in order to better assess their ages of emplacement, the associated climatic conditions, and their triggering mechanisms. In the central Transantarctic Mountains, access to all but Barnes Peak and the Storm Peak locality is relatively easy and does not require helicopter support.

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Data availability statement

All data in support of this paper are available from the author [D. H. E.] on request.

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