

Chapter 2.1a

Ferrar Large Igneous Province: volcanology

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Abstract: Preserved rocks in the Jurassic Ferrar Large Igneous Province consist mainly of intrusions, and extrusive rocks, the topic of this chapter, comprise the remaining small component. They crop out in a limited number of areas in the Transantarctic Mountains and southeastern Australia. They consist of thick sequences of lavas and sporadic occurrences of volcanoclastic rocks. The latter occur mainly beneath the lavas and represent the initial eruptive activity, but also are present within the lava sequence. The majority are basaltic phreatomagmatic deposits and in at least two locations form immense phreatocauldrons filled with structureless tuff breccias and lapilli tuffs with thicknesses of as much as 400 m. Stratified sequences of tuff breccias, lapilli tuffs and tuffs are up to 200 m thick. Thin tuff beds are sparsely distributed in the lava sequences. Lava successions are mainly 400–500 m thick, and comprise individual lavas ranging from 1 to 230 m thick, although most are in the range of 10–100 m. Well-defined colonnade and entablature are seldom displayed. Lava sequences were confined topographically and locally ponded. Water played a prominent role in eruptive activity, as exhibited by phreatomagmatism, hyaloclastites, pillow lava and quenching of lavas. Vents for lavas have yet to be identified.

The discovery of thick dolerite sills in south Victoria Land (Fig. 1) was made by the National Antarctic Expedition, 1901–04, the first of R.F. Scott's expeditions. The field setting of the rocks was described by the expedition geologist H.T. Ferrar (1907) and the petrography by G.T. Prior (1907). Dolerites, mainly erratics, were also collected by members of the British Antarctic Expedition, 1907–09, from the Beardmore Glacier and Ferrar Glacier regions, as well as the coastal region of Victoria Land as far north as the David Glacier (Benson 1916; Mawson 1916). Benson (1916) also noted an erratic with the petrography of a tholeiite and with interstices filled by skeletal feldspar in dark brown glass, a rock that today would be considered a lava. Campbell-Smith (1924) described the dolerites collected from Buckley Island and Mount Darwin at the head of the Beardmore Glacier and from south Victoria Land by the British Antarctic ('Terra Nova') Expedition, 1910–13. Several erratics from the Terra Nova Bay region were later reported (Campbell-Smith 1964) to have glassy mesostases and zeolite-filled amygdaloids, which, with one exception, were thought to be intrusive rocks rather than lavas. The Australasian Antarctic Expedition, 1911–14, collected dolerites from Horn Bluff, as well as dolerite erratics, and these were described by Browne (1923). L.M. Gould, geologist on the first Byrd Antarctic Expedition (1928–30), collected a diabase from Mount Fridtjof Nansen in the Queen Maud Mountains (Gould 1931, 1935), thus extending the known distribution of the dolerite sills yet farther along the Transantarctic Mountains.

Harrington (1958) suggested the name Ferrar Group for the dolerite sills and dykes that are so abundant in the Dry Valleys, and which are known to occur throughout much of the Transantarctic Mountains. During the International Geophysical Year (1957–58) extrusive equivalents were found *in situ* in the Allan Hills–Coombs Hills region and at Westhaven Nunatak, both in south Victoria Land (Fig. 1) (Gunn and Warren 1962). At the same time a layered basic intrusion was discovered at the Dufek Massif, Pensacola Mountains (Aughenbaugh 1961; Walker 1961), and dolerite sills were found in the Theron Mountains and Whichaway Nunataks (Stephenson 1966). Grindley (1963) broadened the name Ferrar Group to

include basaltic lavas that cap the Devonian–Triassic Beacon succession into which the dolerite sills were intruded, and named the lavas the Kirkpatrick Basalt (although commonly referred to as basalts, strictly speaking the majority of the Ferrar rocks have a basaltic andesite composition). Thus, basaltic lavas and pyroclastic rocks first reported by Gunn and Warren (1962) were included in the Ferrar Group. Subsequent field investigations showed that igneous rocks assignable to the Ferrar Group are widespread in the Transantarctic Mountains, cropping out from the Theron Mountains near the Weddell Sea to Horn Bluff, NW of north Victoria Land (Fig. 1). Later, Ford (1976) correlated the Dufek intrusion with the Ferrar Group. Kyle *et al.* (1981) introduced the name Ferrar Supergroup for all the intrusive and extrusive rocks of Jurassic age, and subsequently Kyle (1998) used the name Ferrar Large Igneous Province (FLIP) for these tholeiitic sills, dykes and extrusive rocks. That name now encompasses those rocks as well as the Dufek intrusion and the tholeiites in southeastern Australia, Tasmania and New Zealand (Milnes *et al.* 1982; Hergt *et al.* 1991; Mortimer *et al.* 1995; Bromfield *et al.* 2007). It is proposed here that the name Ferrar Large Igneous Province be formally established for these rocks together with those in southeastern Australasia, all of which are characterized by distinctive chemistry (see Elliot and Fleming 2021). The name Ferrar Group is retained for the rocks belonging to the FLIP but restricted to outcrops in Antarctica.

The extant Ferrar Group is dominated by intrusive rocks, and the subordinate lavas and pyroclastic deposits are scattered in relatively small areas between the Grosvenor Mountains at the head of the Shackleton Glacier and the Litell Rocks situated in the lower reaches of the Rennick Glacier in north Victoria Land (Elliot and Fleming 2008, 2017). Details of the occurrence and distribution of the intrusive rocks are given in the chapter on the geochemistry of the Ferrar LIP (Elliot and Fleming 2021).

The Ferrar province has a limited exposed volume. The volume of dolerite sills is estimated to be about $1.7 \times 10^5 \text{ km}^3$, assuming an outcrop belt 150 km wide. The lavas are estimated to have a volume of several thousand cubic kilometres assuming continuity within the principal areas of outcrop

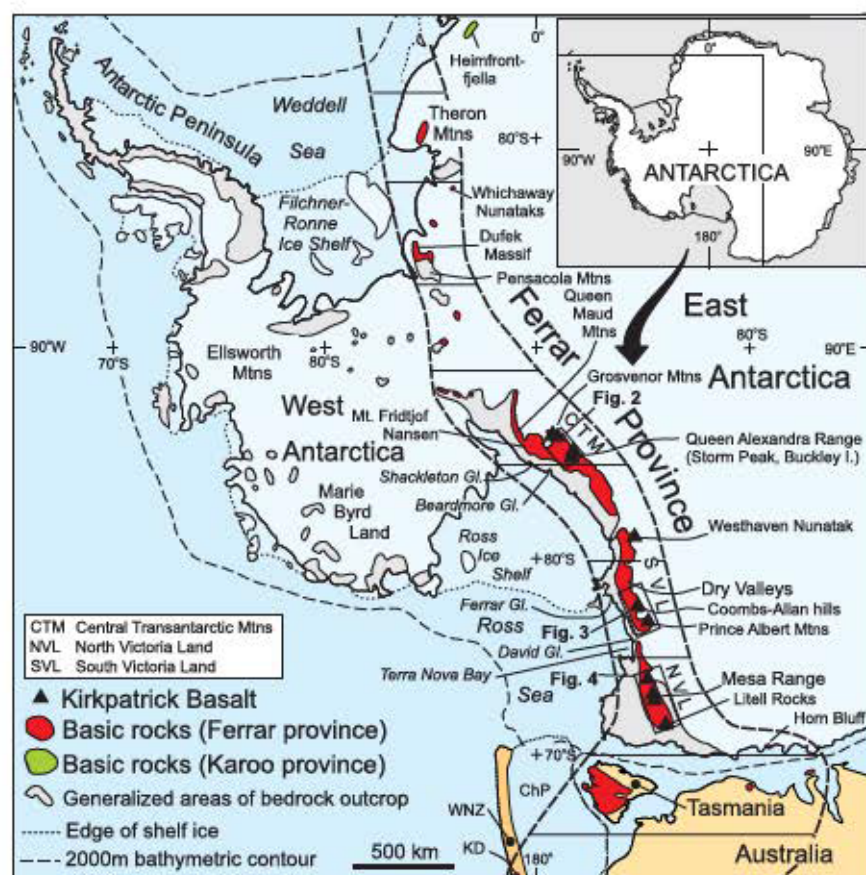


Fig. 1. Location map for the Ferrar Large Igneous Province. In a Gondwana reconstruction, New Zealand would have been off Tasmania. WNZ (solid outline), South Island, New Zealand, west of the Alpine fault. Kirwans Dolerite (KD) with Ferrar composition crops out in northwestern South Island. ChP, Challenger Plateau. Heavy dotted outline includes both WNZ and ChP.

(Fleming *et al.* 1995), but originally it must have been much greater. Although the geochemistry is described in the next chapter (Elliot and Fleming 2021), the division into two chemical types is noted here (Fleming *et al.* 1992, 1995). The Mount Fazio Chemical Type (MFCT) forms about 99% of the province and most of the analysed rocks. The Scarab Peak Chemical Type (SPCT) occurs only as the capping lava of most sequences and as a few sills in the Weddell Sea sector of the Ferrar province.

The age of the Ferrar rocks remained a little uncertain, other than Mesozoic, until the advent of radiometric age determinations when it was established that they are Jurassic in age. Initial results using the whole-rock K–Ar method were superseded by the analysis of plagioclase using the $^{40}\text{Ar}/^{39}\text{Ar}$ technique, but issues remained that were concerned mainly with monitor ages and the differences compared to U–Pb ages. The early U–Pb age determinations by multigrain zircon analysis (Encarnación *et al.* 1996; Minor and Mukasa 1997) have been overtaken by the single-crystal chemical-abrasion isotope-dilution thermal ionization mass spectrometry (CA-ID-TIMS) method (Table 1) (Burgess *et al.* 2015). Because of the sparsity of zircon in the extrusive rocks, U–Pb age determinations are primarily for dolerites. A restricted duration of emplacement (<0.4 Ma) is suggested for 14 Ferrar dolerite sills, with ages ranging from 182.78 ± 0.04 to 182.59 ± 0.08 Ma; two granophyre samples of the Dufek intrusion gave ages of 182.70 ± 0.05 and 182.63 ± 0.03 Ma (Burgess *et al.* 2015). A dolerite from Red Hill, Tasmania, part of the Ferrar LIP, gave an age of 182.54 ± 0.06 Ma (Burgess *et al.* 2015). Ivanov *et al.* (2017) reported three ID-TIMS U–Pb zircon ages for granophyres in Tasmanian dolerites, the ages ranging between 182.90 ± 0.21 and 182.65 ± 0.42 Ma. Kirkpatrick Basalt lavas from three different sections and forming the capping lava in each case yielded ages of 182.64 ± 0.08 , 182.54 ± 0.20 and 182.43 ± 0.04 Ma

Table 1. Single-grain and multigrain U–Pb zircon ages determined for the Ferrar Large Igneous Province

Location	Sample no.	Rock type	Age (Ma)
<i>CA-ID-TIMS (single grain), Antarctica (Burgess <i>et al.</i> 2015)</i>			
Forrestal Range	PRR -8633	Granophyre	182.700 ± 0.045
Forrestal Range	PRR-09305	Granophyre	182.629 ± 0.029
Nilsen Plateau	96-65-11	Dolerite	182.590 ± 0.079
Roberts Massif	96-74-6	Dolerite	182.746 ± 0.054
Rougier Hill	96-51-67	Dolerite	182.753 ± 0.037
Mount Falla	90-53-12	Dolerite	182.85 ± 0.34
Wahl Glacier	85-6-16	Dolerite	182.753 ± 0.037
Mount Picciotto	85-4-4	Dolerite	182.616 ± 0.049
Mount Picciotto	85-4-18	Dolerite	182.633 ± 0.049
Dawson Peak	85-5-6	Dolerite	182.779 ± 0.033
Pandora Spire	A-236-A	Dolerite	182.689 ± 0.038
Pearse Valley	90-76-13	Dolerite	182.776 ± 0.059
Labyrinth	04-03-04	Dolerite	182.750 ± 0.048
Bull Pass	05-06-01	Dolerite	182.680 ± 0.038
Mount Bumstead	96-55-2	Lava	182.48 ± 0.20
Mount Bumstead	96-52-1	Lava	182.54 ± 0.20
Storm Peak	85-76-63	Lava	182.430 ± 0.036
Brimstone Peak	97-55-1	Lava	182.635 ± 0.077
<i>CA-ID-TIMS (single grain), Tasmania (Burgess <i>et al.</i> 2015)</i>			
Red Hill	97-17	Granophyre	182.540 ± 0.059
<i>AA/CA-ID-TIMS (single grain), Tasmania (Ivanov <i>et al.</i> 2017)</i>			
Northwest Bay	2013-289	Granophyre	182.90 ± 0.21
Northwest Bay	2013-288	Granophyre	182.65 ± 0.42
Cape Q Elizabeth	2013-290	Granophyre	182.75 ± 0.45
<i>TIMS (multigrain), Antarctica (Encarnación <i>et al.</i> 1996)</i>			
Dawson Peak	90-63-9	Dolerite	183.4 ± 1.4
Pearse Valley	90-76-12	Dolerite	183.8 ± 1.6
<i>TIMS (multigrain), Antarctica (Minor and Mukasa 1997)</i>			
Forrestal Range	93D-76	Granophyre	183.9 ± 0.3
Forrestal Range	93D-86	Granite dyke	182.7 ± 0.4

CA-ID-TIMS, chemical-abrasion isotope-dilution thermal ionization mass spectrometry; AA, air abrasion.

(Burgess *et al.* 2015). The latter suggests that at least part of the capping unit is permissibly slightly younger than the bulk of the Ferrar LIP. One lava from close to the base of the lava sequence gave a poorly constrained age of 182.48 ± 0.20 Ma, which is indistinguishable from the sill ages. The duration of Ferrar LIP magmatism was estimated to be 349 ± 0.49 ka (Burgess *et al.* 2015).

The Ferrar LIP, like the Karoo LIP of South Africa, exposes the supracrustal architecture of the plumbing system. The Ferrar LIP differs from the Karoo and many other LIPs in the very limited duration of emplacement (Burgess *et al.* 2015), a linear outcrop pattern along the Transantarctic Mountains, the dominance of a single set of chemical compositions and, perhaps most importantly, a distinctive isotopic signature (Elliot and Fleming 2008, 2017). The geochemistry is discussed in the next chapter.

Extrusive rocks (distribution, volumes and extent)

Extrusive rocks occur as isolated and limited outcrops in two distinct regions in the central Transantarctic Mountains (Fig. 2): at the head of the Shackleton Glacier in the Grosvenor Mountains and Otway Massif; and adjacent to the Beardmore Glacier in the Queen Alexandra Range (Barrett *et al.* 1986). In Victoria Land, outcrops, apart from an isolated occurrence at Westhaven Nunatak (Fig. 1) (Gunn and Warren 1962), are scattered over 600 km between the Allan–Coombs hills region (Kyle *et al.* 1983; Bradshaw 1987; Roland and Wörner 1996; Demarchi *et al.* 2001; Ross *et al.* 2008a) and Litell Rocks (Skinner *et al.* 1981) (Figs 3 & 4), with the only extensive outcrops found in the Mesa Range (Gair 1966; Elliot *et al.* 1986b; Brotzu *et al.* 1988; Homig 1993; Hanemann and Viereck-Götte 2004; Viereck-Götte *et al.* 2007). Lavas comprise the bulk of the extrusive rocks, with a small proportion being volcanoclastic and formed by explosive eruptions, and an even smaller proportion being reworked volcanoclastic debris. The minimum volume of lavas, assuming continuity between outcrops within the extant areas, has been estimated to be c. 7000 km³ (Fleming *et al.* 1995), and for the volcanoclastic rocks is estimated to be c. 60 km³ (c. 37 km³ in the Queen Alexandra Range and Otway Massif; c. 20 km³ in south Victoria Land; c. 1.0 km³ in the Prince Albert Mountains; and c. 0.2 km³ in the southern and eastern Mesa Range region, north Victoria Land). The extrusive rocks are remnants of what must have been, at one time, extensive volcanic fields.

Volcanoclastic rocks

Distribution and thickness. Volcanoclastic rocks are assigned to the Prebble Formation in the central Transantarctic Mountains (Hanson and Elliot 1996; Elliot and Hanson 2001); to the Mawson Formation in south Victoria Land (Ballance and Watters 1971; Korsch 1984; Bradshaw 1987; White and McClintock 2001; Reubi *et al.* 2005; Ross and White 2005a; Elliot *et al.* 2006; McClintock and White 2006; Ross *et al.* 2008a) and in the Prince Albert Mountains (Elliot 2002); and are known as the Exposure Hill rocks (formerly Exposure Hill Formation; Elliot *et al.* 1986a) in north Victoria Land (Viereck-Götte *et al.* 2007). Volcanoclastic rocks typically underlie the lavas, but in a few places are found intercalated in the lower part of the lava sequence. Thicknesses of stratified volcanoclastic rocks range up to 200 m (Hanson and Elliot 1996), but unstratified accumulations infilling vent complexes comprising diatreme structures (called 'phreatocauldrons' by White and McClintock 2001) have a vertical extent of at least 370 m at the Otway Massif (Elliot and

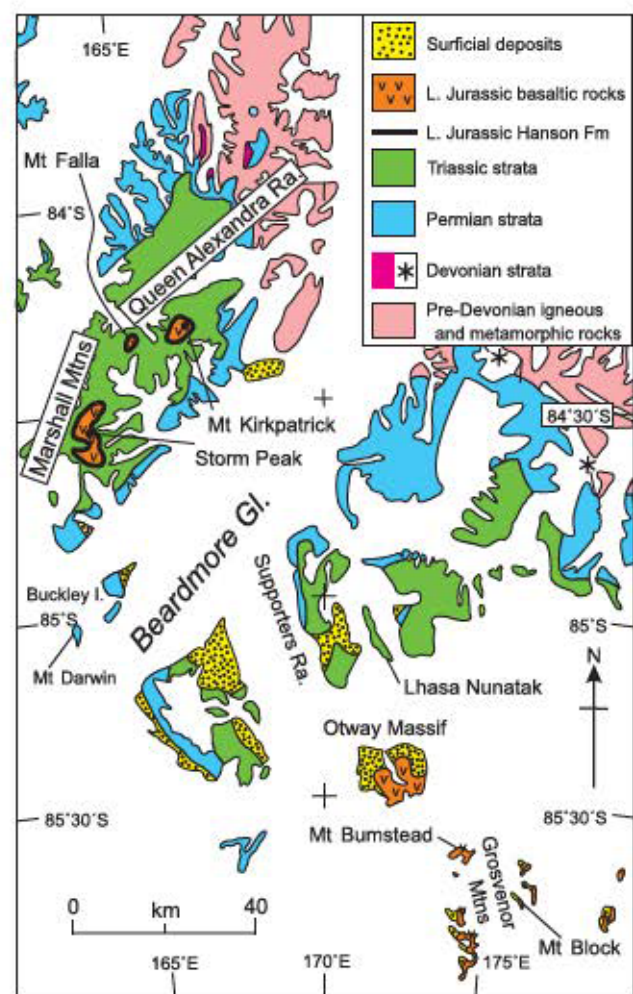


Fig. 2. Simplified geological map of the upper Beardmore Glacier region, central Transantarctic Mountains, illustrating the distribution of the Prebble Formation and Kirkpatrick Basalt lavas (combined as L. Jurassic basaltic rocks in the explanation). Ferrar Dolerite sills and dykes occur throughout the Devonian–Triassic Beacon strata.

Hanson 2001) and more than 400 m in the Coombs–Allan hills area where the outcrop area is more than 30 km² (White and McClintock 2001). Similar features are known from the Karoo province (McClintock *et al.* 2008), but have not been widely identified in the massive volcanoclastic deposits of other flood basalt provinces (Ross *et al.* 2005).

Exposed stratigraphic contacts with underlying Jurassic or Triassic beds are few and far between. Contacts are present in the Queen Alexandra Range (Hanson and Elliot 1996), possibly at Shapeless Mountain and Coombs Hills, south Victoria Land (Korsch 1984; Elliot and Grimes 2011, respectively), and in the Deep Freeze Range and east of Gair Mesa, north Victoria Land (Viereck-Götte *et al.* 2007), where the basal part of the basaltic pyroclastic succession is interbedded with the upper part of the underlying silicic Shafer Peak Formation. More complex contact relationships characterize the northern part of Coombs Hills, where pyroclastic rocks and lava rafts (with an enclosed fossil tree stump; Garland *et al.* 2007) are closely associated with tilted and broken 'rafts' of Lashly Formation country rock of varying sizes, and all enclosed in rock mapped as a mixture of Ferrar Dolerite sills and dykes cutting the Lashly Formation (Grapes *et al.* 1974). White *et al.* (2009) interpreted these relationships to indicate that at the edge of the Coombs Hills phreatocauldrons the intruding sills had shoaled to the surface (cf. Muirhead *et al.* 2014). These sills engulfed broken blocks of the shallow

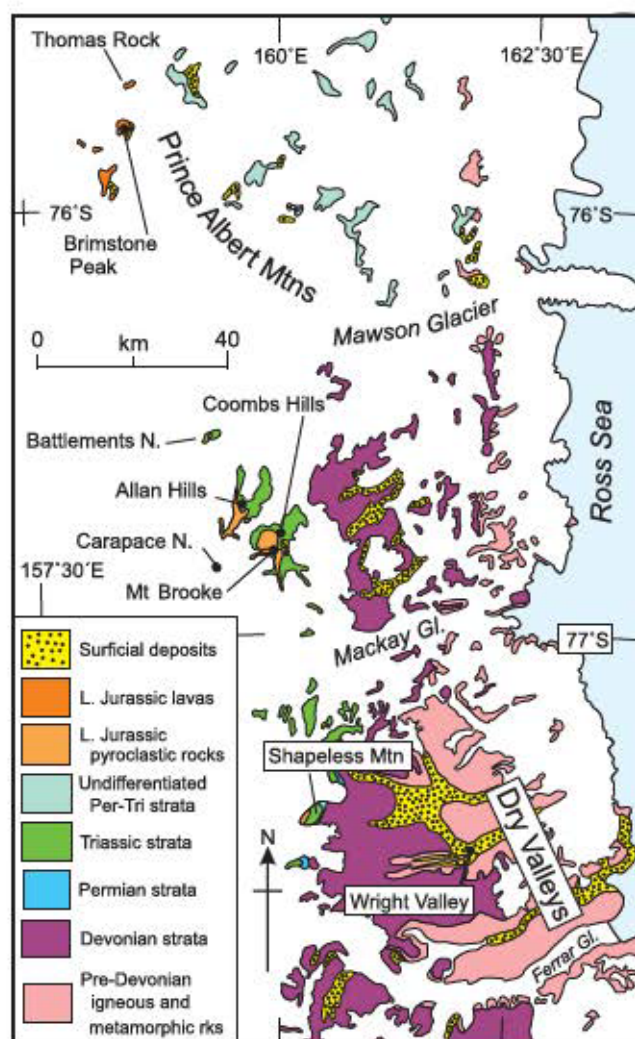


Fig. 3. Simplified geological map of south Victoria Land to show the distribution of the Mawson Formation and Kirkpatrick Basalt lavas. Ferrar Dolerite sills and dykes occur throughout the Devonian–Triassic Beacon strata; however, at Battlements Nunatak sills alone are present.

country rock beneath which magma was initially intruded, and enclose surface-emplaced pyroclastic deposits.

Facies and origin in south Victoria Land. Stratified Ferrar volcanoclastic rocks include both primary volcanoclastic deposits and others inferred to comprise debris redeposited in stream and lake environments. All of these are varieties of mafic volcanoclastic deposits (MVD) such as are known to be associated with many large igneous provinces (Ross *et al.* 2005). The lithofacies scheme used is shown in Table 2, with summary interpretations of the different lithofacies. Outcrops in the Allan and Coombs hills areas provide exceptional exposures of these rocks, which at all known localities mark the initiation of Ferrar volcanism.

At Allan Hills in South Victoria Land (Figs 1 & 3) prominent stratified deposits exposed over c. 6 km (Fig. 5a) include two different kinds of beds (Ballance and Watters 1971; Grapes *et al.* 1974; Ross and White 2005a; Ross *et al.* 2008a). Thick beds with associated accretionary lapilli are laterally persistent and contain small 'rags' of once-glassy basalt indicating inhomogeneous temperatures within the pyroclastic currents that deposited them. Their thickness and extent indicate emplacement from large-volume eruptions from a source beyond Allan Hills, inferred to have been in neighbouring Coombs Hills (Fig. 3). With metres-thick deposits of basaltic

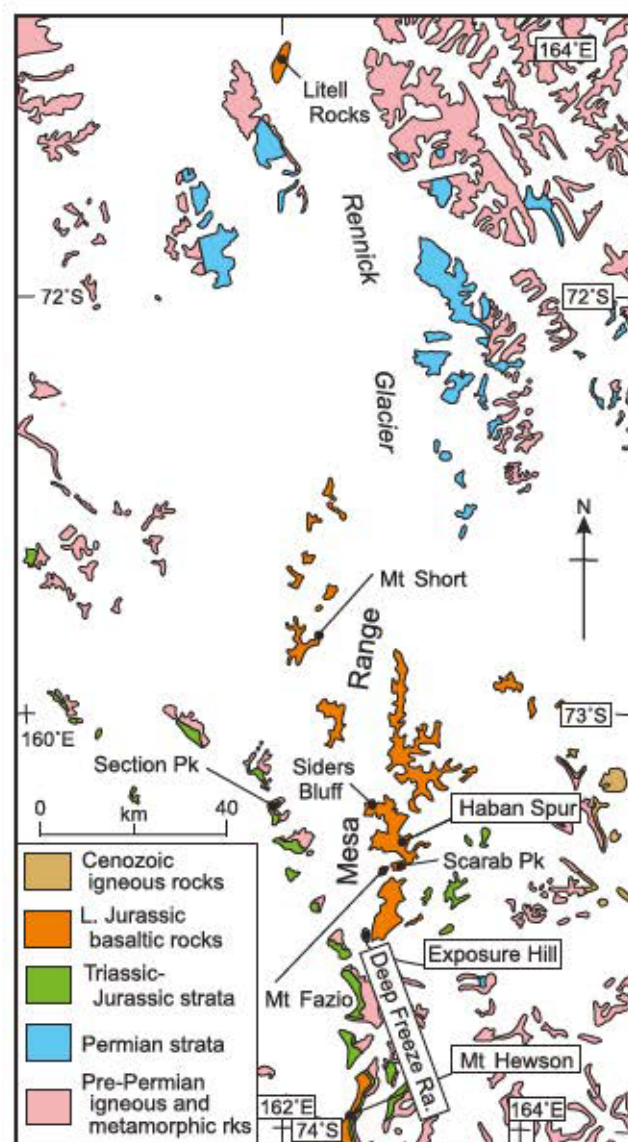


Fig. 4. Simplified geological map of north Victoria Land to illustrate the distribution of the Exposure Hill rocks and Kirkpatrick Basalt lavas. Ferrar Dolerite sills and dykes occur throughout the Permian–Triassic Beacon strata.

pyroclastic density currents (PDCs) covering areas of at least 100 km², the volumes produced by single eruptions were on the order of 1 km³. The PDCs are inferred by Ross and White (2005a) to have been dilute, moist and turbulent, based on lack of welding, abundance of lithic particles and the presence of large (up to 4.5 cm) rim-type (Schumacher and Schmincke 1991) accretionary lapilli.

The thick beds overlie thin-bedded deposits with bedding features like those of small tuff rings (planar lamination, local dunes, bombs and blocks with bedding sags: Ross and White 2005a). The thin-bedded unit extends for more than 1 km without clear thinning or fining trends, and comprises numerous lenses overlapping along a single stratigraphic level. Multiple local sources, probably lying along a major fissure, are inferred from the lenticularity and large blocks (some exceeding 2 m). The layering, blocks and thickness variations are reminiscent of tuff ring deposits, but in a linear array rather than surrounding a single vent (cf. Sohn and Park 2005).

At Coombs Hills (Fig. 3), non-stratified volcanoclastic rocks are dominant over a large area (Bradshaw 1987; White and McClintock 2001; Ross and White 2006; White *et al.* 2009).

Table 2. *Facies descriptions and interpretations for rocks in the Mawson Formation, Coombs Hills and Allan Hills areas*

Facies	Observations	Interpretations
Heterolithic lapilli tuff (LTh and TBh)	1. Volumetrically dominant facies at Coombs Hills (locally tuff breccia in grain size) 2. At Coombs Hills lacks bedding planes for >300 m vertically; forms thick (up to 15 m) widespread layers at Allan Hills 3. Lateral variations in grain size and componentry at Coombs Hills, but no systematic vertical variations; various vertical variations in thick beds at Allan Hills 4. Comprises formerly glassy basalt fragments (mostly blocky ones), microcrystalline basalt fragments (at Allan Hills), sand-grade detrital quartz particles, Beacon fragments, composite clasts (recycled peperite) and rare granite fragments 5. Basaltic clasts are variably vesicular, mostly dense to incipiently vesicular (vesicularity index of Houghton and Wilson 1989) 6. Host for LTa and TBj zones 7. Contains rafts of Lashly Formation (sandstone to siltstone, plus silicic tuff and silicic tuffaceous sandstone) and rafts of layered, fine-grained, mafic volcanoclastic rocks (some with accretionary lapilli); most of these rafts dip steeply	• The clast assemblage suggests phreatomagmatic eruptions affecting the upper part of the Beacon sequence, occasionally as far down as the base of Victoria Group • At Coombs Hills, an origin as one or several lahars (Hanson and Elliot 1996) – or subaerial pyroclastic flows – filling a pre-existing topographical depression is not favoured because of observations 2, 3 and 7 • At Coombs Hills, emplacement in a vent complex by subterranean debris jets is inferred • Several cycles of eruption were probably necessary to reach proportions of formerly glassy basalt clasts observed in LTh (e.g. Bélanger and Ross 2018)
Beacon-rich lapilli tuff (LTa and TBa)	• Forms steep, pipe-like bodies cross-cutting LTh, with sharp contacts • Locally a tuff breccia • Outlines in map view are generally simple, elliptical in shape, with a long-axis length a few decimetres to a few hundreds of metres, aspect of ratio 0.2–0.8 • Same types of clasts as LTh, but with more abundant country-rock fragments	• Phreatomagmatic fragmentation, vent complex setting • Deposited by Beacon-rich subterranean debris jets • Jets originated when phreatomagmatic explosions occurred near the walls or floor of the vent complex, causing fragmentation of abundant Victoria Group material
Basalt-rich lapilli tuff and tuff breccia (LTj and TBj) – general	• Forms zones, metres to hundreds of metres-wide, cross-cutting LTh • Locally a lapilli tuff • Contains more vesicular basaltic clasts than in other facies (except TBhr) • No more than 5% Beacon clasts in the lapilli + block size fraction	• Vent complex setting
TBj – type 1	• Often relatively sharp contacts with host, relatively compact shapes • Same types of clasts as LTh, but with more abundant basaltic fragments; blocky basalt clasts present • Not strongly associated with basalt pods and/or peperite domains	• Phreatomagmatic fragmentation • Debris jets propelled by explosions taking place well away from country rocks • Material in the jets is richer in basalt than the surrounding LTh debris because of the addition of juvenile basalt
TBj – type 2	• Abundant in western Coombs Hills • Generally diffuse gradational contacts with the host rock, outlines can be very complex (e.g. octopus-like) • Contains fluidal basalt fragments and composite clasts, and few or no blocky clasts • Spatially associated with <i>in situ</i> peperite domains, and/or pods of glassy basalt	• Somewhat less violent origin than for LTa, LTh and type 1 TBj zones • Juvenile clast-forming processes inferred to be similar to those in peperite (mostly non-explosive: e.g. surface tension effects, magma-‘sediment’ density contrasts, instabilities in vapour films) • Mixing of juvenile fragments with surrounding volcanoclastic material (incorporation of quartz grains, etc.) • Rags transported while still plastic (high temperature) • LTh-type material simultaneously transported with rags was probably cool (quenched, blocky basalt clasts and Beacon material) • Zones containing ‘randomly’ or subvertically aligned rags could have formed when phreatomagmatic explosions accelerated vesiculating melt not directly involved in the explosions • Zones containing subhorizontally aligned rags at Allan Hills reflect emplacement by pyroclastic density currents
‘Raggy’ heterolithic tuff breccia (TBhr)	• Volumetrically minor at Coombs Hills • Tuff breccia version of LTh with abundant ‘rags’; common at the top of thick beds at Allan Hills • Rags = relatively vesicular, glassy basaltic fragments, elongate, up to several decimetres long, with bent shapes, delicate ends that form spiral shapes and displaying accommodation of the surrounding clasts • Rags can be aligned in any orientation or be ‘randomly’ dispersed in unbedded Coombs Hills deposits; weak subhorizontal orientations at the top of Allan Hills thick beds	• Rags transported while still plastic (high temperature) • LTh-type material simultaneously transported with rags was probably cool (quenched, blocky basalt clasts and Beacon material) • Zones containing ‘randomly’ or subvertically aligned rags could have formed when phreatomagmatic explosions accelerated vesiculating melt not directly involved in the explosions • Zones containing subhorizontally aligned rags at Allan Hills reflect emplacement by pyroclastic density currents

After Ross and White (2005a, 2006), Ross *et al.* (2008a, b, c).

Within this area, domainal outcrop patterns are characteristic (Fig. 6), with steeply dipping, irregular and diffuse contacts separating different lithofacies defined primarily by

differences in the relative abundance of juvenile particles v. country-rock clasts and country-rock-derived sediment grains (Fig. 7; Table 2). The most abundant lithofacies is a

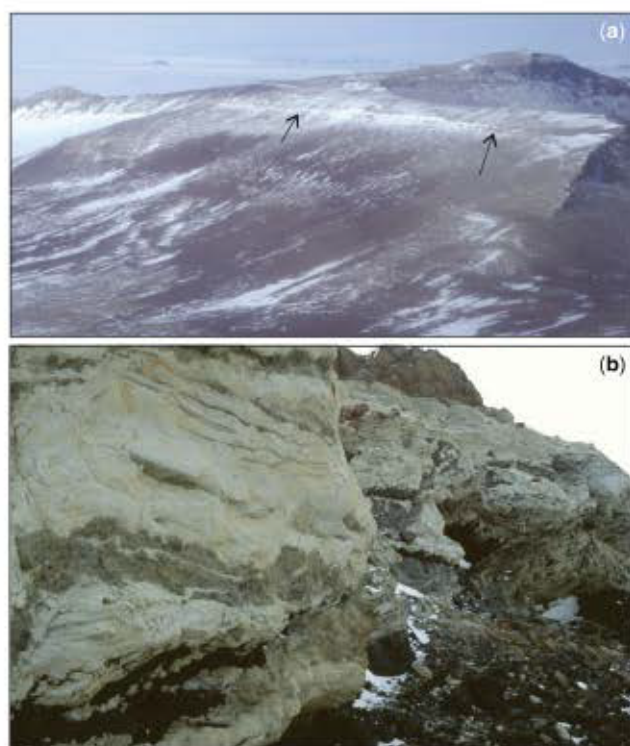


Fig. 5. Layered Mawson deposits at Allan Hills. (a) Aerial view northwards of laterally extensive layering (arrows). (b) Avalanche deposits in the central Allan Hills. Lashly Formation strata, disrupted Ferrar magma intrusion, showing (lower left) broken coal and fine-grained beds enclosed in sandstone. Locally the deposits (not in image) contain domains of fluidally deformed basalt inferred to have triggered avalanching.

heterogeneous lapilli tuff (locally coarser-grained tuff breccia) consisting of 70–80% glassy juvenile fragments and up to 30% Beacon sedimentary rock fragments. Other domains are much richer in sedimentary rock fragments (50–90%) or in juvenile material (up to 70–90%).

The non-stratified rocks locally contain tilted rafts of thin-bedded deposits including accretionary lapilli, interpreted to be remnants of tuff-ring deposits (White and McClintock 2001), and similar beds locally overlie the non-stratified deposits (McClintock and White 2006). Another very common component of the non-stratified deposits, also present in both thick and thin beds at Allan Hills, is a kind of composite clast (White and Houghton 2006). These comprise thin tendrils of glassy basalt mingled on centimetric to millimetric scales with clastic material, most commonly ash (Fig. 8), and are interpreted as fragments produced by disruption of the peperitic zones commonly formed where basalt magma intruded and mingled with previously deposited tephra (Ross and White 2006).

Clastic dykes, specifically pyroclastic ones (Fig. 6), are another distinctive type of deposit. These dykes cut both stratified and non-stratified rocks at Coombs Hills (Ross and White 2005b, 2006), and are also present at Allan Hills (Grapes *et al.* 1974; Ross and White 2005a). Some of these have exceptional width (up to 75 m).

A clastic deposit with a very low proportion of volcanic material at Allan Hills was emplaced as a large avalanche of Lashly Formation sandstone, shale and coal beds (Fig. 5b), into which basalt was injected before or during emplacement (Reubi *et al.* 2005; Lockett and White 2008). It comprises a chaotic assemblage of breccia domains and megablocks as large as 80 m, derived from underlying rocks of the Beacon

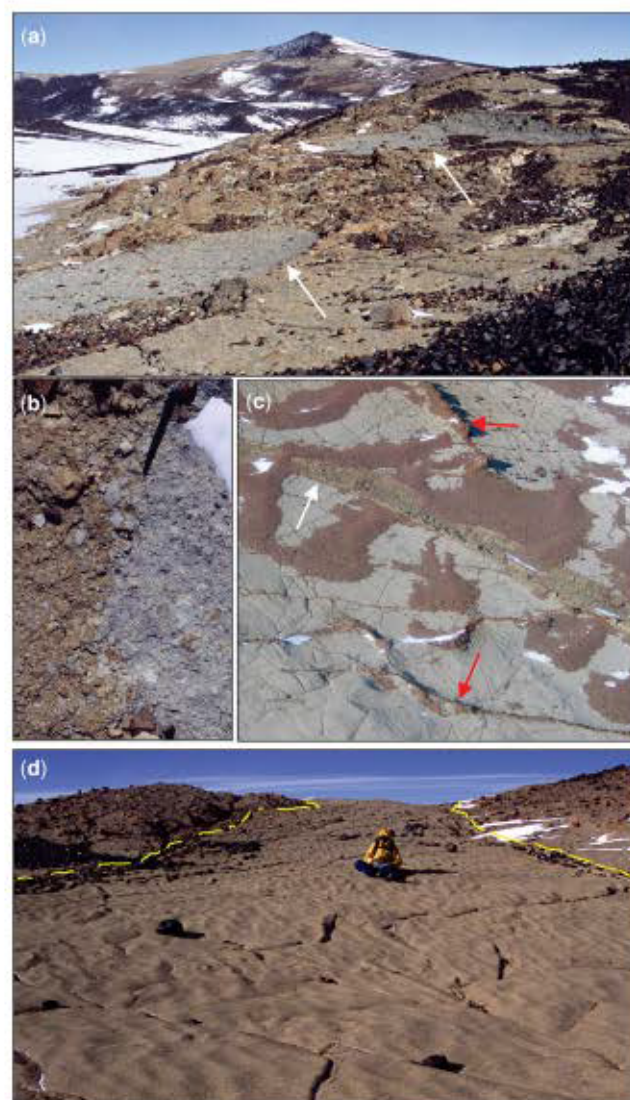


Fig. 6. Distinctive elements of Mawson vent-complex architecture at Coombs Hills. (a) Domainal outcrop interpreted as cross-cutting clastic deposits; pale domains in the midground (white arrows) are LTa, c. 10 m across. (b) Detail of a boundary between LTh and LTa domains. The pencil (upper edge) is for scale. (c) Regular thick clastic dyke (white arrow) and several irregular thin basaltic dykes (red arrows). Field of view is c. 100 m wide. (d) Clastic dyke c. 10 m wide (pale grey rock between the yellow dashed lines); the clast mixtures of the dyke are typically heterolithic like the host, but lack the host's coarser lapilli and blocks (dolerite clasts have accumulated on the surface of the host rocks).

Supergroup, produced by progressive, pervasive and relatively uniform fragmentation of initial megablocks during transport plus minor disruption and ingestion of the substrate. The avalanche was emplaced by northward flow into a pre-existing topographical depression carved into the Beacon sequence (Reubi *et al.* 2005).

At Carapace Nunatak (Fig. 3), pillow lavas (Ballance and Watters 1971), hyaloclastites and coarse tuffs with accretionary lapilli (Bradshaw 1987; Ross *et al.* 2008a) crop out. The tuffs are most probably primary eruption-fed deposits of distal lahar runout flows or possibly PDCs, perhaps including those that emplaced the thick beds at Allan Hills. More significant reworking by streams is a less-favoured possibility.

Thin, subvolcanic dolerite dykes and sills that pass into peperite are widely scattered, and best developed at Shapeless Mountain (Korsch 1984) and at Coombs Hills (Elliot and Grimes 2011). Phreatic explosion vent deposits, comprising

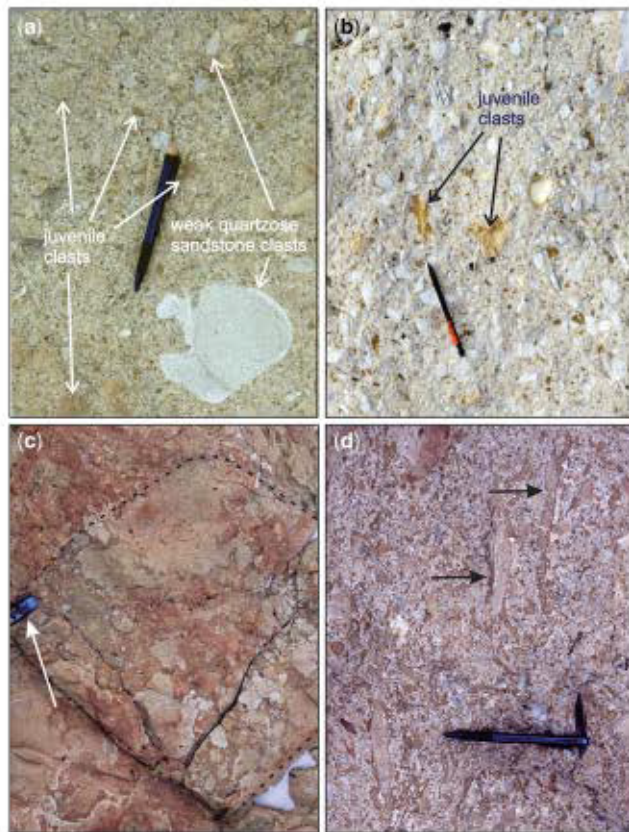


Fig. 7. Dominant clastic lithofacies of Ferrar primary volcanoclastic deposits, with examples illustrated from Coombs Hills. (a) Heterolithic lapilli tuff (LTh). (b) Accidental-rich lapilli tuff (LTa). (c) Juvenile-rich tuff breccia (TBj). The pencil head is for scale (white arrow). (d) Juvenile-rich lapilli tuff (LTj) with 'raggy' fluidal clasts (arrows).

a mixture of fragments from the host strata in a matrix lacking basaltic clasts, have been identified at Coombs Hills where they cut Triassic Lashly Formation strata (Elliot and Grimes 2011).

The mafic volcanoclastic deposits of south Victoria Land have been interpreted in different ways. Gunn and Warren (1962) interpreted the thick-bedded to unbedded deposits comprising a mixture of clast sizes ranging from blocks or boulders to silt-grade fragments as a tillite of glacial origin. Ballance and Watters (1971) called the rocks in the Allan Hills a 'diamictite', identified possible vent deposits and inferred deposition of layered rocks by volcanic mudflows. Bradshaw (1987) further identified vent deposits at Coombs Hills, and noted that granite boulders contained in them had been carried upward, during eruption, through hundreds of metres of stratigraphy.

Interpreting both past work and new mapping results, White and McClintock (2001) recognized the grouping of 'vent deposits' at Coombs Hills as comprising a large vent complex, or 'phreatocauldron', formed by the same suite of processes that form diatremes (Fig. 9). Ross and White (2006) presented new mapping of the internal structure of this complex, and followed their field analysis with a series of experiments to investigate structures characteristic of subterranean vent-excavation processes (Ross *et al.* 2008a, b, 2013; McClintock *et al.* 2009; Valentine *et al.* 2015). White *et al.* (2009) and Muirhead *et al.* (2014) also discussed the role of sills in this setting, suggesting that an alternative interpretation would be that dykes feeding this complex were rooted in a sill(s) at a shallow level below the complex. Such a reconstruction would be consistent with apparent 'shoaling' of sills just north of the diatreme complex (White *et al.* 2009), and inferred processes of generation for

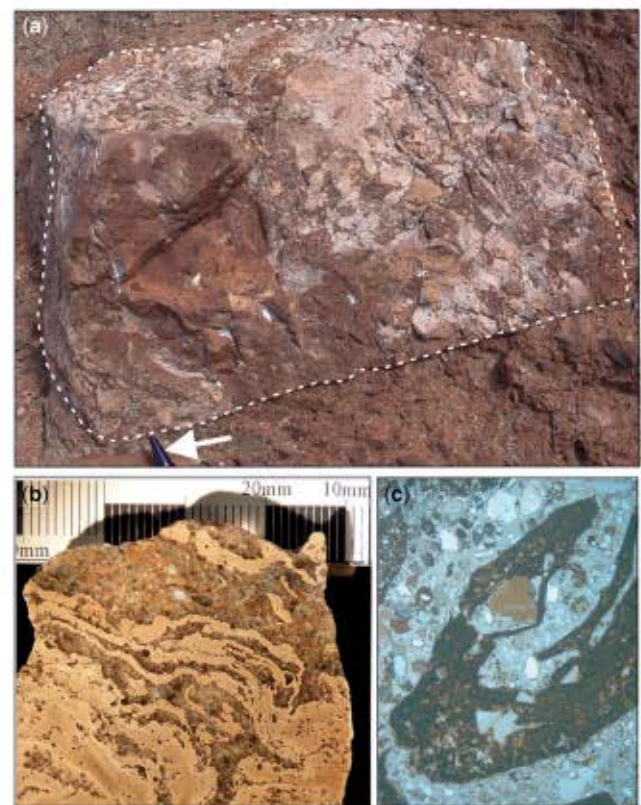


Fig. 8. Distinctive Mawson composite clasts, lapilli (2–64 mm) and clasts >64 mm that are composite bombs and blocks; all are considered to be fragments of peperite. (a) Angular peperite block. Scale given by the pencil end (arrow). (b) Small-scale mingling of basalt (tan, smooth) with LTh lapilli tuff containing quartzo-feldspathic sand from Beacon country rock (darker, grainy). (c) The photomicrograph, with polarizer at 23° (1/4 crossed), shows once-glassy basalt enclosed in, and enclosing, sedimentary grains; field of view is 5 mm.

70 m-wide clastic dykes at Coombs Hills (Ross and White 2005b).

Facies and origin in other areas. At the Otway Massif several hundred metres of massive tuff breccias overlain by thin sequences of tuff beds and capped by lavas at the Otway Massif are interpreted as forming another phreatocauldron (Elliot and Hanson 2001; Elliot and Fleming 2008). These tuff breccias are less well exposed than those at Coombs Hills, but have many of the same characteristics: megablocks of Hanson Formation strata are enclosed in tuff breccia; accidental blocks of silicic tuff and dolerite are up to 10 m in length; areas within the tuff breccias show divergent particle orientations, and varying clast sizes and types; and clastic dykes are scattered widely. Non-bedded tuff breccias with cross-cutting relationships occur at Ambalada Peak, Prince Albert Mountains (Elliot 2002), and probably are another example. At Agate Peak and Exposure Hill, adjacent to the Mesa Range, tuff breccias are interpreted as vent fillings (Elliot *et al.* 1986a; Schöner *et al.* 2007).

Hanson and Elliot (1996) identified extensive deposits of stratified volcanoclastic rock underlying Kirkpatrick Basalt lavas in the Queen Alexandra Range, central Transantarctic Mountains (Fig. 2). Based on a particle population comprising a mixture of glassy basaltic (sideromelane) fragments with sand grains comminuted from country rock, tabular deposit geometries, local entrainment of surface debris into the basal layers and local fluvial channelling, these are interpreted as deposits of lahars. Associated deposits contain accretionary lapilli, and the lahars may have been fed directly from very water-rich

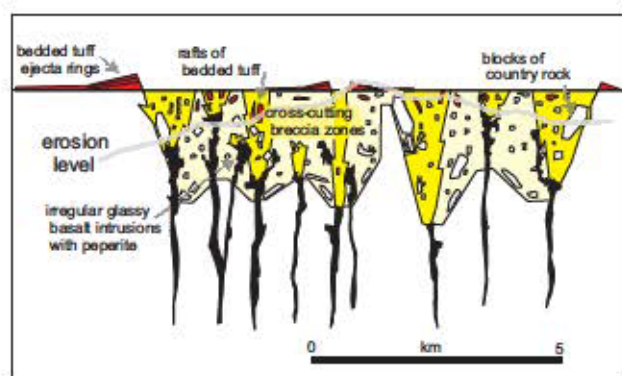


Fig. 9. Schematic cross-section of the Coombs Hills geology showing inferred subterranean development of non-bedded lithofacies in a 'phreatocauldron' or diatreme complex (after White and McClintock 2001) locally capped by, and containing tilted remnants of, bedded tuff and lapilli tuff from surficial tuff rings.

eruptions. Remnants of phreatomagmatic vents, along with small sills and dykes of pyroclastic material, attest to a depositional setting that included volcanic centres. This took place in a basin that was filling with thick Prebble Formation deposits, and beneath which magma interacted with weakly consolidated Falla Formation and older sandstones.

Stratified volcanoclastic rocks (tuff breccias, lapilli tuffs, tuffs and reworked debris) crop out in isolated areas from Mount Pratt, east of Mount Bumstead, to the Mesa Range, and represent remnants of volcanic constructs. Accretionary lapilli are not uncommon. Finer-grained tuffs are attributed to air-fall or base-surge deposition and interpreted as remnants of tuff rings and tephra cones. Most deposits can be attributed to phreatomagmatic activity in which the violent interaction between hot magma and water-rich sedimentary rocks produced mixtures of juvenile pyroclasts and disaggregated siliciclastic sandstones. In contrast, at one locality in the Prince Albert Mountains a c. 10 m-thick breccia, with blocky basalt clasts, spatter up to 2 m long, basalt shreds up to 30 cm long and carbonaceous sedimentary clasts, suggests a locally derived deposit. The overlying 20 m, which lack the number and size of large clasts although basalt shreds are ubiquitous, is succeeded by a few metres of tuff breccia with basalt and sedimentary clasts up to 3 m long (Elliot 2002). Together with stratigraphically associated massive and weakly bedded tuffs, these rocks suggest close proximity to a phreatomagmatic vent.

At Allan Hills and elsewhere in the Ferrar LIP, widespread layers of the same sorts of commonly heterolithic lapilli tuffs were emplaced in different ways: (1) by PDCs; (2) by lahars; and (3) by streamflow. Because they have the same components and similar textures, rocks with features indicating deposition by rivers may well include both primary volcanoclastic rocks from runout of eruption-fed lahars, and those produced by erosion of primary deposits and redeposition of their particles. Those derived from erosion of earlier deposits may be better termed volcanoclastic sandstones and conglomerates. It is important, for context, to appreciate that the variations in dimensions of both non-bedded (tens to hundreds of metres) and bedded (hundreds of metres to kilometres) deposits reflects processes affecting those particular sites, and that the exposed areas are very small relative to the scale of the Ferrar outcrop belt (hundreds to thousands of kilometres). Details of individual beds at different sites hold the record of which depositional process(es) were dominant from one area to the next, and a rich archive of information to be extracted in future research.

Effusive rocks

Distribution and thickness. Lava sequences (Fig. 10) are commonly 400–500 m thick but attain more than 750 m in the Mesa Range (Elliot *et al.* 1986b; Mensing *et al.* 1991; Elliot and Fleming 2008). The lower contact is locally exposed only at the Otway Massif, the Queen Alexandra Range and in Victoria Land, and not at all in the Grosvenor Mountains and the Prince Albert Mountains. Individual lavas range from 1 to 230 m in thickness; many are tens of metres thick, which leads to the layer-cake aspect evident in the field. Although it cannot be demonstrated in the field, lavas thicker than about 100 m are interpreted to be confined topographically in a rift system and locally ponded in the case of the 230 m-thick lava at the Otway Massif (Elliot and Fleming 2008). Thin lavas, in some sections, occur in packages and may occur anywhere in the sequence. Although the thin lavas might be the products of a local volcanic centre, at Storm Peak in the Queen Alexandra Range (Elliot and Fleming 2008, fig 10A) lavas 3–8 on the western ridge are represented by a single lava (designated X in that figure) on the adjacent ridge. This suggests that multiple thin lavas are most probably flow lobes of a much thicker single lava, and the similarity in chemistry of five of those six lavas at Storm Peak provides support for this interpretation. The Mesa Range in north Victoria Land has an almost continuous exposure of lavas over a distance of about 75 km. In a photogeological study of these rocks, Petri *et al.* (1997) distinguished five lava units that were traced throughout the range.

Lava characteristics. The facies classification scheme of Jerram (2002) can be applied to the Kirkpatrick Basalt lavas (Elliot and Fleming 2008). The lavas belong predominantly to the 'tabular-classic flow facies', and, being laterally extensive, many probably represent the sheet lobes of Self *et al.* (1997). These lavas range in thickness from several metres to more than 100 m. Within individual lavas a variety of fracture types commonly occur, but no consistent pattern has been observed (Fig. 10), and the classical colonnade and overlying entablature is seldom displayed. Perhaps one of the most remarkable lavas crops out in the Queen Alexandra Range: at Storm Peak it is as much as 135 m thick, and consists of a thin basal chilled-contact followed by a 3 m-thick colonnade (the latter not always present), 130 m of entablature and a 3 m-thick, slightly vesicular, upper crust. The entablature is entirely tachylitic and comprises downward and outward, diverging and branching, columns 10–20 cm across (Elliot and Fleming 2008, fig. 11). Although occurring as a tachylitic lava at all measured localities in the Marshall Mountains, on the north face of Mount Falla the tachylitic interval appears to thin and pinch out over a few hundred metres, and then reappear and pinch out again in the central part of the face (Fig. 11). The thickness changes in the tachylitic interval probably reflect patterns of quenching and alteration rather than real pinching and swelling of the lava. Stratigraphic sections measured on the NE and NW ridges of Mount Falla illustrate the challenges in lava correlation and in identification of lavas across and between outcrops (see Barrett *et al.* 1986, pl. 1c, secs 71 and 10). The sedimentary interbeds provide the only reliable framework for correlation within the central Transantarctic Mountains (Barrett *et al.* 1986) and in the Mesa Range (Elliot *et al.* 1986b), although detailed geochemical analysis of lavas might yield additional acceptable correlations. The second lava at Mount Kirkpatrick (Elliot and Fleming 2017, fig. 7b), given the MgO concentration and the presence of a thin colonnade and overlying thick black conchoidally fractured basalt, is permissibly correlative with the tachylitic lava at Storm Peak, Mount Falla and elsewhere in the Marshall Mountains. The much greater apparent thickness, half as much again as the

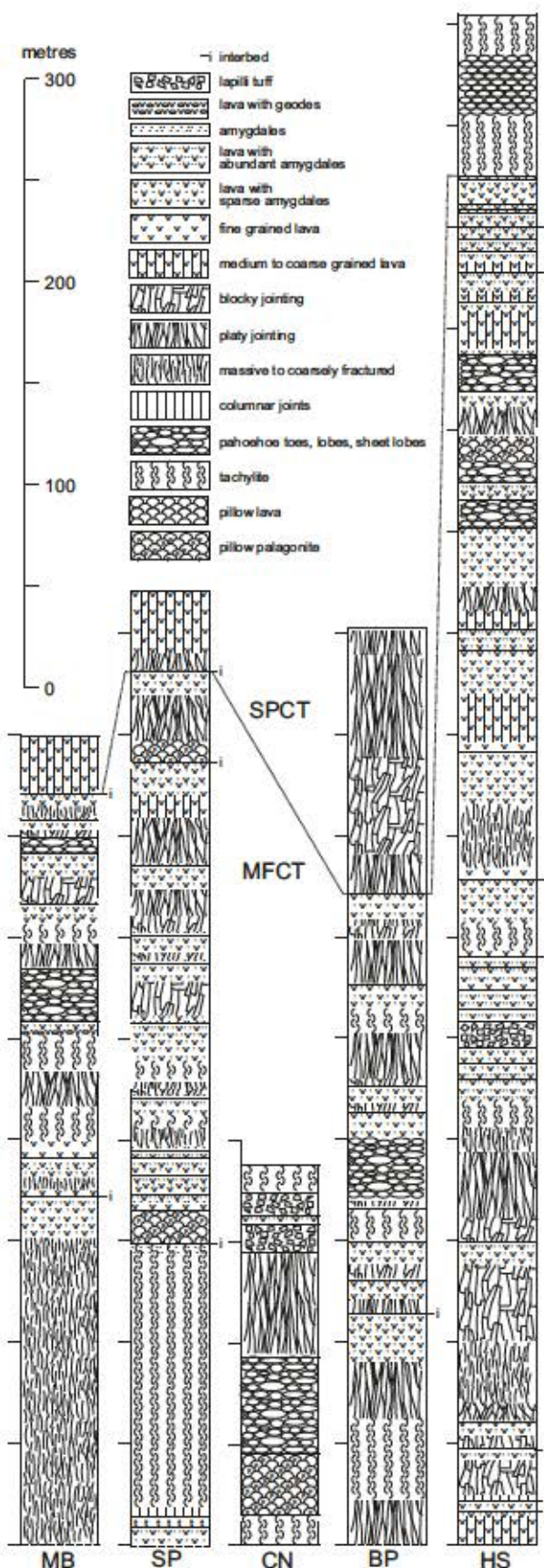


Fig. 10. Simplified stratigraphic columns for Kirkpatrick Basalt lava sequences in the five principal areas of outcrop. Mount Bumstead (MB), Storm Peak (SP), Carapace Nunatak (CN) and Brimstone Peak (BP) from Fleming (unpublished; see also Barrett *et al.* 1986 for MB and SP). Haban Spur (HS) from Elliot *et al.* (1986b). If the inferred correlation between Mount Bumstead and the Otway Massif (Barrett *et al.* 1986) is correct, then the lava sequence in that region is more than 550 m thick.

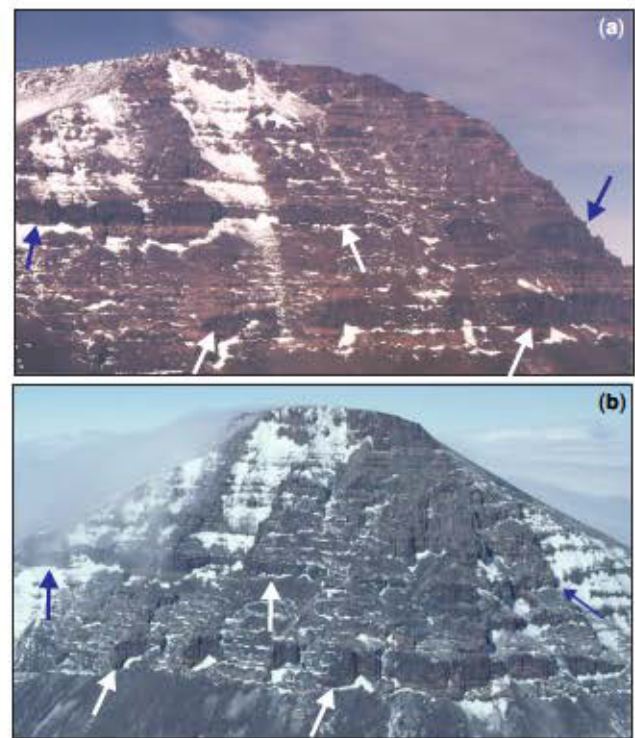


Fig. 11. Kirkpatrick Basalt lava sequence, 475 m thick, exposed on the north face of Mount Falla. (a) Aerial view from near the NE ridge. The black tachylitic lava is exposed on the NE ridge where it is 101 m thick, but this tachylitic interval pinches out westwards. It reappears as prominent cliffs just above the talus slope across most of the width of the mountain (white arrows) but pinches out farther to the west (out of view). Another prominent black tachylitic lava (68 m thick; indicated by black and white arrows) crops out about one-third of the way up the lava sequence along the NE ridge, and that tachylitic interval also thins and thickens westwards. (b) Aerial view from the NW. The arrows correspond with those in (a), and illustrate the lava sequence farther across the mountain front. (Images: D.H. Elliot.)

lava at Mount Falla, is attributed to a fault, down to the north, which is supported by a thin vertical zone of brecciated lava. In north Victoria Land the capping lava of the sequence at all examined localities is also tachylitic (Elliot and Fleming 2008, fig. 10B) and exhibits a variety of fracture patterns, but lacks any columnar jointing. Examples of relatively thin tachylitic lavas are present in many measured sections. A role for water in the rapid chilling and formation of tachylite is implied by the overlying lacustrine bed at Storm Peak and other outcrops in the Marshall Mountains, and is suggested by the lacustrine bed that underlies the capping lava in the Mesa Range; elsewhere such direct evidence is lacking for the tachylitic lavas. Quenched lavas of such thickness (>100 m) must record abrupt flooding of lava surfaces in a topographically confined setting, which has been interpreted as a rift (Elliot and Fleming 2008). The proximal cause could have been a rearrangement of drainage resulting from lava emplacement.

Regional ponding of lavas by topography, on a scale of at least tens of kilometres, is inferred for the very thick lavas (>100 m), and in particular for the basal lava at Mount Bumstead (>175 m) and the >230 m-thick correlative lava at adjacent Otway Massif (Barrett *et al.* 1986, pl. 1c). The palaeosol and tuffaceous beds at the Otway Massif (Elliot and Hanson 2001), which are conformable with the overlying lavas, suggest that the lavas are not simply filling a collapsed or partially collapsed phreatocauldron. All measured sections at the Otway Massif include one or more basal lavas more than 150 m thick.

Intervals of thin lavas are present in many sections, constitute the 'compound-braided flow facies' of Jerram (2002) and represent the flow lobes of Self *et al.* (1997). These sets of thin lavas are scattered among the sections, and are interpreted as either sets of pāhoehoe toes, lobes and sheet lobes marginal to, and break-outs from, thicker lavas of the 'tabular-classic flow facies' or, much less probably because of the absence of evidence for local vents, locally derived thin lavas. However, the lack of a clear relationship, in most instances, to thicker lavas makes their interpretation problematic, as does the absence of any sign of local vents. At Mount Bumstead, where there is an accessible, north-facing, snow-free outcrop, the irregular character of these sets of thin lava units is evident (Fig. 12). Marked differences in individual lava lobe thicknesses are dependent on the degree of inflation, and the commonly thicker, pod-shaped, dense bodies may represent lava tunnels.

The intra-lava characteristics are typical of tholeiitic flood lavas (Self *et al.* 1997). Pāhoehoe toes may be present at the base of a lava, in some instances stacked three or four high, and pass up into sheet lobes. Chilled margins of lobes and lava sheets are normally only a few tens of centimetres thick and are vesicular. Only rarely are there pipe vesicles that show stretching and inclination (Fig. 13), and spiracles are equally uncommon (Barrett *et al.* 1986, fig. 40). Nevertheless, at Mount Kirkpatrick small pipe vesicles in the lowest lava show a consistent west to SW flow direction, and in the Mesa Range inclined pipes suggest an overall westerly flow. Within the body of some thick lavas, vertical pipes (Fig. 14), a few to tens of centimetres in diameter, are filled by segregations of coarser-grained basalt; these pipes interconnect with similar horizontal pipes and sheets. Vesicular upper crusts are almost uniformly present, and in some cases are metres thick. Vesicles may be scattered or in distinct layers, reflecting pulses of inflation, and commonly are filled by secondary



Fig. 12. Kirkpatrick Basalt lavas exposed on the north face of Mount Bumstead. Numerous thin strongly altered lava lobes occupy the interval between a prominent, c. 15 m-thick, black tachylitic interval (indicated by the white arrow; the top of this interval lies 9 m below the top of a 68 m-thick lava), which crosses the whole image, and the capping lava (c. 45 m thick on the right-hand edge of the image). (Image: D.H. Elliot.)



Fig. 13. Siders Bluff, north Victoria Land. Inclined pipe vesicles in the base of a lava overlying a thin pyroclastic interbed filling low spots in the underlying lava (Elliot *et al.* 1986b; unit 35 at 571.5 m in section 81-2). Ice axe for scale. (Image: D.H. Elliot.)

minerals. The upper surface of a lava is seldom exposed in map view and only one instance of a ropy surface has been noted. The tops of some lavas are strongly weathered, form weakly defined palaeosols, which in rare instances include root traces (Fig. 15a, b), and in the Grosvenor Mountains may exhibit the marked oxidation of a red bole. The upper surfaces may carry plant fragments. Geodes and irregular masses



Fig. 14. Haban Spur, north Victoria Land. Pipes, up to 3–4 cm across, of coarser-grained basalt in the upper part of a lava, which here pass abruptly to scattered vesicles (Elliot *et al.* 1986b; unit 11 at 264 m in section 82-3). Hammer is for scale (arrow). (Image: D.H. Elliot.)

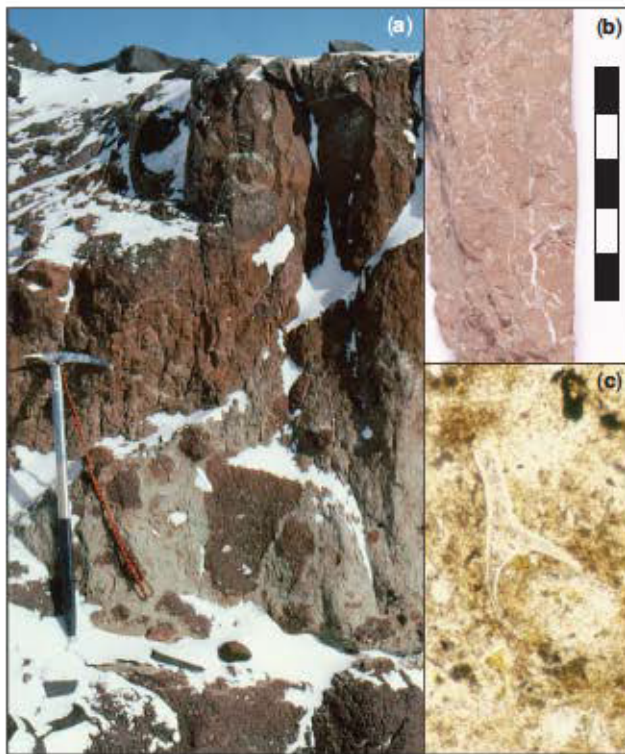


Fig. 15. Mount Block, Grosvenor Mountains. (a) Strongly oxidized and weathered upper part of a lava, forming a red bole. Amygdaloidal basalt fragments, partially destroyed by weathering processes, are scattered throughout the profile. Ice axe (80 cm long) is for scale. (b) Branching tube-like bodies interpreted as rootlets (scale in centimetres). (c) Tricuspid glass shards (c. 0.25 mm across) identified in the weathering profile. Mount Block, Grosvenor Mountains (Fig. 2) (section 61: Barrett *et al.* 1986; Elliot *et al.* 1991). (Images: D.H. Elliot.)

of secondary minerals occur between many pāhoehoe toes or lobes, between pillows and at some lava contacts.

Secondary minerals include quartz, chalcedony, calcite, zeolites, apophyllite, green phyllosilicate and gypsum. Zeolites recognized include stilbite, epistilbite, heulandite, chabazite,

mordenite, clinoptilolite, analcite, natrolite, scolecite and erionite (Barrett *et al.* 1986; Vezzadini *et al.* 1994; Conaway *et al.* 2005). Crystallization of apophyllite in mid-Cretaceous time, with associated zeolite overgrowths, suggests elevated temperatures long after thermal perturbations associated with the Ferrar magmatic event (Fleming *et al.* 1999). At least some of the zeolites post-date apophyllite crystallization and, with their presence in lavas high in the sequence, suggest the possibility of a now-eroded thick lava sequence (Elliot 1970) and/or a middle Jurassic–early Cretaceous sedimentary overburden (Elliot and Fleming 2008), either of which could have been more than 1 km thick. Apatite fission-track data from Victoria Land have also been interpreted to suggest a now-eroded Mesozoic sedimentary sequence formerly overlying the Ferrar lavas (Lisker and Läufer 2013).

Thick intervals of hyaloclastite and pillow lava are prominent at the base of the sequence at Carapace Nunatak (south Victoria Land) and at Thomas Rock in the Prince Albert Mountains. At Mount Fazio in the Mesa Range a pillow lava stack about 50 m high is present (Fig. 16) and passes laterally into a large tachylite mound. Thinner intervals of hyaloclastite and pillow lava are also present higher in sections at the Mesa Range (Fig. 17) (see Elliot and Fleming 2008, fig. 14) and in the Marshall Mountains (Storm Peak). Such intervals indicate standing water in lakes or fluvial channels. The c. 100 m thickness of hyaloclastite and pillow lava at Thomas Rock (Fig. 18) must have originally extended laterally for more than 1000 m (the length of the outcrop), which implies a major lacustrine setting. Water played a significant role in the formation of the Ferrar extrusive rocks as shown by the phreatomagmatic deposits, hyaloclastite formation and quenching of lavas. The Thomas Rock hyaloclastite rocks in particular suggest eruption into an environment with significant local topography. Features advocated by Deschamps *et al.* (2014) to indicate subaqueous eruptions have not yet been observed.

Interbeds

The lava sequences are broken by interbeds, which are mainly lacustrine but also include tuff and lapilli tuff beds (a number



Fig. 16. Mount Fazio, north Victoria Land. Lavas are numbered: (1) lowest lava; (2) 15 m-thick lava with columnar jointing; (3) 3 m-thick lava consisting of small pāhoehoe toes and lobes, which is overlain by a thin volcaniclastic interbed; (4) c. 50 m-thick lava with a 5 m-thick columnar jointed interval at the base, followed by c. 45 m of tachylite; the tachylite passes laterally to small pillows within the white dashed lines, and then to larger pillows (above and to the left of the dashed white lines); the large pillows extend over the top of the tachylite interval; the irregular top of the lava is separated from the overlying lava (5) by patchy thin volcaniclastic beds; and (5) lava with crude columnar jointing (section 82-4: Elliot *et al.* 1986b).

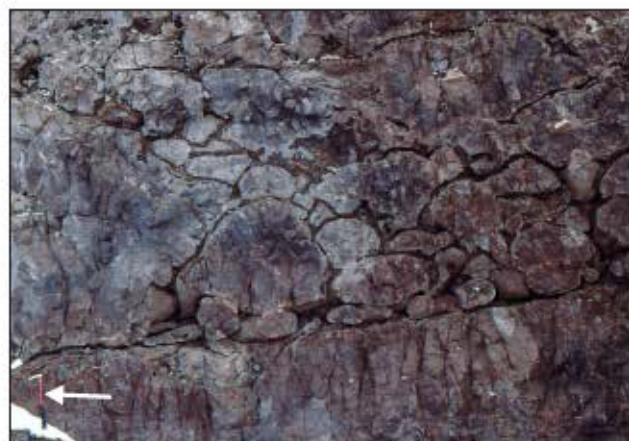


Fig. 17. Mount Short, north Victoria Land. A 14 m-thick stack of lava pillows (Elliot *et al.* 1986b; 6 m above base of section 82-23). Pillows are from 0.5 m to tens of metres across; the upper 2 m consists of pillow breccia. Lava is overlain by 25 cm of bedded pyroclastic debris. Ice axe (80 cm long) in the lower left is for scale (arrow). (Image: D.H. Elliot.)



Fig. 18. At Thomas Rock, Prince Albert Mountains, Kirkpatrick Basalt hyaloclastite and pillow lava overlies Mawson Formation pyroclastic deposits (lower left). The hyaloclastite and pillow lava interval is about 100 m thick, and is overlain by lavas. Inset illustrates the right-hand end of the hyaloclastite and pillow lava outcrop (ice axe in lower left is for scale). (Image: D.H. Elliot.)

containing accretionary lapilli), redeposited tuffs, and palaeosols developed on lava crusts (e.g. Elliot *et al.* 1986b, 1991; Elliot and Hammer 1996). The occurrence of tricusate shards in several weathering profiles in the Grosvenor Mountains (Fig. 15c) has been interpreted as the vertical mixing of silicic ash by vertisol processes (Elliot *et al.* 1991), and continuation at a low level of the silicic magmatism recorded in the underlying Hanson Formation (Elliot *et al.* 2016). Interbeds are common near the base of the lava sequences and characteristically separate the capping SPCT lava from the underlying MFCT lavas. The SPCT lava and the underlying interbed are a significant marker for correlation in the Kirkpatrick Basalt throughout the Transantarctic Mountains, and the lacustrine bed above the 135 m-thick lava at Storm Peak is similarly a widespread marker bed in the Queen Alexandra Range.

The interbeds have yielded a varied flora and fauna. Identifiable plant remains, which include ferns, cycads, conifers and plant microfossils, have been found at Carapace Nunatak and the Mesa Range region (Plumstead 1962; Townrow 1967;

Ribecai 2007; Bomfleur *et al.* 2011; Heiger *et al.* 2015). Vertebrates are restricted to Pholidophoroid fish remains recovered from the interbeds at Storm Peak and elsewhere in the Marshall Mountains (Schaeffer 1972). Lake beds, from the Grosvenor Mountains to the Mesa Range, have yielded abundant invertebrates which are principally conchostracans but include ostracods, notostracans, syncarids, molluscs, beetle elytra and insects (Carpenter 1969; Ball *et al.* 1979; Tasch 1987; Shen 1994; Stigall *et al.* 2008; Bomfleur *et al.* 2011).

Vent locations and eruption rates

The locations of phreatocauldrons are clear and pyroclastic rocks, where exposed, always occur beneath the lavas, suggesting that numerous eruptive centres existed in the early stages of extrusive activity. Locally pyroclastic rocks (hyaloclastite with pillow basalt, tuff and lapilli tuff) may be intercalated higher in the lava successions. Co-location of vents for effusion of the lavas might be expected, but possible connections between a sill or a dyke and the lava sequences have not been observed. Northern Coombs Hills shows apparent 'shoaling' of a sill toward the surface, but an overlying flood lava sequence is not present at that site.

The Dry Valleys region (see the discussion of the sills in Chapter 2.1b, Elliot and Fleming 2021) was clearly the centre for emplacement of the Basement Sill if not the stratigraphically higher sills as well. Although three principal centres for magma intrusion and eruption (central Transantarctic Mountains, south Victoria Land and north Victoria Land) have been suggested (Elliot and Fleming 2008), this distribution may simply reflect extant intrusive and extrusive rock outcrop. The scattered outcrops and the nature of the volcanoclastic lithofacies in the central Transantarctic Mountains and Victoria Land point to multiple local centres for the initial stages of Ferrar extrusive activity. Intervals of volcanoclastic rocks within the basalt sequences also suggest continuing local sources. The phreatocauldrons demonstrate major eruptive centres lying within the present outcrop belt from at least north Victoria Land to the central Transantarctic Mountains, and suggest that vents for the lavas were probably also located within the outcrop belt. If correct, and given that the present narrow outcrop belt is only a fraction of its original extent, then magmas must have migrated laterally away from that linear system of vents.

Ross *et al.* (2008a) noted basalt plugs at Coombs Hills and estimated potential magma eruption rates. Their calculations for a conduit with a 10 m radius suggested that the rate of magma effusion would be sufficient for construction of a typical flood basalt field. No feeder dykes or dyke swarms, such as identified for the Columbia River Basalt (Tolan *et al.* 1989; Self *et al.* 1997), have been found. However, shallow dyke patterns are complex at some sites (Airolidi *et al.* 2011, 2012, 2016; Muirhead *et al.* 2012, 2014) and have been interpreted as possible feeders. Strombolian deposits interbedded with the lavas, such as observed in West Greenland and interpreted as possible vents (Pedersen *et al.* 2017), have not been observed.

Palaeoenvironments

Several lines of evidence point to groundwater and lakes exerting significant control on both the near-surface eruptive processes for the pyroclastic rocks and the flood lavas. The phreatomagmatic deposits already noted indicate abundant groundwater in the underlying Triassic–Jurassic strata, and probably played a crucial role for the generation of the phreatocauldrons in an overall rift setting. Hyaloclastite and pillow



Fig. 19. Mount Fazio, north Victoria Land. Tree (arrow) rooted in an interbed and engulfed by lava (Elliot *et al.* 1986b; unit 4, section 82.4). In Figure 16 this locality is out of sight just to the right of the tachylite bluff. Person is for scale at the base of the tree. (Image: D.H. Elliot.)

lava intervals formed where lava entered standing water, which occurred regionally at the base of the lava stack in south Victoria Land (Carapace Nunatak and Thomas Rock), and later in the eruptive cycle at Storm Peak, Brimstone Peak and the Mesa Range. Lake beds are direct evidence of more protracted intervals of water ponding on the lava field surfaces. Conchostracans and the other invertebrates suggest relatively shallow water, and this shallowness is supported in the Marshall Mountains by a lacustrine bed that passes laterally into the weathered upper part of a lava. Plant debris associated with the weathering horizons, together with fossil logs (Storm Peak, Haban Spur and other sites) and tree stumps (Mount Fazio, Fig. 19; Haban Spur, Fig. 20), indicate flourishing vegetation between eruptive events. The flora and fauna recovered have been interpreted to suggest a temperate but strongly seasonal climate (Elliot and Hammer 1996; Garland *et al.* 2007).

Volcanic gas input to the Jurassic atmosphere from the phreatomagmatism at Allan and Coombs hills was unlikely to have been significant (Ross *et al.* 2008a), but no estimates have been made for Ferrar magmatism as a whole. Thordarson and Self (2003) have made such calculations for the historical Laki eruption in Iceland and Self *et al.* (2014) for the Roza eruption of the Columbia River Basalt province, and show the possible magnitude of such atmospheric perturbations. McElwain *et al.* (2005) speculated on the possibility that devolatilization of coals in the Gondwana sequence by Ferrar and Karoo magmatism might have been responsible for the Toarcian Oceanic Anoxic Event. High precision dating of the Ferrar province (Burgess *et al.* 2015) strengthens the possibility that Ferrar (and Karoo) magmatism is indeed related to, if not responsible for, the Toarcian event. The large sill complexes



Fig. 20. Haban Spur, north Victoria Land. The rooted tree-stump diameter is approximately 80 cm (Elliot *et al.* 1986b; base of unit 23 at 524 m in section 82-3). Hammer is for scale. (Image: D.H. Elliot.)

of the Ferrar LIP (and Karoo LIP) may have been important for that event, in an analogous way to the Siberian Traps and end-Permian extinction (Burgess *et al.* 2017), although coal beds in Antarctica were not extensive.

Future studies

Although the Ferrar LIP in Antarctica has been studied for more than 50 years, access is neither easy nor simple and much remains to be discovered. The following are suggested lines of future research that may prove particularly fruitful.

1. Can flow directions of lavas and sills be determined using AMS (as has been applied in various studies of dykes and sills) in order to establish the location of eruption centres, and whether linear or point sources?
2. What are the physical conditions and cooling patterns for thick lavas with tachylitic intervals?
3. Can lava correlations be established in the central Transantarctic Mountains and the Mesa Range that would enable the evaluation of the three-dimensional relationships of lavas?
4. Would detailed investigation of the lateral relationships in both thick lavas and sets of thin lavas confirm the applicability, to the Ferrar Province, of the general model of flood basalt field development (Self *et al.* 1997)?
5. There are many aspects of the primary volcanoclastic rocks (and deposits of material reworked from them) that merit further study. For example, what do the bedded deposits say about the distribution and nature of the volcanic province prior to eruption of the flood lavas?
6. The relationships of pyroclastic rocks at Coombs and Allan hills with country rock, sills, dykes and lavas are complex. They provide a window of fortuitous exposure that could allow assessment of near-surface (both closely below, and initial eruptive) processes taking place elsewhere in the Ferrar and other flood basalt provinces worldwide.
7. Different studies have come to strongly differing conclusions about the significance of intrusion geometry characterizing the Ferrar LIP. Do the apparently overwhelmingly predominant sills suggest a near-neutral stress regime, or is the dyke-swarm signature of an East Africa-like rift concealed in Antarctica? Were mid-crustal dykes formed which then fed upper-crustal sills and, if so, why?

Summary

The extrusive component of the Ferrar Large Igneous Province comprises an initial phreatomagmatic phase of activity followed by massive lava effusion. The early stages, involving large-scale magma–water interaction, created very large and unusual diatreme complexes (phreatocauldrons with vertical exposure of more than 300 m). Tuff rings and tuff cones that were built early were largely removed by later volcanic activity or by subsequent erosion. Locally, with diminished involvement of water, magmatic activity switched to Vulcanian and/or Strombolian. Only scattered remnants exist of any constructional volcanic topography. With exhaustion of subsurface water and/or increased magma supply rates, activity changed to the quiet effusion of flood lavas, with early but local deposition of thick hyaloclastites and pillow lavas in pre-existing topography. Lavas accumulated during a relatively short time interval to form sequences as much as 750 m thick and constructed from as many as 41 lavas, although commonly much fewer. Lavas in excess of 100 m in thickness are present in most measured sections. The internal features of the individual lavas are typical of flood lavas. Lacustrine interbeds and palaeosols mark more substantial breaks in eruptive activity.

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