Physical volcanology of continental large igneous provinces: update and review

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Abstract: Large igneous provinces (LIPs) form in both oceanic and continental settings by the emplacement and eruption of voluminous magmas ranging from basalt to rhyolite in composition. Continental flood basalt provinces are the best studied LIPs and consist of crustal intrusive systems, extensive flood lavas and ignimbrites, and mafic volcaniclastic deposits in varying proportions. Intrusive rocks are inferred to represent the solidified remnants of a plumbing system that fed eruptions at the surface, as well as themselves representing substantial accumulations of magma in the subsurface. The vast majority of intrusive rock within the upper crust is in widespread sills, the emplacement of which may structurally isolate and dismember upper crustal strata from underlying basement, as well as spawning dyke assemblages of complex geometry. Interaction of dykes and shoaling sills with near-surface aquifers is implicated in development of mafic volcaniclastic deposits which, in better-studied provinces, comprise large vent complexes and substantial primary volcaniclastic deposits. Flood lavas generally postdate and overlie mafic volcaniclastic deposits, and are emplaced as pahoehoe flows at a grand scale (up to 10^4 km^2) from eruptions lasting years to decades. As with modern Hawaiian analogues, pahoehoe flood lavas have erupted from fissure vents that sometimes show evidence of high lava fountains at times during eruption. In contrast to basaltic provinces, in which volcaniclastic deposits are significant but not dominant, silicic LIPs are dominated by deposits of explosive volcanism, although they also contain variably significant contributions from widespread lavas. Few vent sites have been identified for silicic eruptive units in LIPs, but it has been recognized that some ignimbrites have also been erupted from fissure-like vents. Although silicic LIPs are an important, albeit less common, expression of LIP events along continental margins, the large volumes of easily erodible primary volcaniclastic deposits result in these provinces also having a significant sedimentary signature in the geologic record. The inter-relationships between flood basalt lavas and volcaniclastic deposits during LIP formation can provide important constraints on the relative timings between LIP magmatism, extension, kilometre-scale uplift and palaeoenvironmental changes.

Large igneous provinces (LIPs) have been the subject of many previous papers and books, most with a petrological or geodynamic focus. The papers in this volume devoted to George Walker focus, in contrast, on physical processes of magmatism, and for LIPs a diversity of physical magmatic phenomena are known to be involved in their emplacement. George had an interest in the styles of lava that form LIPs and his early work was influential – including his Deccan Traps-based paper that proposed compound v. simple flows (Walker 1972, 1999). In this article, we update and review aspects of physical volcanology for continental basaltic and silicic LIPs. For basaltic continental

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LIPs, we assess the hypabyssal magma distribution system for eruptions, the emplacement of extensive basaltic lava flows, and the extent and significance of mafic volcaniclastic deposits accompanying flood lavas. Silicic LIPs are dominated by pyroclastic deposits but in contrast to the basaltic examples, their plumbing systems are less well exposed and studied. We conclude with a brief evaluation of the context for physical volcanological studies in LIPs, and a summary of key volcanological processes active during their emplacement.

Magma distribution systems: dykes and sills of continental LIPs

Although the most prominent and longest studied rocks of continental large igneous provinces are thick stacks of basaltic lavas, the first section of this manuscript addresses the solidified lithospheric magma distribution systems that fed the lavas. These 'plumbing systems' are represented by extensive sills and dykes, now exposed at different levels in variously eroded provinces (e.g. Richey 1948; Ernst & Baragar 1992; Tegner *et al.* 1998; Chevallier & Woodward 1999; Elliot & Fleming 2004). Giant dyke swarms and other intrusions inferred to have been coupled with surface eruptions are exposed in deeply eroded continental provinces (Piccirillo *et al.* 1990; Ernst & Baragar 1992; Hatton & Schweizer 1995; Ernst & Buchan 1997, 2001; Ernst *et al.* 2005; Ray *et al.* 2007), whereas a range of intrusive complexes, sill networks and populations of smaller dykes are known from settings within a few kilometres of the palaeoeruption surface.

Whatever the origin of LIP magmas or the tectonic regime associated with their emplacement, the resulting intrusive rocks represent substantial volumes of unerupted magma (Crisp 1984; Walker 1993). The underplated igneous volume can be up to 10 times larger than the associated extrusive volume. For example, in the North Atlantic Igneous Province, Roberts *et al.* (1984) estimated the total volume of Palaeocene to early Eocene basalt to be 2 million km^3, whereas White *et al.* (1987) and White & McKenzie (1989) suggested a total volume of up to 10 million km^3, and Eldholm & Grue (1994) estimated a total crustal volume of 6.6 million km^3.

Magma that solidified in sills, dykes and other intrusive complexes developed in host rocks as a result of mechanical coupling between magmatic pressure and the stress regime extant during their emplacement (Anderson 1951; Rubin 1995). Assuming that dyke–sill orientations reflect deformation in homogeneous media at crustal or lithospheric scales, the geometries of the solidified magmatic plumbing networks have been used to infer stress regimes during emplacement, and to infer tectonic context and magma origin (Wilson 1993; Head & Kreslavsky 2002; Wilson & Head 2002; Ernst & Desnoyers 2004; Elliot & Fleming 2004). The nature of magma transport at depth is not, however, readily determined in regions where only shallower exposure exists, as illustrated by the range of possibilities considered by Elliot & Fleming (2004) for delivery of magma to the Ferrar Group intrusions and flood basalts in Antarctica (Fig. 1). This uncertainty makes it more challenging to determine the ultimate sources of magma for various LIPs, whether it is generated in linear zones below eruption fissures or distributed along such zones over large distances from a central source (e.g. MacKenzie dyke swarm; Baragar *et al.* 1996).

Given the dynamics of magma intrusion and structural decoupling of strata buoyed above extensive sills, it may not be valid to assume that dyke–sill orientations can be used directly to infer regional tectonic stresses. This may be particularly relevant for LIPs characterized by widespread and voluminous sills, such as the Karoo Dolerite of southern Africa (Fig. 2) and its spectacularly exposed Antarctic counterpart, the Ferrar Dolerite. Consider the enormous Peneplain Sill in the Dry Valleys, Antarctica (19 000 km^2, 0.25 km thick), which was intruded beneath *c*. 2 km of sedimentary rock (Gunn & Warren 1962). Had the sill been intruded 'instantaneously', the overlying sedimentary rock would have been decoupled from underlying basement rock by a liquid–plastic layer of magma; the lid would have been isolated from any tectonic stress exerted on the rocks below. Emplacement is not instantaneous, but sills maintain deformable interiors during emplacement (Marsh 1996), which limits mechanical coupling through them (Hawkesworth *et al.* 2000; Marsh 2004). Also, sills grow by fluid-dynamic insertion of magma which, under triaxial stress regimes and into homogeneous or simply layered host rocks, produces saucer-shaped or stepped-saucer sills (Chevallier & Woodward 1999; Malthe-Sørenssen *et al.* 2004). As a sill spreads from a magma supply site, the rock above is progressively wedged and buoyed upward (Chevallier & Woodward 1999; Thomson & Hutton 2004). This process transmits stress through the uplifting rock, and cracks thus created are filled by magma to produce dykes (Pollard & Johnson 1973; White *et al.* 2005). Dykes spawned in this way reflect near-field stresses from the intrusion process itself, rather than far-field tectonic stresses affecting the crust below the sill.

In South Victoria Land, Antarctica, many Ferrar intrusions change their shape and orientation along their length; horizontal sills locally feed into subvertical dykes, dykes change strike abruptly and the

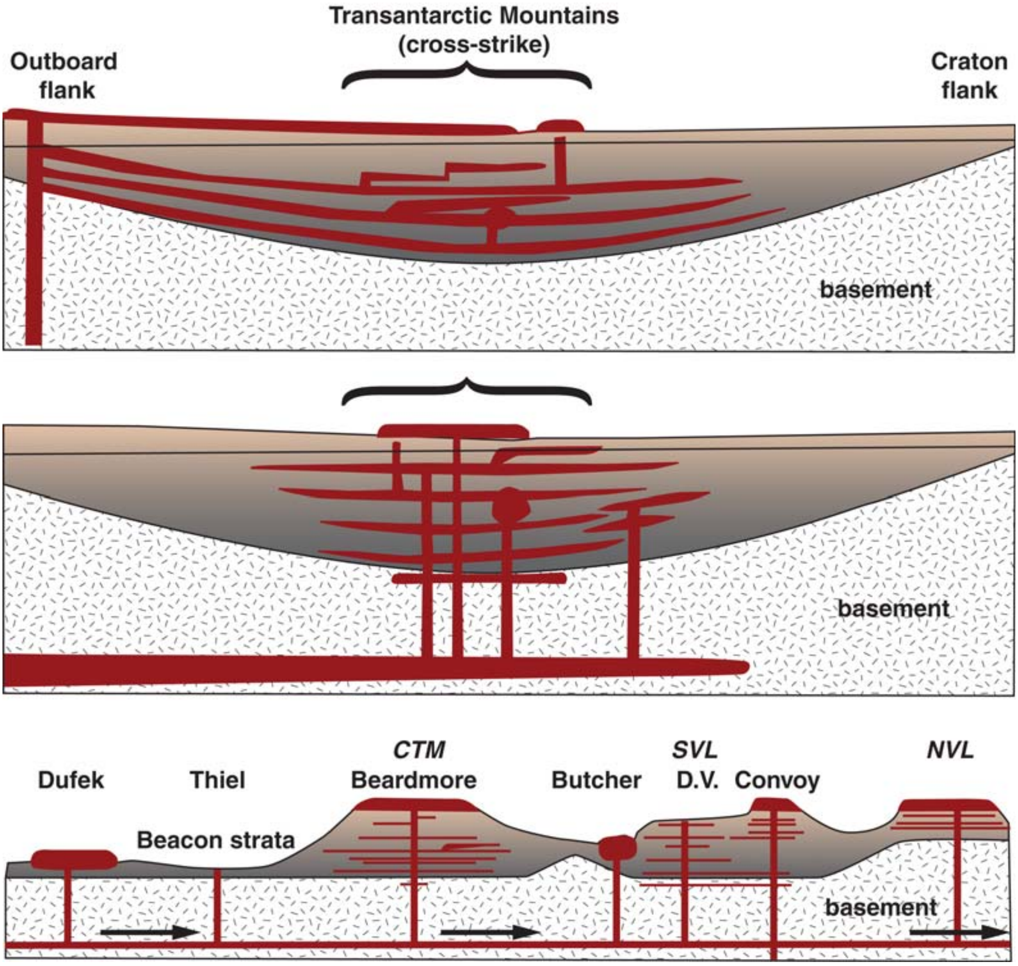

Fig. 1. Different styles of long-distance magma transport considered by Elliot and Fleming (2004; diagram redrawn from their fig. 7) for the distribution of Weddell Sea-derived magmas throughout the Ferrar LIP. In the top two cartoons, magma feeding the Ferrar Dolerite, exposed in the Transantarctic Mountains, is provided by sills extending cratonward from a megadyke farther toward an outboard convergent margin, whereas in the lower cartoon the main transport is in a megadyke or dyke complex beneath the current outcrop belt, with delivery toward the surface by vertical dyking.

form of the fractures occupied by magma varies widely across small areas (Elliot & Fleming 2004). Outcrops at Mount Gran and Terra Cotta Mountain (Fig. 3) illustrate this complexity at paeleodepths of *c*. 1–2.5 km, which is somewhat unexpected, because magma transport at such depths has been treated as being controlled predominantly by the regional stress field.

Terra Cotta Mountain exposes rocks from *c*. 1–2 km below the surface, including a sill separating basement from sedimentary cover rocks, a large 'mega' dyke, and swarms of subparallel to suborthogonal inclined sheets of varying thickness. Higher in the overall sequence, a spectacular cliff at Mount Gran similarly exposes a megadyke at the local terminus of a thick sill, with the megadyke apparently feeding a splay of inclined sheets (Fig. 3b). At both sites, significant displacements of country rock take place across thick dykes, at least partly in response to differential jacking up of strata by sills that terminate at these dykes.

The rather chaotic pattern seen at both Mt Gran and Terra Cotta Mountain is more consistent with country rock acting as a 'floated' lid on top of, or partly within, a fluid magma, with cracks forming in response to very local stresses rather than regional ones (White *et al.* 2005). This conclusion gains additional support from the contact geometries of

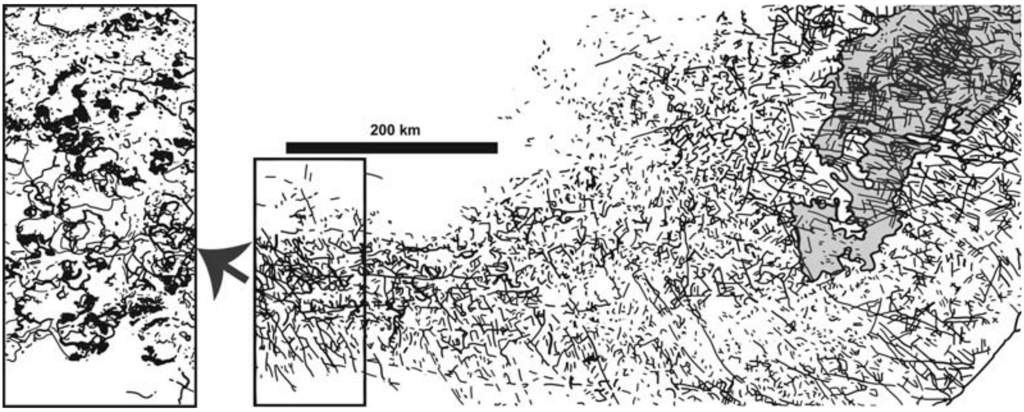

Fig. 2. Illustration of regional and local patterns of intrusion in the Karoo LIP, redrawn after Chevallier & Woodward (1999). The intruded strata are predominantly mudrock and minor sandstone of the lower Beaufort Group (Johnson *et al.* 1996), and were probably intruded at depths of several kilometres below the pre-flood basalt surface. The shaded area to the northeast represents outcrops of the Stormberg lavas, and dykes mapped there are exposed within the flood-basalt sequence. The enlargement of the boxed area shows in more detail the outcrop pattern and the abundance of sills (thick curved lines), many of which form broadly dish-shaped structures. Note that the abundant approximately linear dykes in the simplified regional illustration are not apparent at this scale (drawn from 1:50 000 maps), which instead displays many curved and irregular dykes with only weak, segmented, linearity. The regional map showing rectilinear dykes demonstrates well the extent and intensity of subvolcanic intrusion, but fails to capture the chaotic and irregular pattern of intrusion apparent at larger mapping scales.

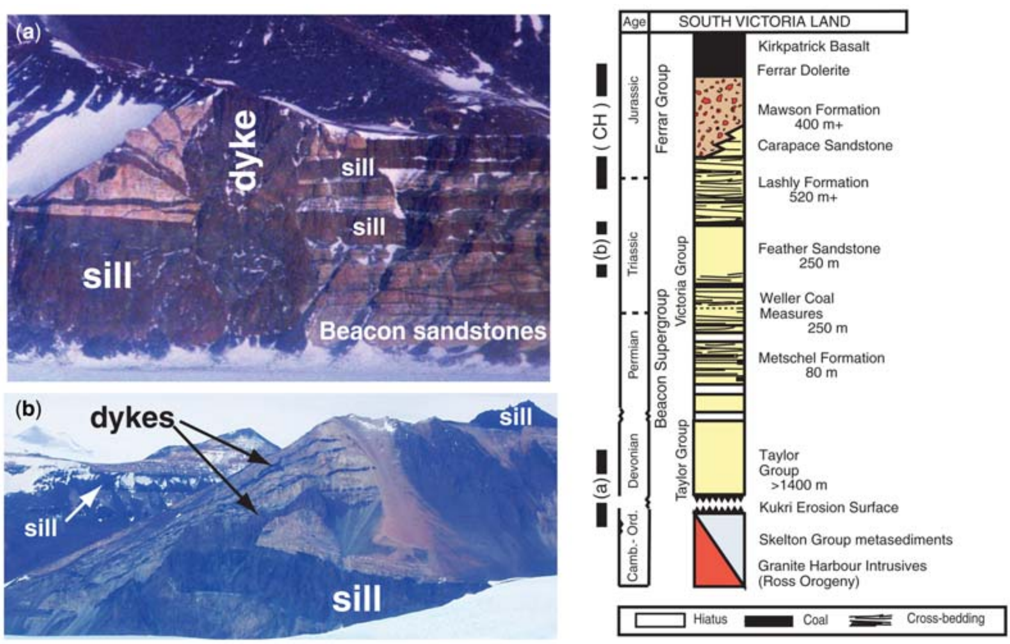

Fig. 3. Simplified stratigraphic column for South Victoria Land, Ferrar LIP (right) shows approximate stratigraphic levels of dyke and sill outcrops shown. In outcrops shown, dark rock is dolerite, country rock is pale sandstone. (**a**) Terra Cotta Mountain, *c*. 800 m topography, with basement exposed below sill at lower right, and (**b**) Mt Gran, cliff height *c*. 400 m, with apparent transition from large sill at left to central mega-dyke; inclined sheets extend from the mega-dyke, and thinner sills are exposed to the right. 'CH' indicates stratigraphic range of outcrops of Coombs Hills (Fig. 4). Column after McClintock (2001), Elliot (1992), Collinson *et al.* (1983) and Ballance (1977).

some dykes, which show irregular buds and extensions that indicate different directions of magma flow and/or of dyke propagation among closely spaced dykes. Local regularities in dyke-set geometries may reflect intrusion dynamics, with wedging and uplift during initial sill propagation causing systematic cracking of the floated lid. Other dyke-set patterns may result from inhomogeneities in the country rock caused by jointing, fracturing, faulting or folding, interlayering of rock units with contrasting rheologies, or the presence of older intrusive rocks.

At Coombs Hills, Ferrar Dolerite outcrops extend to within 200 m of the base of nearby flood lavas adjacent to the Coombs Hills vent complex (Ross *et al.* 2008). At this level, large domains of country rock are isolated and tilted within dolerite bodies (Fig. 4a), and the country rock domains are additionally cut in complex patterns (Fig. 4b) by both wedge-shaped, and small, commonly sinuous, dykes (White & Garland 2007). It appears that as Ferrar sills approached the ground surface at Coombs Hills, at least a hundred metres of overlying country rock was broken into blocks that became separated from one another, commonly rotated and partly to wholly engulfed in incrementally inflated doleritic sills. An absence of flood basalt lavas among the large tilted blocks suggests that at Coombs Hills the process predated emplacement of overlying flood basalts (White *et al.* 2006). Such wholesale breakup of country rock at shallow intrusion levels may well be related to development of vent complexes such as that at Coombs Hills, but the nature of this relationship remains to be determined.

Extensive dyke swarms are exposed in the more deeply eroded silicic LIPs (e.g. Whitsunday, Kennedy–Connors–Auburn; Ewart *et al.* 1992; Stephenson 1990; Bryan 2007). Diffuse swarms ($\geq$100 km wide) of mostly steeply dipping dykes ranging between 1 and 50 m in width are characteristic, and the swarms can extend along strike for over 1000 km (Stephenson 1990). Silicic LIPs may have similarly extensive mid- to upper-crustal granitic batholith underpinnings and dyke swarms, and more-mafic igneous underplate at lower crustal depths (Ferrari *et al.* 2007), but our understanding of the magma plumbing systems of silicic LIPs remains limited in comparison to what is known for CFBPs.

In summary, the emplacement of substantial volumes of magma generated in continental LIPs may be solely as intrusions at relatively shallow depths (upper few kilometres). The geometry and emplacement processes of these intrusions are controlled by the interaction between magma fluid dynamics and tectonic stresses. As large intrusions approach the ground surface, overlying rocks can be broken apart and effectively engulfed within them.

Lava flows in continental flood basalt provinces

In less-eroded continental flood basalt provinces (CFBPs), very thick piles of basaltic lava flows (more than 3 km thick in some cases) are seen to make up the bulk of each province. Although they have been studied since the inception of geology as a science (see Walker 1995 for a review), the flows are so extensive, and the flood basalt provinces so widespread and generally broken up by rifting, that it has taken much painstaking work to piece together a picture of the ‘typical’ product of a flood basalt eruption. The most valuable work so far for flood basalt interpretation has been based upon the Columbia River Basalt province, the smallest, youngest, and arguably the most intact CFBP. Decades of effort by many workers, summarised in papers such as Tolan *et al.* (1989), Reidel *et al.* (1989) and Reidel & Hooper (1989), show that individual flow fields, each the product of a single eruption, are huge in volume, commonly exceeding 1000 km^3 of lava. Furthermore, volcanological studies show that these eruptions were fed by very long fissures (e.g. Swanson *et al.* 1975) and that at least parts of the eruptions were Hawaiian-like in nature at the vent (Reidel & Tolan 1992), featuring small lava ponds.

Lava flow fields

The lava piles in CFBPs are composed of flow fields almost always of pahoehoe or rubbly pahoehoe (as in more-modern lava fields, Guilbaud *et al.* 2005), with the latter forming up to 30% of flows in some provinces (Keszthelyi 2002; Keszthelyi *et al.* 2004). The flow fields have been proposed to have originated from prolonged eruptions that probably lasted for years to decades (Self *et al.* 1996, 1997, 1998; Thordarson & Self 1996, 1998). Each flow field consists of several major lava flows, which in turn consist of multiple flow lobes. The number of individual lava bodies within one flow field must be very large indeed. The major sheet lobes, containing the majority of the lava volume, are commonly 20–30 m thick, several kilometres wide, and show features consistent with *in situ* flow thickening by inflation (Hon *et al.* 1994; endogenous growth; Fig. 5). These extensive lobes, with aspect ratios (length/thickness) ranging from *c.* 50 to *c.* 500, are the basic building-blocks of a CFBP and give the provinces their ‘layer-cake’ or, when eroded, step-like, appearance. The similarity of processes that form sheet lobes within any CFBP gives

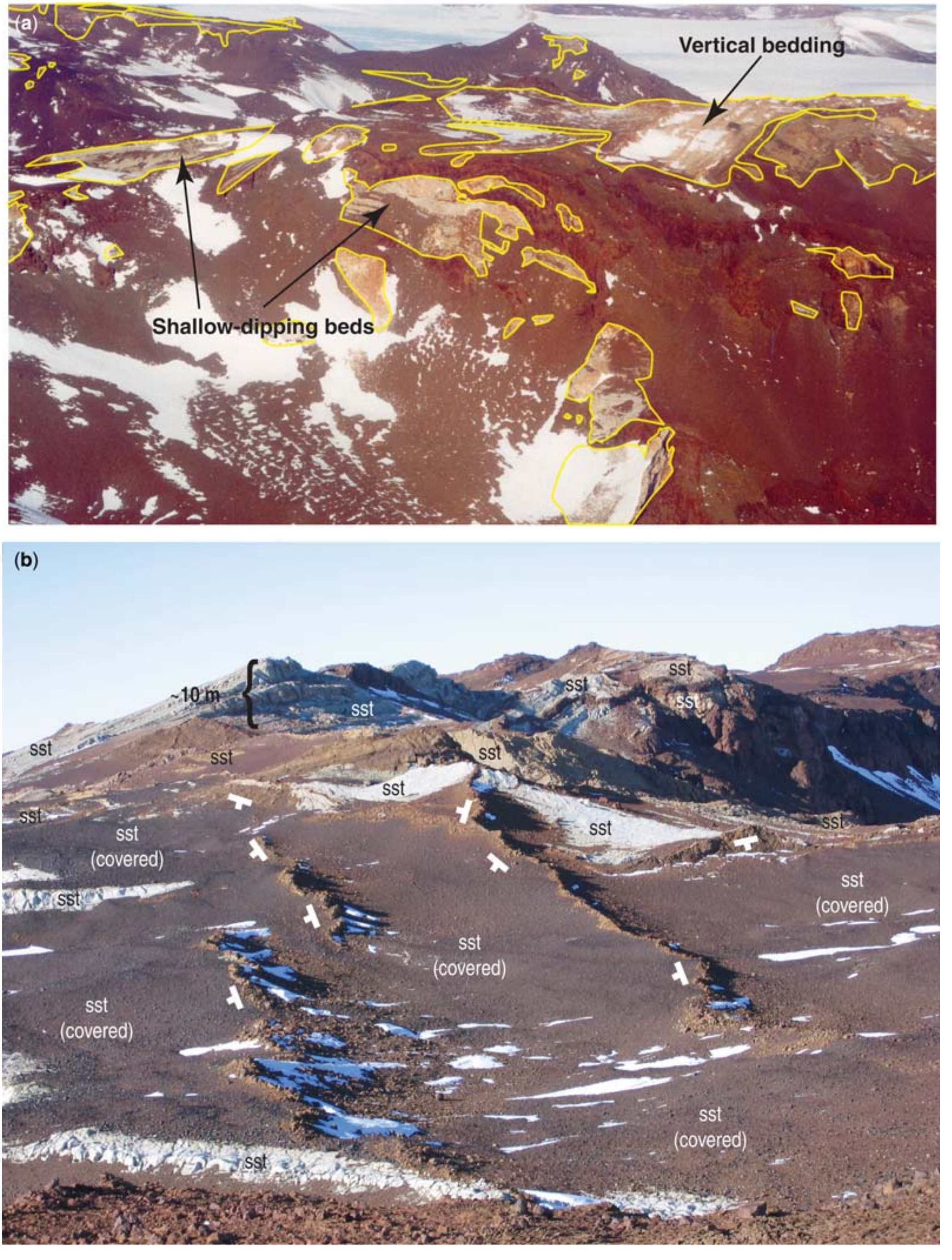

Fig. 4. (**a**) Northern Coombs Hills. Note large, variably tilted, blocks of sandstone enclosed in dolerite; *c*. 200 m relief in image. The Coombs Hills outcrops are adjacent to a large vent complex (see Ross *et al.* 2008 and references therein). (**b**) Detail of dolerite–sandstone relationships illustrated in (a). Numerous dykes and inclined sheets, some dipping at very low angles, penetrate and separate bodies of sandstone that are tilted in varying directions from their *in situ* orientations. The sandstone bodies at this level (*c*. 200 m below nearby base of Ferrar Group flood-basalt lavas) are also penetrated by scores of thin, sinuous dykes that commonly terminate within the sediment as thin pointed tips a few millimetres wide. The dolerite cliff in shadow is the edge of a sill; its contacts against bodies of country rock to the left are subvertical, but it has distinct subhorizontal internal boundaries defined by weathering and, locally, thin lenses of country rock.

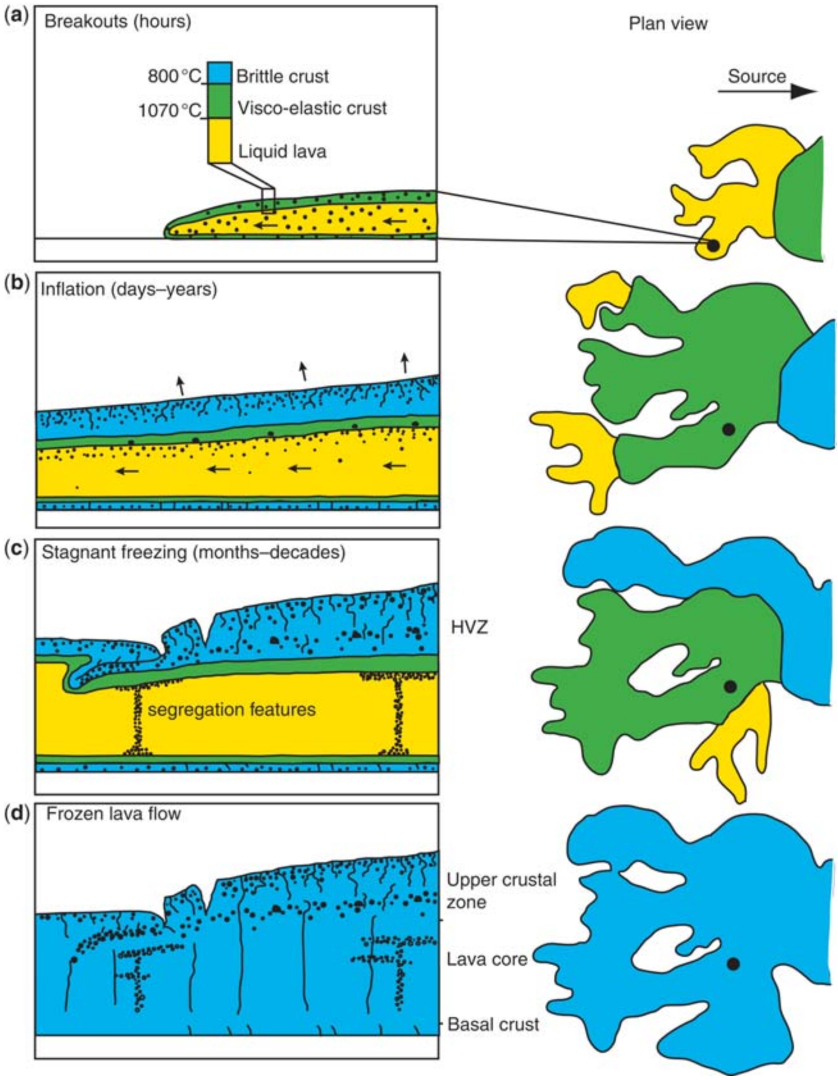

Fig. 5. A four-stage diagram illustrating emplacement of lava by lobes and lobe-breakouts.

a consistent internal structure to the lava units that is typified by the schematic section shown in Figure 6. Lava units are usually pipe-vesicle-bearing pahoehoe lavas (Fig. 7), and the often complexly jointed vesicular upper crustal zone can occupy 40–60% of the sheet lobe thickness.

The morphology of the lava bodies, their surface characteristics and internal textures appear to change little from vent to toe in flood basalt flow fields, which can extend over distances in excess of 500 or even 1000 km (Hooper 1996; Self *et al.* 2008). Proximal lava flows tend to be thinner than the thick sheet lobes that occupy the almost infinitely low slopes of the main parts of a province. *Sheet lobe* refers to a single flow lobe that is a large-scale feature, i.e. wider than an outcrop, and tens of metres thick, as is common in CFBPs. This same basic volcanic architecture has been reported from each flood basalt province where physical descriptions of the lavas have been made, including the Kerguelen plateau (Keszthelyi 2002), Etendeka (Jerram 2002), North Atlantic Igneous Province (Single & Jerram 2004) and Deccan (Bondre *et al.* 2000, 2004*a*, *b*; see review by Jerram & Widdowson 2005). A common variant is flows with pahoehoe bases and internal structures capped by a rubbly top often over 10 m in thickness. These so-called rubbly pahoehoe lava units have been described

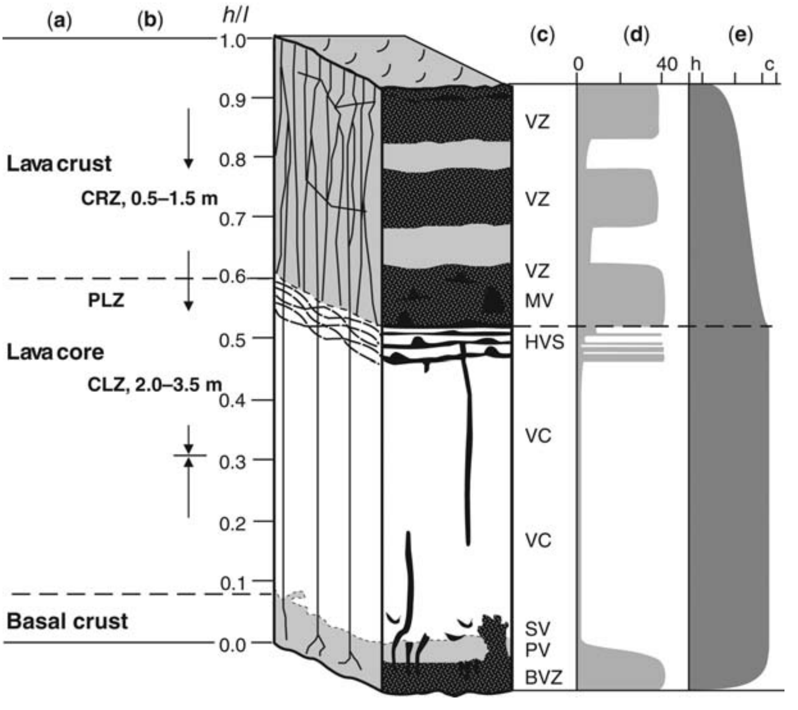

Fig. 6. A composite graphic log showing illustrating characteristic structures of Roza sheet lobes. Left side shows the characteristic three-part division of sheet lobes (**a**) and jointing styles (**b**). Right side of the column shows distribution of vesiculation structures (**c**), vesiculation (**d**) and degree of crystallinity (**e**). The scale h/l indicates normalized height above the base of the sheet lobe (h, height in lobe; l, total lobe thickness). Abbreviations in column (a) are: CRZ = zone of crustal joints, PLZ = zone of platy joints, CLZ = zone of columnar joints. The structures in column (c) are BVZ = basal vesicular zone, PV = pipe vesicles, SV = sheet vesicles, VC = vesicle cylinders. Scale on column (d) is d, dense (0–5 vol% vesicles); m, moderately vesicular (10–20 vol%); and v, vesicular (30–40 vol%). On column (e) h, hyaline; hy, hypohyaline; hc, hypocrystalline; c, holocrystalline.

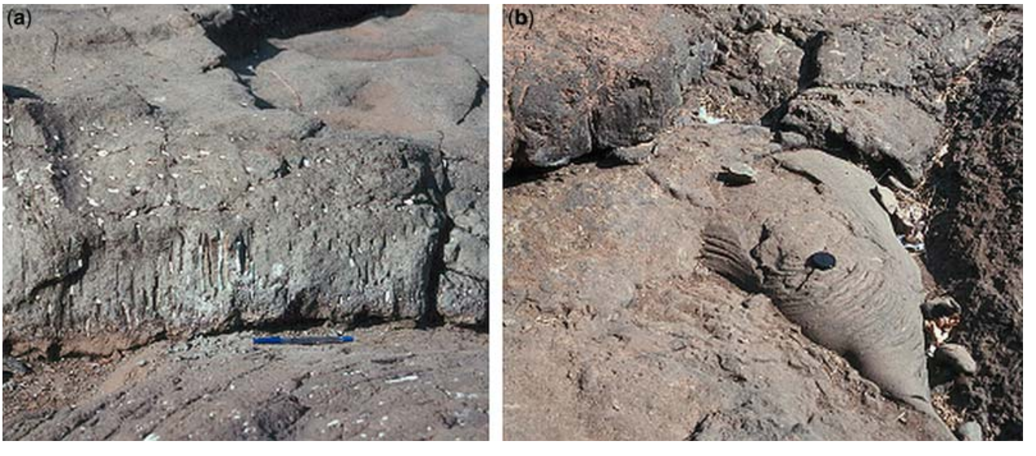

Fig. 7. Two photos of 'Hawaii-size' Deccan pahoehoe lobes, Bushe Formation, Lonavala Sub-group, near Poladpur, India. (**a**) base of a decimetre-thick lobe with pipe-vesicles – pen for scale; (**b**) small lobe with ropes, lenscap for scale.

from smaller, more recent flood lava flow-fields (Laki, Iceland; Guilbaud *et al.* 2005) and CFBPs (Kerguelen, Columbia River Basalts, see references above), and some slabby-topped flows have been described from the Deccan province (Duraiswami *et al.* 2003).

Both geothermometry and thermal modelling of Columbia River lava flows show that the great extent of individual lava flows was not limited by cooling (Keszthelyi *et al.* 2004). The insulated transport of lava under a thick crust is thermally extremely efficient, with measured cooling rates of $\ll 0.1\ ^{\circ}C$ per km flowed (Ho & Cashman 1997; Thordarson & Self 1998). Theoretically, this mode of emplacement can produce lava flows >1000 km long with modest lava fluxes (Self *et al.* 2008). In a study of long Quaternary basaltic lava flows (>100 km long) of the McBride province in northern Australia, it was concluded that lava flows greater than tens of kilometres were favoured by a pahoehoe emplacement style (thermal insulation), sustained eruption over years to tens of years, favourable slopes, unhindered flow conditions (e.g. dry river beds) and an insulated conduit system (lava tubes), but that lava flow size was ultimately limited by supply (Stephenson *et al.* 1998). Thus the key to these 'floods' of lava is the immense volume of magma released during one eruption, rather than the lava viscosity, eruption rate or environmental conditions (Keszthelyi & Self 1998). Recent studies of lavas in the Deccan Traps show that inflated pahoehoe lavas are common in that province, with the implication that insulated transport also played an important role their emplacement (e.g. Keszthelyi *et al.* 1999; Duraiswami *et al.* 2001, 2002, 2003; Bondre *et al.* 2004*a*; Jay 2005). Many details are still not available, however, and different CFBPs may have distinct lava characteristics (Bondre *et al.* 2004*a*, *b*).

Flood basalt vents

Important additional information about the nature of flood basalt eruptions can be gleaned from the nature of the vents, although the number of reports of vent facies from CFBPs is very small. This partly reflects that fact that the vent systems are small, often-linear components in huge lava provinces, so there is low probability that they will be commonly exposed. Moreover, in many CFBPs rifting may occur along the trends of earlier fissure–vent systems, perhaps preferentially destroying evidence of the vent regions (Hooper 1990). The occurrence of dyke systems that can be traced laterally for tens to hundreds of kilometres implies that many flood basalt eruptions are fed from linear vent systems. The fissures appear to cluster in time and space, such that one lava formation within a CFBP is often erupted from a group of sub-parallel linear-vent systems, represented by dyke groupings (Hooper 1990; Walker 1995). The best documented examples of the apparently rarely outcropping surface-vent constructs and vent successions are found within the Columbia River Basalts (Swanson *et al.* 1975), where associations of fountain-fed flows, spatter and lapilli scoria units, and spongy and shelly pahoehoe lobes appear to define complex linear vent systems that bear a resemblance to the cone complexes formed by modern-day fissure eruptions. One in-depth study of a vent structure within the *c.* 16 Ma old Teepee Butte Member of the Grande Ronde basalts shows that it featured a lava pond surrounded by cone ramparts that were constructed by at least three distinct episodes of Hawaiian-style fountaining (Reidel & Tolan 1992). An important consideration is that evidence preserved around basaltic vents may represent processes occurring in the dying stages of vents and fissure segments, and not what was occurring during the periods of maximum effusion rate.

The near-vent succession of the *c.* 14.7 Ma old Roza Member is a sequence of fountain-fed lava flows overlain by 1–10 m thick bedded lapilli scoria units, which in turn are capped by either fountain-fed lava or pahoehoe sheet lobes (Thordarson 1995; Thordarson & Self 1996). The scoria units are of particular interest as they consist of uniform fine to medium lapilli scoria and exhibit a near-horizontal internal bedding. This sequence of fountain-fed lavas, scoria beds and 'normal' lavas is identical to that found in the near-vent successions of the 1783–1784 AD Laki and 934–940 AD Eldgjá fissure-fed flood lava eruptions (Fig. 8a), where each sequence is the product of one eruption episode. The resemblance is also enhanced by the similar grain-size distributions and clast morphologies in these deposits to those found in flood basalt provinces. In the historic eruptions, the scoria units were produced by sub-Plinian explosive phases at the beginnings of individual eruption episodes because, at the onset of degassing, the exsolved volatiles streamed up through the magma column to form a two-phase flow in the upper part of the conduit. This resulted in gas-driven explosive eruptions at the surface that produced the early fountain-fed flows. Because of this initial bulk loss of volatiles the conduit flow was converted to the bubbly flow regime and the style of the eruption changed to weak fountaining and effusion of normal lava. This pattern then repeated in the subsequent eruption episodes. Although the tephra falls produced by the explosive phases in flood basalt volcanism are minor components compared with the volume of lava erupted, the significance of these phases for assessing eruption dynamics

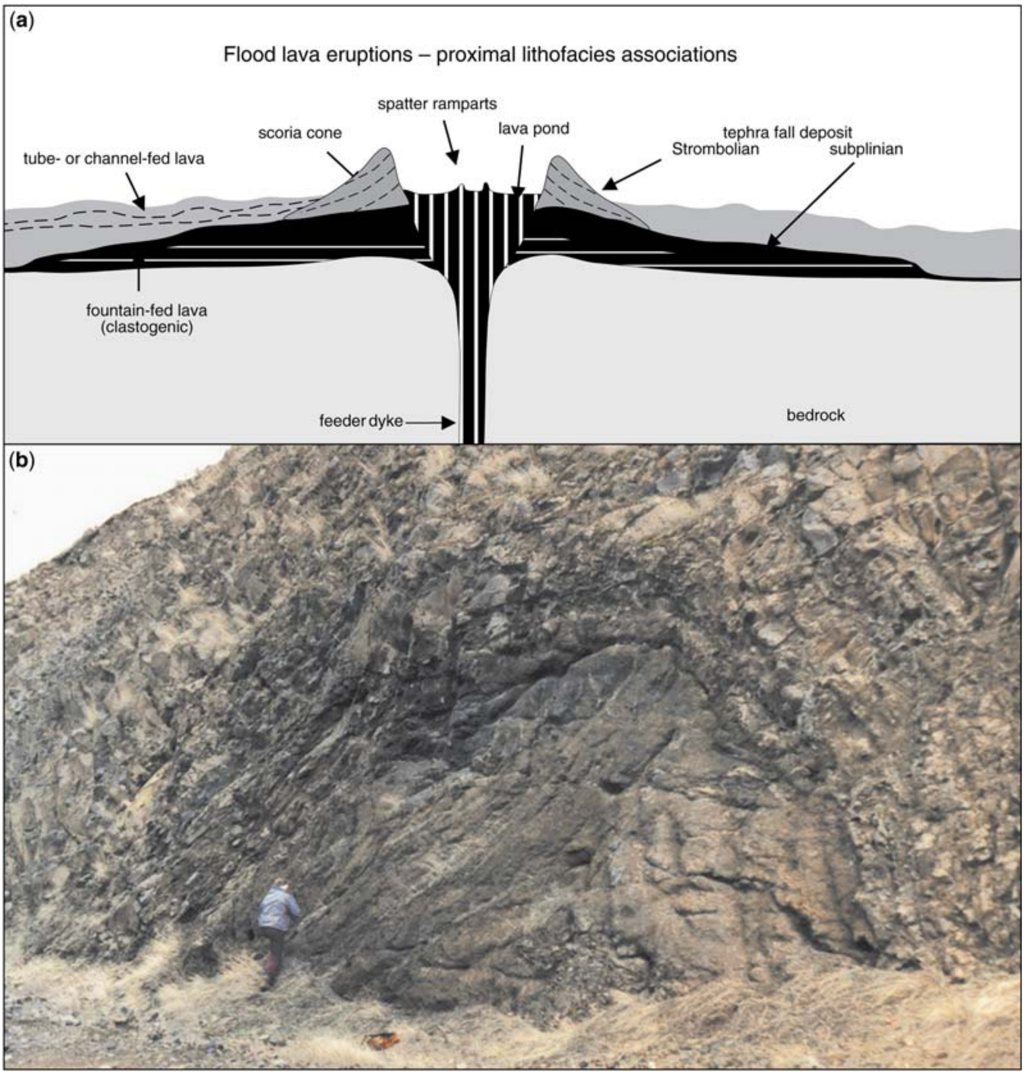

Fig. 8. (**a**) Schematic illustration showing the stratigraphy of the near-vent successions produced by a single eruption episode during the Laki and Eldgjá eruptions. The diagram is unscaled, but spatter ramparts are metres to a few tens of metres in scale typical scoria cone crater widths and heights are a few hundred metres or less, and tephra fall deposits of significant thickness extend only a few km from the vent. (**b**) Exposure of near-vent Roza eruption products at Winona, Washington, consisting of a scoria-fall mound or rampart overlain by agglutinated spatter-fall facies (person for scale).

and possible atmospheric effects should not be underestimated (Self *et al.* 2005). They record periods of peak magma discharge that produced eruptions of sub-Plinian intensities.

The picture that is emerging conforms well to the notion that these fissure-fed lava-producing events are large-scale versions of the historic flood lava eruptions in Iceland. Thus by analogy, it is likely that flood basalt eruptions featured multiple episodes, each beginning with a relatively short-lived explosive phase followed by a longer-lasting effusive phase. After fissure activity, effusion may have settled down to one of a few points along the linear vent system, and small shields (and cones; Fig. 8b) are known along the Roza fissure of the Columbia River province (Swanson *et al.* 1975). Another important conclusion that can be drawn from this comparison is that it is unlikely that the entire vent system erupted concurrently. It is more likely that at any one time the activity was confined to distinct fissure segments on the vent system, as indicated by mapping of the Roza lava flow field (Thordarson & Self 1998). The lava flow-fields also grew incrementally, with active lava emplacement in only one part, or a few parts, of the whole field at any one time. It should be noted that ten years of effusion at the maximum sustained Laki eruption rate (estimated at *c.* 4000 $m^3 s^{-1}$; Thordarson & Self 1993) would yield a 'flood-basalt-magnitude' flow field (*c.* 1250 km^3). We also note that the Laki flow field has been shown to contain abundant lava tubes in the proximal to medial regions (Wood & Watts 2002), and can be considered, in the current state of knowledge, largely a tube-fed lava field.

Volcaniclastic rocks in LIPs

Not all the eruptions of LIPs were predominantly effusive. A range of volcaniclastic deposits, in addition to the informative but relatively small-volume tephra falls of sub-Plinian eruptions mentioned above, are found. The various volcaniclastic deposits contain information on erupted magma compositions, primary fragmentation mechanisms, eruptive processes, depositional environments and tectonomagmatic evolution. This section reviews the main characteristics and proposed origins of volcaniclastic rocks associated with LIPs.

We follow White & Houghton (2006) in defining primary volcaniclastic deposits and rocks as 'the entire range of fragmental products deposited directly, by explosive or effusive eruption'. In this classification, 'primary volcaniclastic' replaces the broadest use of 'pyroclastic' in Fisher & Schmincke (1984) as the core term for the family of particles and deposits formed by volcanic eruptions. Primary volcaniclastic deposits may include older fragments ejected or moved during an eruption. Reworked volcaniclastic deposits refer to those comprising particles that have been derived from primary volcaniclastic deposits and redeposited by surface processes (e.g. wind, rivers, non-eruptive density currents, ocean currents), either during an eruption or after a storage period (syn-eruptive resedimented volcaniclastic deposits of McPhie *et al.* 1993). Epiclastic deposits or volcanogenic sedimentary rocks (McPhie *et al.* 1993) are those produced by weathering and erosion of volcanic (including lithified volcaniclastic) rocks, and their rates of production are hence controlled largely by weathering. White & Houghton (2006) recognized four end-member groups of primary volcaniclastic deposits: (1) pyroclastic deposits from pyroclastic plumes and jets or pyroclastic density currents; (2) autoclastic deposits formed when effusing magma cools by contact with air and the fragments produced accumulate to produce approximately *in situ* deposits; (3) hyaloclastic deposits (hyaloclastite) and pillow breccia formed when magma effuses subaqueously, is quenched in contact with water and produces fragments accumulated as approximately *in situ* deposits; and (4) peperite, formed during shallow intrusion of magma into a clastic host; fragments of magma or lava form by mingling with the debris (typically wet), with deposition effectively *in situ* (e.g. White *et al.* 2000).

Pyroclastic deposits thus defined may comprise fragments produced by both phreatomagmatic and magmatic fragmentation processes. We therefore use 'magmatic' fragmentation (e.g. Houghton & Wilson 1989) to describe fragmentation occurring within the conduit due to gas expansion and/or magma shear, without the influence of external water.

Mafic volcaniclastic rocks in LIPs

Kilometre-thick piles of basaltic lava are not only characteristic of continental flood basalt provinces but also characterize other LIP-types such as volcanic passive margins and oceanic plateaus (e.g. Coffin & Eldholm 1994; Menzies *et al.* 2002; Kerr 2003; Kerr & Mahoney 2007). Volcaniclastic deposits constitute significant stratigraphic thicknesses and volumes of several mafic LIPs (Bryan *et al.* 2002; Ross *et al.* 2005; Ukstins-Peate *et al.* 2005), but are reportedly sparse for others (e.g. Columbia River, Deccan) or have been misinterpreted (e.g. as epiclastic alluvial fan deposits in the Emeishan flood basalt province; He *et al.* 2003). Mafic volcaniclastic deposits have been a relatively neglected research topic thus far, despite their implications for palaeoenvironmental reconstructions, magma production and supply rates, eruption dynamics and climatic impacts.

Mafic volcaniclastic deposits, now mostly lithified, exist in various proportions in most Phanerozoic CFBPs (Table 1), as well as in some Precambrian examples (Blake 2001) and in silicic LIPs (Pankhurst *et al.* 1998; Bryan *et al.* 2000). Several hundred metres of mafic volcaniclastic deposits have also been found in one drill hole from the Ontong Java Plateau, a largely submarine LIP (Shipboard Scientific Party 2001; Thordarson 2004). Salient points, in part drawn from the review paper by Ross *et al.* (2005), include the following:

1. Mafic volcaniclastic deposits occur principally as intercalated horizons among lava flows in some flood basalt provinces (e.g. Vøring Plateau in the North Atlantic; Noril'sk area on Siberian platform; Deccan plateau), whereas in others they are concentrated in the lower part of the volcanic stratigraphy

Table 1. *Summary of mafic volcaniclastic deposits from Phanerozoic mafic LIPs (continental flood basalt provinces except the Ontong Java), after Ross* et al. *(2005) and references therein*

Province	Features
Columbia River Basalts	• Possible phreatomagmatic vent infills (Fuller 1928) • Pillow–palagonite complexes, common at base of lava flows, especially near plateau margins (Swanson & Wright 1981) • Proximal pyroclastic accumulations near linear vents (Swanson *et al.* 1975; Thordarson & Self 1996)
Afro-Arabia	• Mafic mega-breccia in upper part of sequence in Yemen (Ukstins Peate *et al.* 2005) • Primary deposits intercatated with earliest lavas in Ethiopia and Yemen • Mafic volcaniclastic deposits overlying basement in NE Yemen Plateau
Deccan Traps	• Fine-grained mafic material in clastic layers between flood lavas ('intertrappean' beds; e.g. Widdowson *et al.* 1997) • Both fine and coarse mafic volcaniclastic deposits reported from base of lava sequence in Mumbai region • Thick late Cretaceous deposits in Pakistan (Khan *et al.* 1999)
North Atlantic Igneous Province	• East Greenland: near the coast, mafic volcaniclastic deposits constitute 35–50% of the lower volcanic rocks, much less in the overlying plateau lavas (Ukstins Peate *et al.* 2003*a*) • Faeroe Islands: over 1100 m of mafic volcaniclastic deposits underneath the flood lavas in one drill hole (Ellis *et al.* 2002) • North Sea and Denmark: over 130 basaltic tephra layers intercalated in sediments of the Balder Formation and correlatives (Larsen *et al.* 2003), possibly causing early Eocene cooling (Jolley & Widdowson 2005) • Ireland: numerous exposures of vent-filling breccia containing basaltic clasts near the Giant's Causeway (Patterson 1963) • Several other occurrences in other areas
Ontong Java	• Over 300 m of mafic volcaniclastic deposits, rich in accretionary lapilli, occur in one drill hole without any overlying lavas (Thordarson 2004)
Ferrar	• Flood lavas are almost everywhere underlain by mafic volcaniclastic deposits ranging in exposed thickness from 10 m to over 400 m, interpreted as phreatomagmatic deposits (Elliot & Fleming 2008) • Overall mafic volcaniclastic deposits are dominated by poorly sorted, structureless to diffusely-layered tuff-breccias and coarse lapilli-tuffs, with subordinate tuffs and fine lapilli-tuffs • Interesting features include tuff ring remnants (Ross *et al.* 2008), debris avalanche deposits (Reubi *et al.* 2005), huge clastic dykes (Ross & White 2005*b*), and thick mafic pyroclastic flow deposits (Ross *et al.* 2005) • Among the coarser-grained rocks, some are filling diatreme-like vents or vent complexes (White & McClintock 2001; Ross & White 2006), whereas other form layers filling pre-existing topographic depressions (Ross & White 2005*a*; McClintock & White 2006)
Karoo	• Mafic volcaniclastic deposits exposed locally within an area about 530 by 240 km in Lesotho and South Africa, underneath the main flood lavas (Du Toit 1954) • Includes thickly bedded to structureless, mainly coarse-grained mafic volcaniclastic deposits, 100–250+ m thick, within steep-walled depressions (5–40+ km^2) in pre-existing country rock; these centres are surrounded by sheets of thinner-bedded, mainly lapilli and ash-grade deposits, 10–100 m thick (e.g. McClintock *et al.* 2008)
Siberian Traps	• Mafic volcaniclastic deposits are thought to represent about a quarter of the total volume of the province on the Siberian platform (Viswanathan & Chandrasekharam 1981) • The thickest volcaniclastic accumulations (up to 700 m) are older than the lavas • In the Noril'sk area, some mafic volcaniclastic layers (including agglomerates) are up to 100 m thick, and a 15–25 m thick layer can be traced over 30 000 km^2 (Czamanske *et al.* 1998)

(*Continued*)

Table 1. *Continued*

Province	Features
Emeishan	• Mafic volcaniclastic deposits and sedimentary rocks containing mafic lava fragments are relatively widespread but their volume probably represents <10% of the province • Mafic volcaniclastic deposits up to 170 m thick with a potential distribution of 400 by 30–70 km • Occur in the lower parts of the stratigraphy and dominated by thick bedded, limestone and basalt block-bearing tuff breccias interbedded with lavas and accretionary lapilli tuffs

(e.g. East Greenland; Emeishan; Ferrar; Karoo; Tunguska basin on Siberian platform); in the latter cases, clastic accumulations can reach hundreds of metres in thickness (Fig. 9). In the silicic LIPs, mafic volcaniclastic rocks generally occur in the upper parts of the eruptive stratigraphy (Fig. 12) following short-lived, large volume pulses of silicic ignimbrite volcanism (e.g. Whitsunday, Sierra Madre Occidental).

2. The known areal extent of mafic volcaniclastic deposits ranges from quite restricted (e.g. the vent-proximal pyroclastic accumulations on the Columbia River plateau) to hundreds of thousands of square kilometres (Siberian platform, Fig. 10).

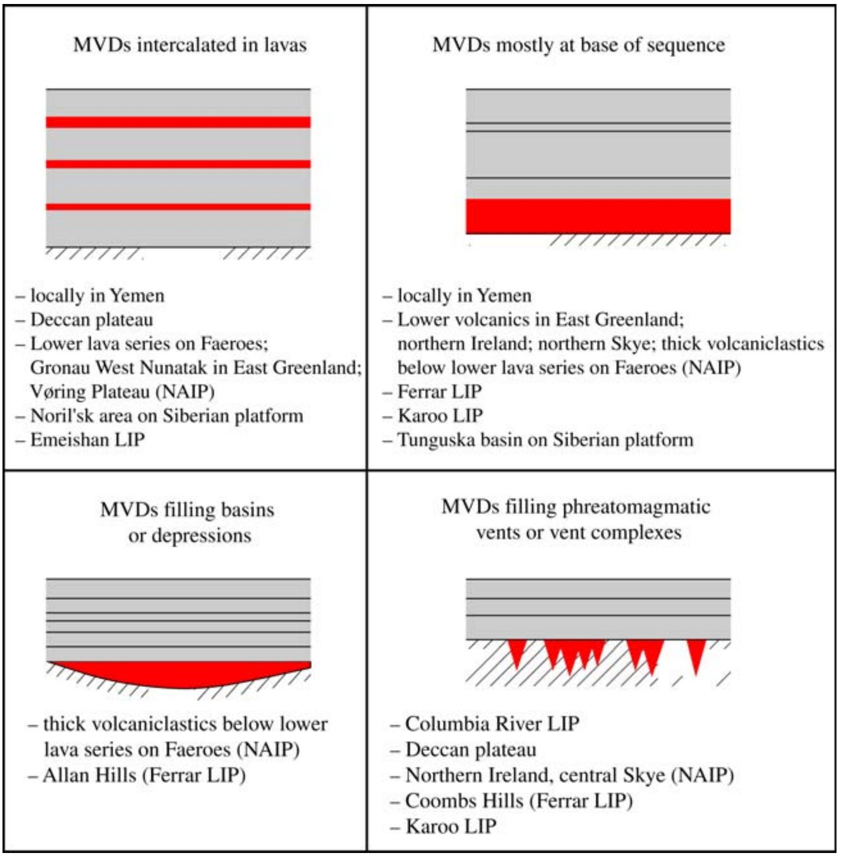

Fig. 9. Schematic illustration summarizing the different positions that mafic volcaniclastic deposits (MVDs) can occupy in flood basalt provinces. NAIP = North Atlantic Igneous Province. Modified from Ross *et al.* (2005).

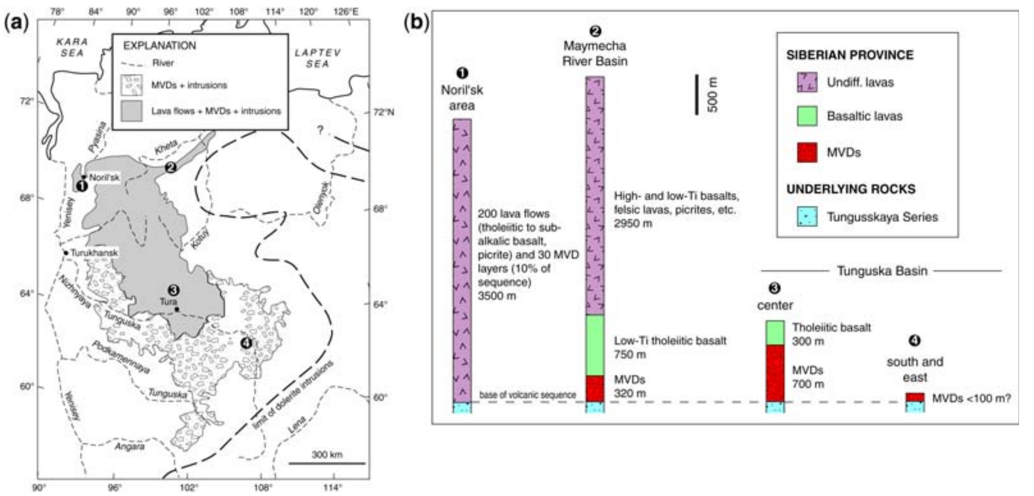

Fig. 10. Simplified map (**a**) and summary stratigraphic sections (**b**) of the Siberian Traps (modified from Ross *et al.* 2005). On the Siberian Platform, the area covered by mafic volcaniclastic deposits (MVDs) only, without flood lavas, is equivalent to that covered by flood lavas plus MVDs. Circled numbers indicate the general location of stratigraphic sections. River names in italics.

3. In many provinces, the deposits include abundant coarse lapilli-tuffs and tuff-breccias of mostly phreatomagmatic origin (e.g. East Greenland, Ferrar, Karoo, Emeishan), and are presumably exposed close to the source vents (Fig. 11).
4. Particles forming the volcaniclastic deposits were distributed by a range of processes including pyroclastic density currents (Ross & White 2005*a*), vent- or conduit-confined debris jets (White & McClintock 2001; Ross & White 2006; McClintock & White 2006), fall from volcanic plumes, mass flows (e.g. Hanson & Elliot 1996) and stream flows (e.g. McClintock *et al.* 2008).
5. Mafic volcanism in the silicic LIPs has produced the same volcanic structures as modern basaltic intraplate (largely monogenetic) volcanic fields such as the Newer Volcanics Province (Johnson 1989) and Auckland Volcanic Field (Johnson 1989; Houghton *et al.* 1999).

Mafic volcaniclastic deposits produced as a result of magmatic fragmentation are generally of limited volume in LIPs, and those of the Columbia River Basalts described previously are typical. It is inferred that fire fountaining was relatively common at vent sites for mafic flood lavas (Thordarson & Self 1996), but in many cases, the very hot pyroclasts (lava clots, spatter) coalesced upon landing to form clastogenic lava feeding into extensive flow fields. The original volume of basalt that underwent magmatic fragmentation during Columbia River (and other mafic LIP) eruptions was probably greater than that preserved in the geologic record, because coalescence processes convert fountain pyroclasts to lava, and because pyroclastic deposits are locally eroded by lava-flow 'bulldozing' (as seen during modern eruptions in Hawaii; Swanson *et al.* 1975). Rare, weakly welded pyroclastic deposits composed of degassed basalt fragments with few lithics have been described from the Karoo (McClintock *et al.* 2008), but the style of magmatic eruption that produced them remains incompletely understood.

Much larger volumes of preserved mafic volcaniclastic deposits are attributed to magma–water interaction, i.e. quench-induced hyaloclastic and explosive phreatomagmatic fragmentation (Ross *et al.* 2005). In the East Greenland, Ferrar, Karoo and Emeishan provinces, coarse phreatomagmatic lapilli-tuffs and tuff-breccias are composed of dense to poorly vesicular blocky sideromelane (or altered basaltic glass) clasts mixed with country rock fragments, or in some cases, loose quartz particles derived from sandstones (Hanson & Elliot 1996; Elliot & Hanson 2001; Ukstins Peate *et al.* 2003*a*; McClintock & White 2006). Limestone-bearing mafic volcaniclastic rocks toward the base of the Emeishan province reflect explosive interaction between flood basalt magmas and a carbonate reef platform. Basaltic accretionary lapilli tuffs, found in several provinces (North Atlantic, Ontong Java Plateau, Ferrar, Karoo, Emeishan), are commonly associated with phreatomagmatic eruptions and indicate the generation of subaerial plumes. Mafic phreatomagmatic activity in some silicic LIPs where lakes were present locally produced

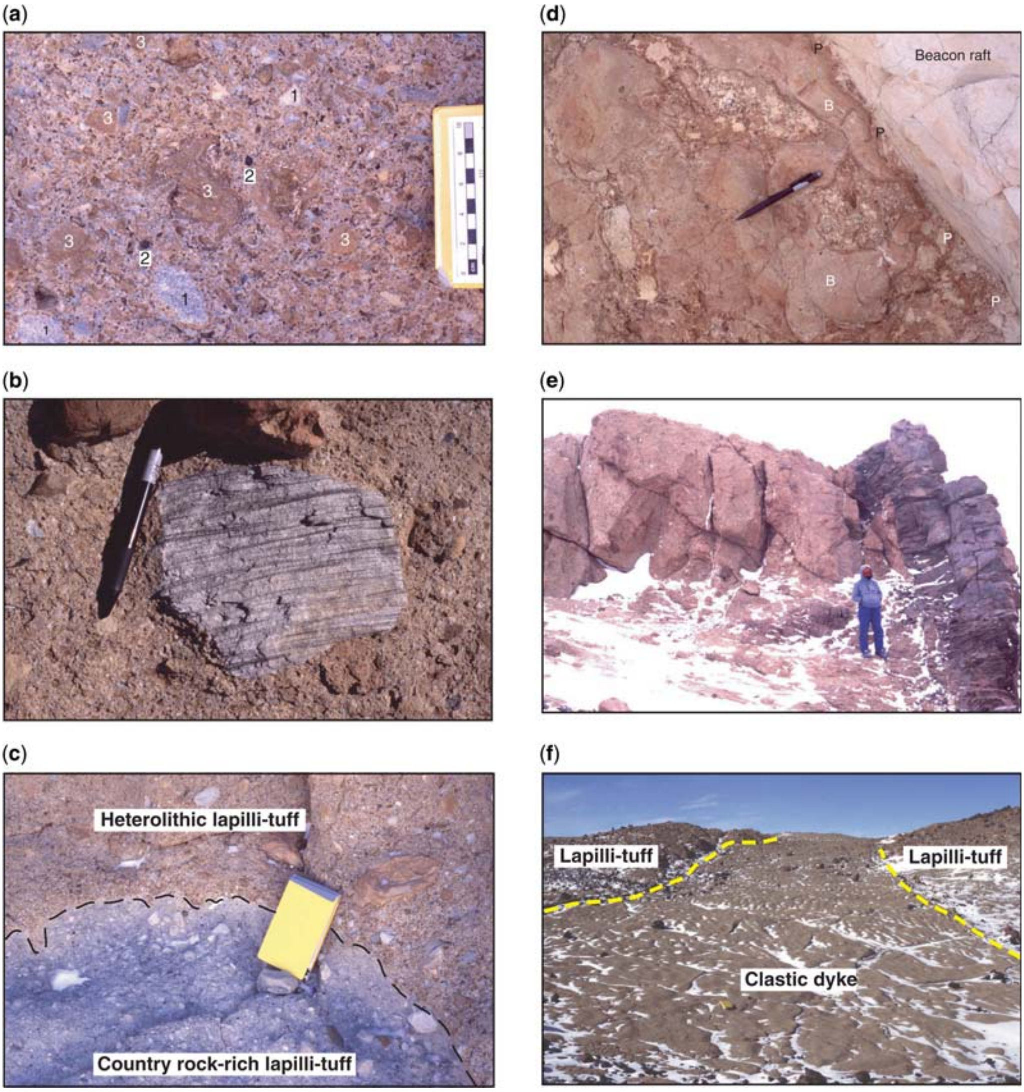

Fig. 11. Illustration of mafic volcaniclastic deposits and associated features in a phreatomagmatic vent complex at Coombs Hills, Ferrar LIP (after Ross 2005). (**a**) The dominant volcaniclastic rock filling the vent complex is a heterolithic lapilli-tuff containing (1) Beacon sandstone clasts, (2) coal clasts (in black, above the numbers) and (3) basaltic fragments of various shapes. Unlabelled Beacon (country rock) fragments are pale to white, unlabelled basaltic fragments are in medium and dark shades. Scale bar in centimetres. (**b**) A fragment of medium to coarse sandstone with dark laminae in the heterolithic vent fill. (**c**) View from above of a pipe of country rock-rich lapilli-tuff cross-cutting the dominant vent fill. It has been inferred that such pipes were generated by vertically travelling 'debris jets' above explosion sites (Ross & White 2006). (**d**) Rafts of quartz-rich sandstone from the Beacon Supergroup, up to hundred of metres in length, 'float' inside non-bedded volcaniclastic rock in the Coombs Hills phreatocauldron. The sandstone in the raft was wet and poorly consolidated during volcanism, so the rafts interacted with the basaltic magma (B), creating basalt-sandstone peperite (P) near the raft margins. (**e**) Sub-vertical, >2 m wide basaltic dyke (right of person) invading tuff-breccias. Note columnar joints perpendicular to the dyke wall. Such dykes, and/or cylindrical plugs of basalt cross-cutting the volcaniclastic rocks, could have served as feeders for the flood lavas higher up in the sequence. (**f**) Dykes filled by tuff (clastic dykes), including this 12 m wide example, cross-cut the vent filling deposits (see Ross & White 2005*b* for details).

vent-proximal accumulations (tens to hundreds of metres thickness) of basaltic base surge deposits and/or accretionary lapilli-bearing tuff rings or cones. Mafic phreatomagmatism in silicic LIPs appears to be far less significant than in the CFBPs, no doubt attributable in part to the much smaller volumes of mafic magma erupted.

Where abundant surface water is present on continents or when lava enters the sea, mafic lava flows generally experience quenching and spalling; typical products include pillow lavas, pillow–palagonite breccias and hyaloclastite. Pillow–palagonite complexes are common at the base of basalt flows in parts of the Columbia River Plateau (Swanson & Wright 1981; Hooper 1997), and are present in rift basins (e.g. Hartford Basin) of the Central Atlantic Magmatic Province. Pillow lavas are also present in the Karoo succession (McClintock *et al.* 2008), in lower and central sections of the Emeishan flood basalt province (Binchuan section of Xu *et al.* 2004), and at the base of the Mull lava succession in the North Atlantic Igneous Province (Kerr 1995). Hyaloclastite piles are documented onshore in East Greenland (Nielsen *et al.* 1981) and mafic volcaniclastic deposits may be important in syn- to post-rift sequences along volcanic rifted margins (Planke *et al.* 2000). A hyaloclastite complex is also documented from an Archaean LIP in the eastern Pilbara craton of Western Australia (Blake 2001). Pillowed lavas and peperitic intrusions have also been reported, formed in marine or, more commonly, lacustrine environments (Pankhurst *et al.* 1998; Bryan *et al.* 2000).

Autoclastic fragmentation of mafic lavas generates breccias (e.g. aa clinker, slabby and rubbly pahoehoe). Interaction of lavas with aeolian sand formed peperitic rocks in the arid landscape in which the Etendeka LIP was emplaced (Jerram & Stollhofen 2002). Other peperites associated with mafic LIPs clearly formed in the presence of water, as where the Pomona Basalt invaded water-saturated silicic ash of the Ellensburg Formation to form peperites exposed over an area of about 5400 km^2 in south-central Washington (Schmincke 1967). Peperites are also known from inside the vents of mostly phreatomagmatic mafic volcaniclastic deposits, where wet fragmental material was in abundant supply (McClintock & White 2002; Ross 2005).

Reworked mafic volcaniclastic deposits are known from East Greenland (Heister *et al.* 2001; Ukstins Peate *et al.* 2003*a*), the North Sea (phase 1 ashes, Knox & Morton 1988; Morton & Knox 1990), the Ferrar (Carapace Sandstone, see Ross *et al.* 2008) and the Karoo (Ross *et al.* 2005; McClintock *et al.* 2008). Reworked deposits are a natural accompaniment, in variable proportions, to primary mafic volcaniclastic deposits wherever they are found.

Silicic volcaniclastic rocks and lavas in LIPs

Silicic volcaniclastic rocks have long been recognized within LIPs otherwise dominated by flood basaltic lavas and ranging in age from Precambrian (e.g. Twist & French 1983; Thorne & Trendall 2001; Blake *et al.* 2004) to Cenozoic. Detailed descriptions are relatively limited in the literature (see review in Bryan *et al.* 2002), and our knowledge of the scale and magnitude of silicic eruptions from CFBPs has been greatly improved by the detailed studies on silicic volcanic units in the Paraná-Etendeka (Milner 1988; Milner *et al.* 1992, 1995; Bellieni *et al.* 1986; Garland 1994; Garland *et al.* 1995; Ewart *et al.* 1998, 2004), and most recently, the Afro-Arabian flood volcanic provinces (Baker *et al.* 1996; Ukstins Peate *et al.* 2003*b*, 2005).

Silicic LIPs represent the largest accumulations of volcaniclastic rocks on Earth and have dimensions comparable to those of CFBPs (Bryan *et al.* 2002; Bryan 2007; Sheth 2007; Bryan & Ernst 2008). Like their better-known CFBP counterparts, silicic LIPs bear all the hallmarks of LIP events, in particular, large erupted volumes ($>$0.25 Mkm^3) and areal extents ($>$0.1 Mkm^2), evidence for very high magma-emplacement rates over short periods and intraplate tectonic settings or geochemical affinities (Bryan 2007; Bryan & Ernst 2008). Ignimbrite-dominated silicic LIPs of Late Palaeozoic to Cenozoic age are the best preserved, and represented by the Early Cretaceous Whitsunday Igneous Province of eastern Australia (Ewart *et al.* 1992; Bryan *et al.* 1997, 2000), the Jurassic Chon Aike Province of South America-Antarctica (e.g. Pankhurst *et al.* 1998, 2000; Riley & Leat 1999); the middle Cenozoic Sierra Madre Occidental Province of Mexico (e.g. McDowell & Clabaugh 1979; Swanson & McDowell 1984; Ferrari *et al.* 2002), and the Permo-Carboniferous Kennedy–Connors–Auburn Igneous Province of eastern Australia (Bain & Draper 1997; Bryan *et al.* 2003). The Whitsunday and Chon Aike provinces were spatially and temporally related to emplacement of other LIPs and episodes of continental break-up. Proterozoic examples have significantly smaller preserved volumes (1×10^5 km^3), and may include the *c.* 750 Ma old Malani (India; Sharma 2005) and the *c.* 1590 Ma old Gawler Range-Hiltaba (South Australia; Daly *et al.* 1998; Allen *et al.* 2003) igneous provinces.

The Sierra Madre Occidental province of Mexico, being the best preserved example of a silicic LIP, is representative of their general architecture. It is an extensive, relatively flat-lying ignimbrite plateau covering from 10^5 to 10^6 km^2 (2000 km long and 200–500 km wide) to an average thickness of 1 km (e.g. McDowell & Clabaugh 1979;

Ferrari *et al.* 2002). Sierra Madre Occidental volcanism was concurrent with the widespread 'ignimbrite flare-up' in the Basin & Range Province of western USA, where an additional 10^5 km^3 of dacitic to rhyolitic ignimbrite was emplaced between *c*. 35 and 20 Ma, but which has now largely been dismembered by basin and range extension (Lipman *et al.* 1972; Gans *et al.* 1989; Best & Christiansen 1991; Johnson 1991; Axen *et al.* 1993; Ward 1995). In general, the combination of burial, tilting, faulting and exhumation have hindered mapping and volume estimation of individual ignimbrite eruptive units in LIPs. As an example, $<10\%$ of the youngest and least deformed Sierra Madre Occidental province has so far been mapped (Swanson & McDowell 2000; Swanson *et al.* 2006); an important item on the 'future research' agenda is to improve the database on large-volume silicic provinces (see also Mason *et al.* 2004).

Even compared with flood-basaltic lava flows, silicic volcaniclastic deposits may be emplaced over vast areas (*c*. 2% of the globe for Toba; Rose & Chesner 1987), and may also have eruptive volumes (e.g. Mason *et al.* 2004) that equal or exceed those of the largest mafic lava flows. The large eruptive fluxes and upper-atmospheric effects of silicic eruptions may combine with effects of flood basalt eruptions to force environmental and climatic changes during LIP emplacement, so it is important to understand their timing relative to, and petrogenetic relationships with, roughly coeval flood basalts. In the CFBPs, the silicic pyroclastic deposits are also of great stratigraphic utility, forming laterally continuous and distinctive marker horizons that constrain volcanic stratigraphies over several hundreds of kilometres within often monotonous and internally complex flood basalt lava successions (Jerram 2002; Bryan *et al.* 2002; Ukstins Peate *et al.* 2005). For example, large-volume silicic units have proved vitally important in establishing stratigraphic correlations between the Paraná and Etendeka flood basalt provinces, which are now separated by the South Atlantic Ocean (Milner *et al.* 1995). Silicic volcanic and volcaniclastic rocks can represent a significant contribution to the total magmatic output of CFBPs (5–10 vol%; e.g. Paraná-Etendeka; Afro-Arabian). The size of some individual Paraná-Etendeka silicic units is truly impressive, with the largest covering areas $>100\,000$ km^2 and representing erupted volumes of *c*. 5000 km^3 dense rock equivalent; they are amongst the largest volume eruptive units so far recognized on Earth (Milner *et al.* 1995; Ewart *et al.* 1998, 2004; Marsh *et al.* 2001; Mason *et al.* 2004).

Pyroclastic rocks, primarily ignimbrites (pyroclastic-density-current deposits) are the dominant products of silicic volcanism within CFBPs, silicic LIPs and along volcanic passive margins. This is particularly the case for those CFBPs having large volumes of silicic deposits ($>10^3$ km^3; Bryan *et al.* 2002). Ignimbrite typically represents $>75\%$ of the total stratigraphic thickness (generally >1 km; Fig. 12) of silicic LIPs, and multiple units can cover areas in excess of 5×10^5 km^2. Thinner intervals of silicic tuffs or bentonites and breccias have been reported from oceanic plateaux (e.g. Northwest Australian margin, von Rad & Thurow 1992; Caribbean–Colombian oceanic plateau; Kerr *et al.* 2004), and submerged syn- to post-rift sequences along continental margins (e.g. Larsen *et al.* 2003). Most recently, deep-sea ash layers thousands of kilometres from source have been correlated to onshore deposits of silicic explosive volcanism in the Afro-Arabian flood basalt province (Ukstins Peate *et al.* 2003*b*; Touchard *et al.* 2003*a*, *b*), with one major consequence being to dramatically increase the known volumes of silicic magma erupted in this province (Ukstins Peate *et al.* 2005, 2007). This result suggests that the volumetric significance of silicic volcanism in mafic LIPs, inferred for most provinces only from onshore records, is likely to have been greatly underestimated.

Ignimbrites in LIPs are dominantly dacitic to rhyolitic welded units that range from tens to a few hundred metres in thickness, and possess tabular geometries with moderate to low aspect ratios reflecting lateral extents of up to hundreds of kilometres, and dense rock equivalent volumes $>10^3$ km^3 (e.g. McDowell & Clabaugh 1979; Milner *et al.* 1995; Pankhurst *et al.* 1998; Bryan *et al.* 2000; Ukstins Peate *et al.* 2005). Intracaldera facies ignimbrite units >1 km thick have been reported from the silicic LIPs, and are commonly associated with coarse lithic lag breccias (e.g. Ewart *et al.* 1992; Pankhurst *et al.* 1998; Bryan *et al.* 2000; Swanson *et al.* 2006). In detail, ignimbrites within and from different silicic LIPs show considerable variety in deposit features, such as welding intensity and crystal, lithic and pumice contents. Rheomorphic ignimbrites are rare in silicic LIPs, reflecting generally low eruptive temperatures (750–900 °C; Cameron *et al.* 1980; Wark 1991; Ewart *et al.* 1992). In contrast, many silicic eruptive units in the CFBPs are lava-like in character, particularly those in the Paraná Etendeka and Karoo provinces (e.g. Cleverly 1979; Milner *et al.* 1992). The preferred interpretation has been that these units formed as rheoignimbrites or extremely high-grade ignimbrites (Branney & Kokelaar 1992; see also Walker 1983) that erupted with relatively low explosivity at high eruptive temperatures (>1000 °C) and low magma viscosities (10^4 Pa s). Pyroclasts apparently maintained high temperatures to the sites of deposition, resulting

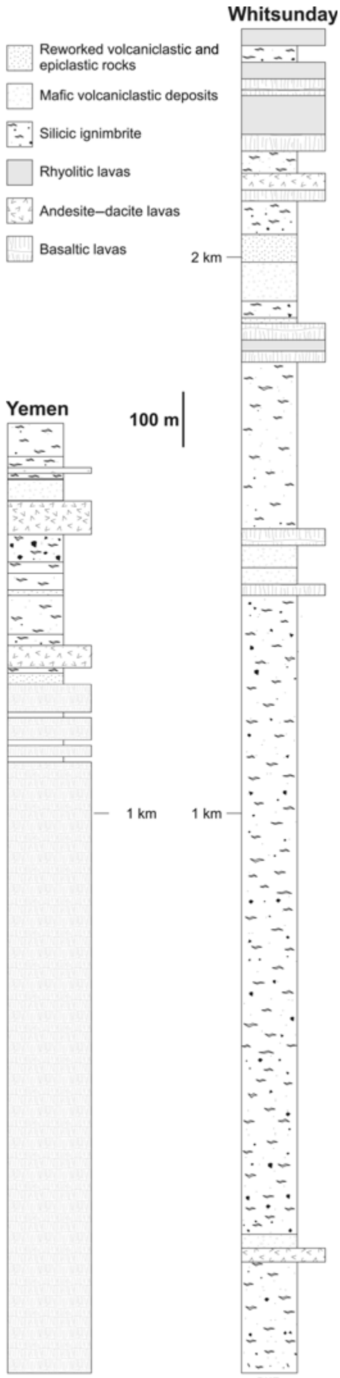

Fig. 12. Generalised composite stratigraphic sections comparing the Afro-Arabian continental flood basalt provinces with the Whitsunday Silicic LIP. Lavas and volcaniclastic rocks are shown at arbitrary thicknesses on the *x*-axis to emphasize distinction in rock types; dykes, sills and other intrusions are not shown. The Yemen (*Continued*) section of the Tertiary Afro-Arabian continental flood basalt province, based on Baker *et al.* (1996) and Ukstins Peate *et al.* (2005); the section for the Early Cretaceous Whistunday Silicic LIP, based on Bryan *et al.* (2000), with the base to section not exposed (BNE). Note the similar trend in both LIP sections towards bimodal volcanism characterizing the upper parts of the preserved stratigraphies (see also Bryan 2007).

in agglutination/coalescence and rheomorphism (Milner *et al.* 1992; Bryan *et al.* 2002). New studies of the quartz latites in the Etendeka province (Mawby *et al.* 2006) have recognized that the lava-like units exhibit the classic valley-fill geometry of ignimbrites, and in contrast to long silicic lava flows, do not possess basal or upper autobreccias along their strike length. Many other examples of ignimbrites in mafic LIPs have unequivocal pyroclastic textures, inferred eruptive temperatures <1000 °C (e.g. Kirstein *et al.* 2000; Ukstins Peate *et al.* 2005), and are potentially the products of eruptions that generated higher atmospheric plumes. This is confirmed by the correlation of such ignimbrites with distal ash fall deposits located thousands of kilometres from source (Ukstins Peate *et al.* 2003*b*; Touchard *et al.* 2003*a*, *b*).

Silicic phreatomagmatic deposits are relatively common in silicic LIPs, occurring as locally thick sequences of base surge deposits and/or accretionary lapilli tuff (e.g. Bryan *et al.* 2000), with several examples formed during early phases of ignimbrite-forming eruptions. Some accretionary lapilli have been interpreted to occur in deposits of co-ignimbrite ash clouds (Riley & Leat 1999). In general, the phreatomagmatic deposits are fine-grained and lithic-poor, reflecting a highly efficient fragmentation process inferred to result from explosive interaction of magma with surface water. Sites of explosive magma–water interaction were lakes within rift valleys or flooded calderas, with the latter analogous to the eruptive environments that have characterized silicic explosive eruptions in the Taupo Volcanic Zone (e.g. Wilson *et al.* 1995). Quench fragmented silicic lava is common where lavas were extruded onto the floors of flooded calderas.

Rhyolitic lavas and associated autoclastic facies are present in continental LIPs, but are clearly subordinate in volume to silicic pyroclastic rocks. Lavas appear to be more volumetrically significant, however, in Precambrian provinces (e.g. Bushveld, Twist & French 1983; Mesoproterozoic Gawler Range Volcanics, Allen & McPhie 2002; Allen *et al.* 2003), and where the total volume of silicic volcanics in a LIP is low ($<10^3$ km^3; e.g. Deccan, Lightfoot *et al.* 1987).

As recognized following recent caldera-forming (26.5 ka ago, Oruanui; Taupo 1800*a*; Manville &

Wilson 2004; Manville *et al.* 2005) and small-volume historical eruptions (e.g. Mount St Helens, Major *et al.* 2000; Pinatubo, Newhall & Punongbayan 1996), the remobilization and redeposition of unconsolidated pyroclastic material by fluvial and mass-flow processes can be extensive and protracted, affecting large areas otherwise untouched by the eruption. Similarly, substantial remobilization and reworking of silicic volcaniclastic material has occurred during eruptive hiatuses in the emplacement of LIPs, resulting in reworked volcaniclastic deposits. Resedimented volcaniclastic deposits are volumetrically larger in silicic LIPs than in CFBPs, reflecting the greater abundance of pyroclastic debris (non-welded ignimbrites, fall and surge deposits) available for reworking. Substantial volumes of reworked volcaniclastic rocks are associated with the Whitsunday silicic LIP in eastern Australia (Bryan *et al.* 1997), which was the source of $>1.4 \times 10^6$ km^3 of predominantly sand-sized volcaniclastic sediment that was rapidly generated and transported over large distances with limited weathering (Smart & Senior 1980), fundamentally altering the basin-fill history of at least two major continental sedimentary basin systems (Bryan *et al.* 1997).

Implications of volcaniclastic rocks in LIPs

The types and volumes of volcaniclastic rocks present within LIPs are strongly related to: (1) LIP composition (i.e. mafic v. silicic); (2) primary volcanic fragmentation mechanisms; (3) environmental conditions of eruptions, particularly for mafic LIPs; and (4) environmental conditions of emplacement (e.g. continental flood basalt provinces v. oceanic plateaux).

Primary fragmentation processes

Mafic and silicic LIPs show a similar range of volcanic fragmentation and eruptive processes, but marked differences in the dominant eruptive styles (Table 2). In mafic LIPs, basaltic effusive eruptions dominate, whereas in silicic LIPs, silicic explosive eruptions dominate. Volcaniclastic deposits are more volumetrically significant for those mafic LIPs that contain: (1) substantial volumes (10^3–10^4 km^3) of silicic magma erupted as ignimbrite or other pyroclastic deposits (e.g. Paraná-Etendeka, Karoo, Afro-Arabia; Bryan *et al.* 2002); and/or where (2) external water sources were involved in

Table 2. *A comparison and rating of the variety of primary fragmentation processes, and the importance of reworking to produce volcaniclastic rocks in LIPs*

Process	Mafic LIPs	Silicic LIPs
'Magmatic' fragmentation	Minor process for basaltic magmas producing mostly near-vent deposits (e.g. spatter ramparts/cones and scoria cones/fall deposits). Dominant process for silicic magmas	Dominant primary process producing widespread ignimbrites, Plinian fallout and co-ignimbrite tuffs
Phreatomagmatic fragmentation	Significant for continental flood basalts emplaced through pre-existing hydrologic reservoirs or where developing rifts are flooded (e.g. NAIP[a]); mafic magma interaction with groundwater in sedimentary aquifers; minor to rare for silicic magmas	Significant, especially in flooded caldera and rift valley environments; magma interaction mainly with surface water (i.e. lakes)
Autoclastic fragmentation	Dominant primary process fragmenting basaltic lava flows; minor for silicic rheoignimbrites	Minor to rare, affecting lavas of all composition (basalt to rhyolite)
Hyaloclastite formation	Minor process in subaerial LIPs; likely to be more significant for subaqueously emplaced LIPs (e.g. oceanic plateaux, submarine ridges, seamounts)	Minor process generally affecting lavas emplaced into hydrologic reservoirs (e.g. lakes, rivers)
Reworking and resedimentation	Minor process in provinces consisting of largely coherent mass of unfragmented lava; becomes more significant where mafic phreatomagmatic and silicic pyroclastic deposits (nonwelded) present	Important; substantial remobilisation of the primary silicic volcaniclastic deposits

[a]NAIP = North Atlantic Igneous Province.

mafic eruptions, leading to large-scale phreatomagmatic basaltic volcanism and resultant mafic volcaniclastic deposits with volumes of up to 10^2-10^5 km^3 (e.g. Siberian Traps, Ferrar, North Atlantic Igneous Province, Emeishan; Ross *et al.* 2005). These contrasting fragmentation mechanisms for the silicic and mafic magmas result in dramatically different eruption and emplacement processes. Knowing the relative contributions of effusive v. pyroclastic deposit is critical for assessing the environmental impact of individual large-igneous provinces (Table 3).

The largest volumes (10^5-10^6 km^3) and proportions of volcaniclastic deposits are associated with silicic LIPs, for which volcaniclastic deposits alone can exceed the total eruptive volume and areal extent of some CFBPs (Bryan *et al.* 1997). Pyroclastic eruptions are critical in producing the large volumes of primary and reworked volcaniclastic rocks (Bryan *et al.* 1997, 2000). In contrast, in the CFBPs, the mafic effusive eruptions produce only minor volumes of coarse fragmental material (e.g. aa lava breccias, rubbly pahoehoe, scoria deposits). The common burial of fragmental deposits by the lava flows that formed them, combined with the limited fragmentation accompanying emplacement of pahoehoe to rubbly pahoehoe lavas that typify these provinces, are both important factors in limiting the amount of clastic material generated and available for reworking, and help explain why CFBPs produce only limited sedimentary signals in the geologic record.

Table 3. *Comparison between caldera-forming (silicic) ignimbrite eruptions and explosive mafic eruptions in LIPs*

Feature	Caldera-forming (silicic) ignimbrite eruptions in LIPs	Explosive mafic eruptions in LIPs
Dominant fragmentation mechanism	'Magmatic' explosive fragmentation of viscous and vesiculated magma	Phreatomagmatic, driven by explosive interaction of magma with external water
Pre-welding/compaction vesicularity of juvenile clasts	Mostly high, moderate to low for rheoignimbrites of CFBPs	Variable but mostly low (Ross *et al.* 2005)
Dominant transport mechanism	Pyroclastic density currents	Varied (see text)
Eruption rates	Very high	Varied
Recurrence	Low (10^3-10^5 years) but probably higher during main eruptive phase(s)	Relatively high?
Eruption duration	Cataclysmic eruptions, hours/days to weeks[a]	Several eruptions over many years? (Ross 2005)
Eruption plumes	Sub-Plinian to ultra-Plinian	Small to ultra-Plinian?
Sulfur concentration in magmas	Unknown, dissolved concentrations in magmas less than basalt but 'excess sulfur' budgets potentially substantial, especially via basaltic underplating	High (e.g. Thordarson & Self 1996; Thordarson *et al.* 1996)
Potential for environmental impacts	High: largest known eruptions have areal extents $>10^5$ km^2; stratospheric injections of ash and aerosols would produce at least short-term climatic impacts; highest impact for eruptions at low latitudes[b]	Good if there is a cumulative effect from several consecutive eruptions (e.g. Jolley & Widdowson 2005)
Volcanic depressions	Large calderas and nested caldera complexes formed by vertical collapse following rapid evacuation of magma chambers; volcanotectonic depressions with subsidence $\pm$ eruption foci controlled by regional tectonic structures	'Phreatocauldrons' (White & McClintock 2001) formed progressively through vent migration and coalescence, plus lateral quarrying; no boundary faults or massive subsidence[c]

Note: As an analogy, the Toba 74 ka 'supereruption' mean eruption rate was *c*. 7×10^9 kg s^{-1} (Oppenheimer 2002).
[a]Large-volume silicic pyroclastic units deposited in hours, e.g. the Bishop Tuff (Wilson & Hildreth 1997).
[b]Long duration of silicic LIP volcanism will also contribute to environmental effects (see also Bryan 2007).
[c]Phreatocauldrons do not fit the current definition of caldera: 'A volcanic structure, generally large, which is principally the result of collapse or subsidence into the top of a magma chamber during or immediately following eruptive activity' (Cole *et al.* 2005).

Vent types

The vent sites for some primary mafic volcaniclastic deposits in mafic LIPs have been identified (e.g. East and West Greenland: Pedersen *et al.* 1997; Ukstins Peate *et al.* 2003*a*; Larsen *et al.* 2003; Ferrar: Ross & White 2006; Ross *et al.* 2008; Karoo: McClintock *et al*. 2008). In the Ferrar and Karoo provinces, exposed composite vent sites are called 'phreatocauldrons' (White & McClintock 2001) because of the inferred phreatomagmatic fragmentation mechanism and the overall cauldron shape of the vent complexes, which can be described as 'nests of cross-cutting diatremes'. The mafic vent complexes described so far are apparently less than 10 km wide.

Known eruptive sources for silicic pyroclastic rocks in the CFBPs and along volcanic rifted margins are caldera-type complexes (e.g. Bell & Emeleus 1988; Ewart *et al.* 2002; Bryan *et al.* 2002). However, locations of silicic eruptive centres are poorly constrained in LIPs even for those provinces with well-mapped, abundant, extensive and voluminous silicic eruptive products (e.g. Paraná-Etendeka, Marsh *et al.* 2001; Afro-Arabian, Ukstins Peate *et al.* 2005). A key target for future work in LIPs is to locate and identify eruptive sites for silicic rocks, in order to constrain the volcanostratigraphy, eruption volumes and emplacement mechanisms.

Caldera complexes are typically parts of multiple-vent volcanic systems, which include numerous extracaldera (and intracaldera) monogenetic edifices. The mafic volcanic deposits consist of scoria/spatter cones and tuff rings/cones or maars, whereas silicic effusive eruptions generate lava domes or nested dome complexes (e.g. Bryan *et al.* 2000). Evidence for calderas as source vents for large-volume ignimbrites includes coarse lithic lag breccias and megabreccias within the ignimbrite eruptive units, and very thick, localised deposits inferred to be caldera-ponded (e.g. Swanson & McDowell 1984; Ewart *et al.* 1992; Bryan *et al.* 2000; Swanson *et al.* 2006). Calderas in most silicic LIPs appear to range between 10 and 30 km in diameter, but larger calderas from *c*. 40 km up to 100 km in diameter have been inferred for the Chon Aike Province (Aragón *et al.* 1996; Riley *et al.* 2001). Remnants of many caldera complexes (cauldrons) are well-exposed in the deeply dissected Permo-Carboniferous Kennedy subprovince of northeastern Australia, where they exhibit a variety of geometries and composite arrangements, with individual calderas ranging from 10 to 40 km in diameter (Bain & Draper 1997).

Locating calderas is also difficult in silicic LIPs because vent sites are often buried by caldera-fill deposits emplaced in the course of eruptions during caldera subsidence. Vent sites may also be buried by products, either primary or redeposited, of eruptions elsewhere in a province, given the volumes of volcanic material generated ($>10^5$ km^3) over relatively short periods of time. Later tectonism and erosion may also obscure vent sites (Bryan *et al.* 2002). Fewer than 15 calderas (Valles-type, Williams & McBirney 1979) have, for example, been identified to date in the Sierra Madre Occidental (Aguirre-Díaz & Labarthe-Hernández 2003), but given the preserved areal distribution of 393 000 km^2, the existence of as many as 400 calderas has been postulated (McDowell & Clabaugh 1979). New work in a small sector of the northern Sierra Madre Occidental indicates a relatively high density of locally overlapping calderas similar to those in the coeval San Juan volcanic field in Colorado (Swanson *et al.* 2006). The problem of caldera recognition is even more extreme for the Whitsunday silicic LIP, where only five eruptive centres have been identified in a province with an extrusive output of $>2.2 \times 10^6$ km^3 (Ewart *et al.* 1992; Bryan *et al.* 2000; Bryan 2007).

Recent work also suggests that some rhyolitic ignimbrites were emplaced from extensive volcanotectonic fissures linked to regional extension. Welded pyroclastic dykes have been reported from the Whitsunday (Bryan *et al.* 2000), Sierra Madre Occidental (Aguirre-Díaz & Labarthe-Hernández 2003) and Chon Aike (Pankhurst *et al.* 1998) silicic LIPs. In the Sierra Madre Occidental, welded pyroclastic dykes or fissures are up to several kilometres in length and associated with regional Basin and Range extensional graben-bounding faults (Aguirre-Díaz & Labarthe-Hernández 2003). Although a general spatial overlap between silicic volcanism and extensional faulting is known, the timing of volcanic pulses and major tectonic episodes is not well constrained for these provinces.

Volcanic environments

Eruptive and depositional environments for many of the better-studied LIPs, both mafic and silicic, were primarily subaerial. On continents, wet environments fostering phreatomagmatic fragmentation and hyaloclastite formation are dominated by intracaldera lakes or volcanotectonic rift basins developed before or during LIP magmatism. In contrast, the Late Permian Emeishan flood basalt province appears to have been emplaced onto a shallow marine carbonate platform developed across the Yangtze craton in southwest China. Phreatomagmatic activity produced limestone-bearing mafic volcaniclastic rocks and accretionary lapilli tuffs, and submarine emplacement of basaltic pillow lava occurred during the early eruptive stages and in the core of the Emeishan flood basalt province. The lateral and temporal distributions of these rocks within the volcanic pile ('inner zone' of He *et al.* 2003) suggest that the upper limits of the accommodation space remained close to, or below, sea level during the early stages of volcanism, indicating that no kilometre-scale pre-volcanic lithospheric doming, such as predicted by the mantle plume hypothesis, took place (cf. He *et al.* 2003; Xu *et al.* 2004; Campbell 2005).

Emplacement during continental LIP volcanism of kilometres-thick pyroclastic piles over short time spans may inhibit formation of sizeable bodies of surface water if the deposits are highly permeable or if the climate is arid. Ignimbrite surfaces commonly have relatively modest permeabilities, however, as illustrated by the formation of a large temporary lake atop the Taupo ignimbrite (Manville 2001), and blocking of streams by ignimbrites commonly leads to drainage disruption, development of new lakes, and deepening of existing ones (Manville *et al.* 1999; Manville 2002). Flood basalt lava flows may develop significant topography by differential inflation across a lava flow field (e.g. Larsen *et al.* 2006), but the extremely high permeability of young lava fields makes surface ponding of water unlikely unless the groundwater table rises above the flows. An example of this extreme permeability is provided by the Snake River Plain aquifer, which occupies a Cenozoic stack of basaltic lavas up to 3 km thick (Greeley 1982). It is 'one of the most permeable large aquifer systems in the world' (p. 7, Hackett *et al.* 1986). Some wells demonstrate transmissivities in excess of 50 000 m^2/day (Welhan & Reed 1997), with the bulk of water moving along tops and bases of pahoehoe flow units, although jointing gives even massive flow interiors moderate lateral permeability (Welhan & Reed 1997). The 'Big Lost River', discharging about *c.* 70 $m^3\ s^{-1}$ before flowing onto the lavas (Ostenaa *et al.* 2002), simply disappears into the surface of the aquifer, with discharge elsewhere feeding springs in excess of 7000 cfs (Hackett *et al.* 1986; *c.* 200 $m^3\ s^{-1}$). Despite the high permeability of young lava fields, however, emplacement of lavas may partially impound streams with sufficient discharge, and this may explain the abundance of small pillow–lava complexes interspersed within the sequences across the Columbia River Basalt Province (see Hooper 1997 and references therein). Where later eruptions encounter ground or surface water, explosive or passive magma–water interactions may take place, as in West Greenland, where lavas of the Maligat Formation erupted from subaerial vents and flowed into a deep and extensive paeleolake, generating a complex of hyaloclastites and rootless cones up to 200 m wide and 25 m high (Larsen *et al.* 2006). Interaction of mafic magmas with water also appears to be more prevalent when volcanism is associated with active lithospheric extension and subsidence. An example of this is the change from non-explosive, terrestrial flood basalt eruptions to highly explosive basaltic phreatomagmatic/phreato-Plinian eruptions towards the close of volcanic activity in the North Atlantic Igneous Province, related to flooding of the nascent North Atlantic Rift (Larsen *et al.* 2003; Jolley & Widdowson 2005).

Phreatomagmatic activity has also been common during opening stages of flood volcanism in sedimentary basins, which have a variety of water reservoirs (aquifers, lakes, rivers, shallow seas) that can supply the water for explosive eruptions (Ross *et al.* 2005). Non-explosive hyaloclastic fragmentation processes generating volcaniclastic deposits in association with sea floor lavas are presumably prevalent in initial deep-water eruptions of oceanic plateaux. Recent results from the Ocean Drilling Program (ODP) on the Ontong-Java Plateau (Leg 192, Site 1184; Shipboard Scientific Party 2001) show, however, that mafic volcaniclastic deposits of phreatomagmatic origin can form deposits of significant thickness in such plateaus, particularly where basaltic volcanism occurred at shallow water depths and/or was emergent (Thordarson 2004). Evidence from ODP Legs 119, 120 and 183 also clearly demonstrates that large parts of the Kerguelen oceanic plateau were originally subaerial during plateau construction (Frey *et al.* 2000).

In summary, the diversity of volcaniclastic material generated during LIP formation reflects the diversity of magma composition, eruption mechanisms and environmental conditions during eruptions, and in primary volcaniclastic emplacement processes as well as those effecting subsequent remobilization. The volcaniclastic deposits are unique in the richness of information they offer regarding environmental conditions during volcanic activity, and represent the most explosive of large igneous events. By better understanding emplacement of the sometimes undervalued volcaniclastic component of LIPs, we can address and constrain key issues of LIP petrogenesis and global impact.

Discussion and conclusions

With the accumulation of studies addressing the physical processes of eruptions in LIPs, it becomes ever more apparent that all the complexity observed in historical eruptions is present, but at a range of scales, and to a far lesser extent, rates, extending from the normal to the extraordinary. In a sense, this is unsurprising – LIPs are, after all, large, and the product of many large magnitude eruptions.

In other ways, however, the similarities with small-volume, modern eruptions may allow us to use LIPs to diagnose behavioural aspects of smaller eruptions. In particular, some provinces expose large tracts of shallow subvolcanic intrusions, which are presumed to represent the magma-feeding systems for numerous effusive eruptions over large areas. Although magma delivery processes may be significantly different for provinces that provide millions of cubic kilometres of magma for eruptions, these tracts can nevertheless offer important insights into the physical operation of other magmatic plumbing systems. With so much magma pumped into the crust, there are mappable geological effects that in smaller systems might require high-resolution mapping or geophysical studies to identify. In provinces such as the Ferrar, the results of this high magma throughput are well exposed at a level of detail that cannot be matched by subsurface geophysical investigations, and in a variety unmatched in any single smaller eroded volcanic system.

In conclusion, here is a listing of major findings from studies on LIPs, with an admitted emphasis on continental examples.

1. LIPs must have large plumbing systems that, at least for CFBPs, include an extensive network of crustal sills and dykes, the emplacement of which subdivides intervening country rock into differentially uplifted and tilted blocks.
2. Shallowing sills locally break through to the surface, disrupting and enveloping blocks of country rock.
3. Flood basaltic lava flows are emplaced as flow fields that advance and thicken by inflation for years or decades during sustained eruptions with fluxes in the range known from other more recent basaltic eruptions.
4. Fragmental deposits proximal to flood basalt vents provide evidence for periods of high fountaining and sub-Plinian plume dispersal, and based on smaller-scale modern analogues, have been associated with regional climate modification.
5. Many CFBPs and volcanic passive margins contain significant volumes of mafic volcaniclastic deposits, and where well studied, these have revealed evidence for extensive phreatomagmatic eruptions that preceded flood-basalt emplacement.
6. Eruptions producing mafic volcaniclastic deposits have produced caldera-scale vent complexes and substantial mafic pyroclastic-flow deposits, the former in some cases subsequently occupied by lakes.
7. Silicic pyroclastic eruptions, both in largely basaltic continental fields and in silicic LIPs, have emplaced deposits as or more voluminous than individual flood-basalt units, and with much greater dispersal.
8. Like flood basalt lava eruptions, ignimbrite-forming eruptions, as evident from the silicic LIPs, can also occur from fissures that tap stored magma through regional faults.
9. Many LIPs have been constructed by multiple long-lived effusive eruptions producing the typical flood basalt plateaus of CFBPs, volcanic passive margins and oceanic plateaus, but in other cases explosive volcanism has been the dominant eruptive style (e.g. Siberian, silicic LIPs). Dominance of explosive volcanism in LIPs is a result of either magma composition, where erupted compositions are dominantly silicic, and/or of environmental conditions at eruptive sites conducive for large-scale phreatomagmatism.
10. The spatial–temporal relationships between flood basalt lavas and volcaniclastic deposits during LIP formation can provide important constraints on the relative timing of LIP magmatism, extension, kilometre-scale uplift and palaeoenvironmental changes. These constraints can significantly advance our aim of understanding the Earth-scale causative processes of LIP events.

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