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Frontiers in Large Igneous Province research

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Abstract

Earth history is punctuated by events during which large volumes of mafic magmas were generated and emplaced by processes distinct from "normal" seafloor spreading and subduction-related magmatism. Large Igneous Provinces (LIPs) of Mesozoic and Cenozoic age are the best preserved, and comprise continental flood basalts, volcanic rifted margins, oceanic plateaus, ocean basin flood basalts, submarine ridges, ocean islands and seamount chains. Paleozoic and Proterozoic LIPs are typically more deeply eroded and are recognized by their exposed plumbing system of giant dyke swarms, sill provinces and layered intrusions. The most promising Archean LIP candidates (apart from the Fortescue and Ventersdorp platformal flood basalts) are those greenstone belts containing tholeiites with minor komatiites. Some LIPs have a substantial component of felsic rocks. Many LIPs can be linked to regional-scale uplift, continental rifting and breakup, climatic shifts that may result in extinction events, and Ni–Cu–PGE (platinum group element) ore deposits.

Some current frontiers in LIP research include:

(1) Testing various mantle plume and alternative hypotheses for the origin for LIPs.
(2) Characterizing individual LIPs in terms of (a) original volume and areal extent of their combined extrusive and intrusive components, (b) melt production rates, (c) plumbing system geometry, (d) nature of the mantle source region, and (e) links with ore deposits.
(3) Determining the distribution of LIPs in time (from Archean to Present) and in space (after continental reconstruction). This will allow assessment of proposed links between LIPs and supercontinent breakup, juvenile crust production, climatic excursions, and mass extinctions. It will also allow an evaluation of periodicity in the LIP record, the identification of clusters of LIPs, and postulated links with the reversal frequency of the Earth's magnetic field.
(4) Comparing the characteristics, origin and distribution of LIPs on Earth with planets lacking plate tectonics, such as Venus and Mars. Interplanetary comparison may also provide a better understanding of convective processes in the mantles of the inner planets.

In order to achieve rapid progress in these frontier areas, a global campaign is proposed, which would focus on high-precision geochronology, integrated with paleomagnetism and geochemistry. Most fundamentally, such a campaign could

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help hasten the determination of continental configurations in the Precambrian back to 2.5 Ga or greater. Such reconstructions are vital for the proper assessment of the LIP record, as well as providing first-order information related to all geodynamic processes.

Keywords: Large igneous province; Flood basalt; Oceanic plateau; Greenstone belt; Komatiite; Dyke swarm

1. Introduction

Large Igneous Provinces (LIPs) have been defined by Coffin and Eldholm (1994) as "massive crustal emplacements of predominantly mafic (Mg and Fe rich) extrusive and intrusive rock which originate via processes other than "normal" seafloor spreading. . . . [and] include continental flood basalts, volcanic passive margins, oceanic plateaus, submarine ridges, seamount groups and ocean basin flood basalts". For the present paper, we explicitly exclude large magmatic events associated with 'normal' subduction processes. We also note that some LIPs may have substantial felsic components (Campbell and Hill, 1988; Bryan et al., 2002).

LIPs occur throughout Earth's history (Isley and Abbott, 1999; Ernst and Buchan, 2001a, 2002; Isley and Abbott, 2002; Fig. 1). Those of Mesozoic and Cenozoic age are the best studied. They most prominently include both continental and oceanic flood basalts, and the former are commonly associated with volcanic rifted margins (e.g., Cox, 1980; White and McKenzie, 1989; Coffin and Eldholm, 1994, 2001; Storey, 1995; Menzies et al., 2002a). Important examples include the 62–56 Ma North Atlantic Igneous Province (NAIP), the 182 Ma Karoo–Ferrar event, the 250 Ma Siberian Traps (Fig. 2A) and the 122 Ma 'greater' Ontong Java event in the Pacific Ocean (Fig. 2B). Paleozoic and Proterozoic LIPs are typically more deeply eroded, exposing their plumbing system of giant dyke swarms, sill provinces and layered intrusions (e.g., Baragar, 1977; Fahrig, 1987; Ernst and Buchan, 1997a,b; 2001a,b; Pirajno, 2000; Condie, 2001). For example, the 1270 Ma Mackenzie giant radiating dyke swarm of the Canadian Shield fans over 100° of arc and extends for more than 2300 km from its focal point (Fig. 3A). In other cases (e.g., Fig. 3B), the distribution of sills, layered intrusions, dykes and volcanics appears less systematic. Flood basalts also occur in the Archean (e.g., Fortescue, Fig. 4A, and Ventersdorp volcanics; e.g., Eriksson et al., 2002). However, most Archean mafic–ultramafic magmatism occurs as deformed and fragmented packages termed greenstone belts. Those Archean greenstone belts that contain thick tholeiite sequences with minor komatiites (Campbell et al., 1989; Nelson, 1998; Arndt, 1999; Kerr et al., 2000; Tomlinson and Condie, 2001; Bleeker, 2002; Arndt, 2003) are excellent candidates for LIPs. An example in the Rae craton of northern Canada (Fig. 4B) extends for more than 1000 km and consists of 2730–2700 Ma komatiite-bearing greenstone belts of the Woodburn Lake, Prince Albert, and Mary River groups (e.g., Schau, 1997; Aspler et al., 1999; Zaleski et al., 2001; Skulski et al., 2003). LIPs have been recognized on Mars, Venus and the Moon where they provide complementary information to that from Earth (Head and Coffin, 1997; Ernst and Desnoyers, 2004).

Coffin and Eldholm (2001) defined the minimum size of a LIP in terms of an areal extent of at least 0.1 Mkm^2, thus including small continental flood basalts (e.g., Columbia River) and small oceanic plateaus (e.g., Shatsky Rise, Magellan Rise, Maud Rise, Wallaby Plateau). However, most Cenozoic and Mesozoic LIPs (Eldholm and Coffin, 2000; Courtillot and Renne, 2003) originally covered >1 Mkm^2. Courtillot and Renne (2003) estimated the volume of the extrusive components of most Mesozoic and Cenozoic continental flood basalts at 2–4 Mkm^3 and that of the Ontong Java oceanic plateau at 6 Mkm^3 (compared with 44.4 Mkm^3 for combined extrusive and intrusive components of Ontong Java). In the ocean basins, a LIP is manifested as oceanic crust that covers an area >0.1 Mkm^2 and has a thickness of at least 10 km, in contrast to the 7-km thickness of normal oceanic crust. In older LIPs where much or all of the volcanic component of the LIP has been lost to erosion, the area encompassed by the exposed plumbing system (i.e., dykes, sills and layered intrusions) is used as the measure of event size

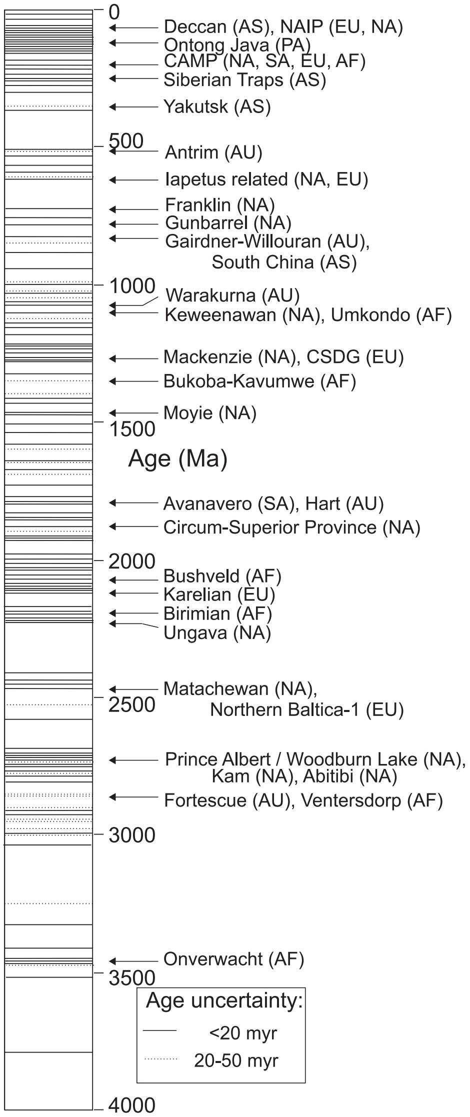

Fig. 1. Age spectrum ('bar code') of large igneous provinces through time. After Figs. 2b and 3b respectively in Ernst and Buchan (2001a, 2002). Includes events rated as "well-established and probable plume head mafic magmatic events" according to criteria in Ernst and Buchan (2001a). Selected events are labeled at the starting age of the main pulse. Locations for events are indicated as follows: NA is North America, SA is South America, EU is Europe, AF is Africa, AS is Asia, AU is Australia, and PA is Pacific Ocean. Names are mostly after Ernst and Buchan (2001a). Exceptions include Iapetus related (Puffer, 2002), Warakurna (Wingate et al., 2004), and Circum-Superior Province (www.largeigneousprovinces.org/LOM.html). Age uncertainty quoted at 2-sigma level.

(e.g., Yale and Carpenter, 1998), and is the basis for the estimated 7 Mkm^2 extent of the 200 Ma Central Atlantic Magmatic Province (CAMP) event (Marzoli et al., 1999), and the 2.7 Mkm^2 for the 1270 Ma Mackenzie event (Fig. 3A; Fahrig, 1987). Although the relative extent of flood basalts and the underlying plumbing system is uncertain, we apply the Coffin and Eldholm (2001) criterion of LIP size (>0.1 Mkm^2) for these older events dominated by exposed plumbing systems.

Most LIPs are emplaced within <10 Myr, with the bulk of the magmatism in <1 Myr. In some instances, magmatic activity may persist at a much-reduced rate for tens of millions of years and generate hotspots or seamount chains. Continental LIPs can often be subdivided into two distinct pulses, a prerift pulse and a postrift (or syn-rift) pulse (Campbell, 1998; Courtillot et al., 1999). The prerift pulse is linked to the arrival of a new mantle plume, and is the first eruptive phase from that plume. The postrift pulse, which includes seaward-dipping reflectors (White and McKenzie, 1989) and zones of high-velocity lower crust (Menzies et al., 2002b) is produced during rifting associated with continental breakup and can be interpreted to result from hot mantle in the plume head being drawn into the zone of rifting. The time gap between the two pulses varies from case to case, but is typically a few million to tens of million of years. The volume of magma produced during the second pulse may exceed that produced during the first (Campbell, 1998; Courtillot et al., 1999). Many alternative models for the origin of LIPs have been proposed, including back arc processes, shallow mantle 'edge' convection and bolide impact (see discussion in Section 2.1).

Independent of the debate regarding their origin, LIPs have special significance throughout Earth's history and potentially involve processes at all depths in the Earth: (1) core (possible links with magnetic superchrons and more subtle variations in magnetic reversal frequency), (2) mantle (links with individual plumes and clusters of plumes which sample both deep and shallow mantle and also provide evidence for the nature and location of geochemical reservoirs), (3) lithosphere (links with plate tectonics particularly rifting and ocean basin formation, and also associated with regional domal uplift and the generation of ore deposits), (4) biosphere (implications for climatic

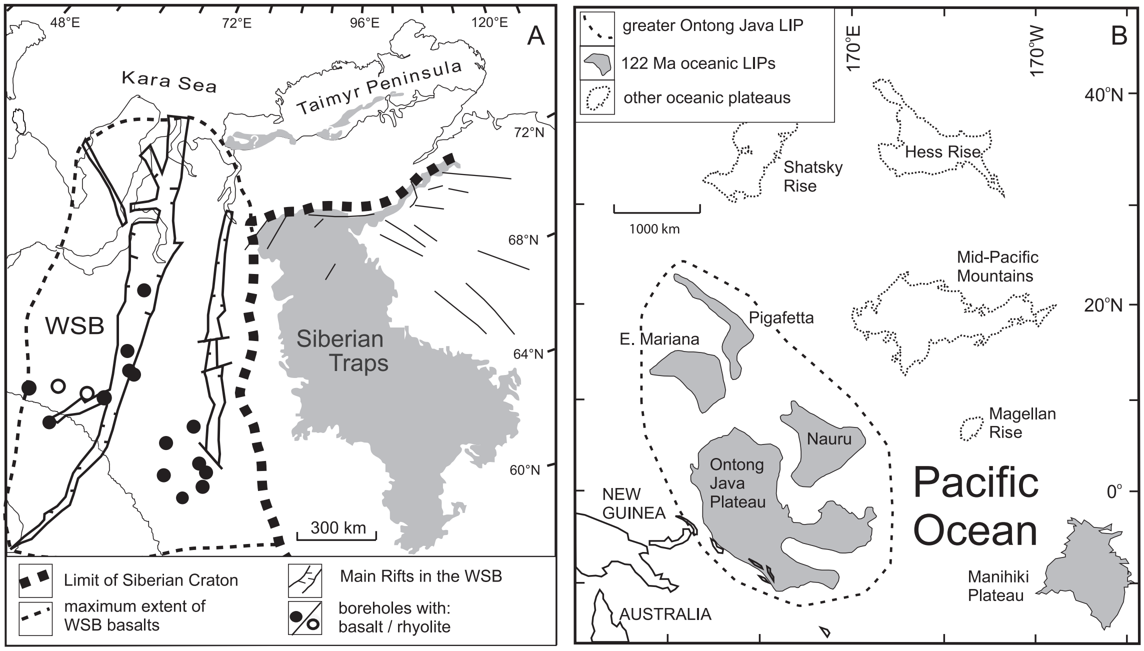

Fig. 2. Examples of young Mesozoic–Cenozoic LIPs, (A) 250 Ma Siberian Traps, modified after Reichow et al. (2004), with additions of dykes from Ernst and Buchan (2001b). WSB is West Siberian Basin, (B) 122 Ma greater Ontong Java Plateau (after Ingle and Coffin, 2004) and other oceanic LIPs (Eldholm and Coffin, 2000; Ernst and Buchan, 2001a): Manihiki Plateau (122 Ma), Shatsky Rise (147 Ma), Magellan Rise (145 Ma), Mid-Pacific Mountains (ca. 130–80 Ma), and Hess Rise (ca. 110 and ca. 100 Ma).

change and extinction events), and (5) extraterrestrial (comparison with the LIP record on Venus, Mars, Mercury and the Moon).

2. New frontiers

As research into Large Igneous Provinces expands, it is appropriate to highlight some important research 'frontiers'. In this paper, we focus on the following frontier issues (Table 1): (1) investigating and testing mantle plume vs. nonmantle plume origins; (2) characterizing LIPs in terms of their size, melt production rate throughout an event, plumbing system for emplacing and distributing magma in the crust, and in terms of geochemical character and location of mantle sources; (3) determining the distribution of LIPs in time and space after correcting for the reconstruction of continents; and (4) comparing LIPs on Earth, Venus, Mars, Mercury and the Moon.

Each frontier is discussed in turn below, and in a later section we propose a targeted LIP research agenda for the next 5 years focused on geochronology (integrated with paleomagnetism and geochemistry), that should accelerate our understanding of the LIP record through time.

2.1. Testing plume vs. nonplume origin of LIPs: developing critical tests

Many LIPs have been linked to the arrival at the base of the lithosphere of a mantle starting plume (Richards et al., 1989; Campbell and Griffiths, 1990; Maruyama, 1994; Ernst and Buchan, 2001a, 2003; Courtillot et al., 2003). Two end-member models were developed in the late 1980s. First, the plume head hypothesis of Richards et al. (1989) and Campbell and Griffiths (1990) suggested that flood basalts and oceanic plateaus are produced by melting within the head of a mantle starting plume, which originates at the base of the mantle. Producing the voluminous melting, before the onset of extension, requires the presence of an eclogite component in the plume head (Cordery et al., 1997; Campbell, 1998).

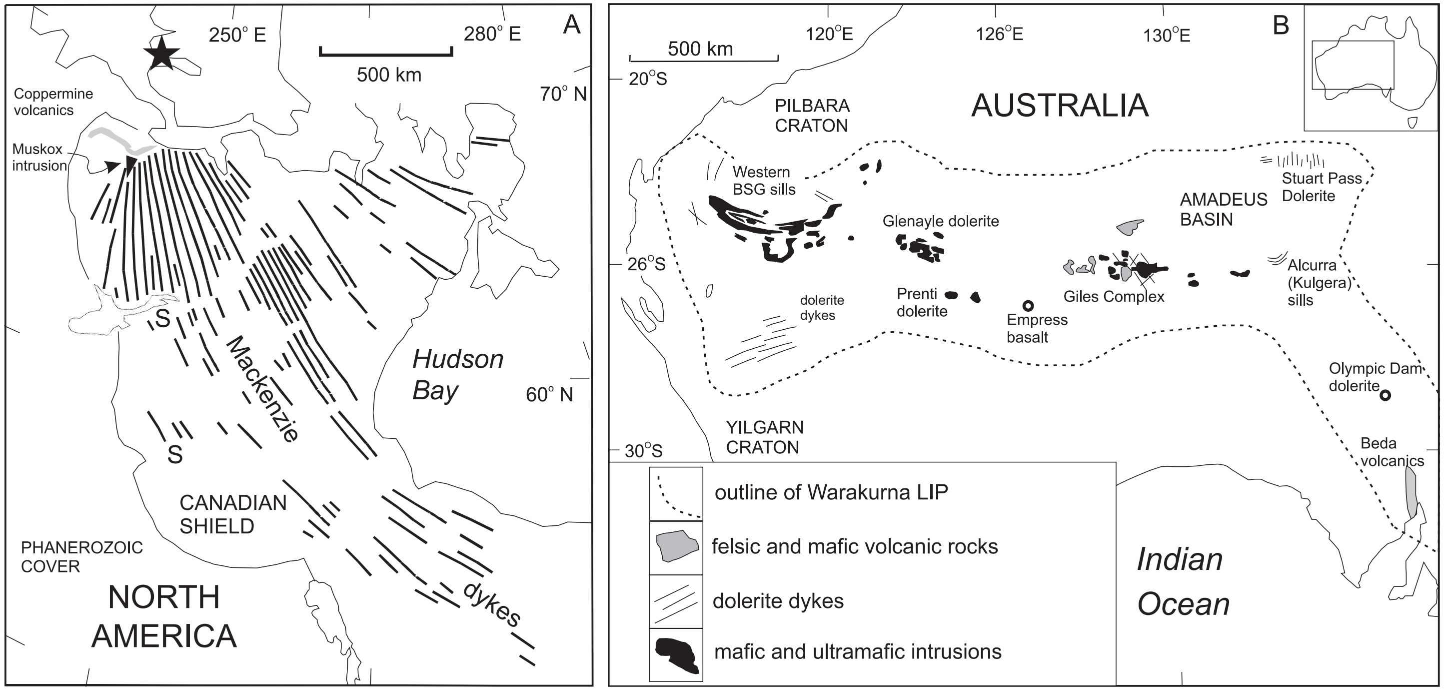

Fig. 3. Examples of Proterozoic LIPs where erosion has exposed the 'plumbing' system of dykes, sills and layered intrusions. (A) 1270 Ma Mackenzie event consisting of Mackenzie radiating dyke swarm, Muskox layered intrusion and Coppermine volcanics, and other regions containing sills ('s') (e.g., Baragar et al., 1996), (B) 1070 Ma Warakurna LIP of central Australia (Wingate et al., 2004).

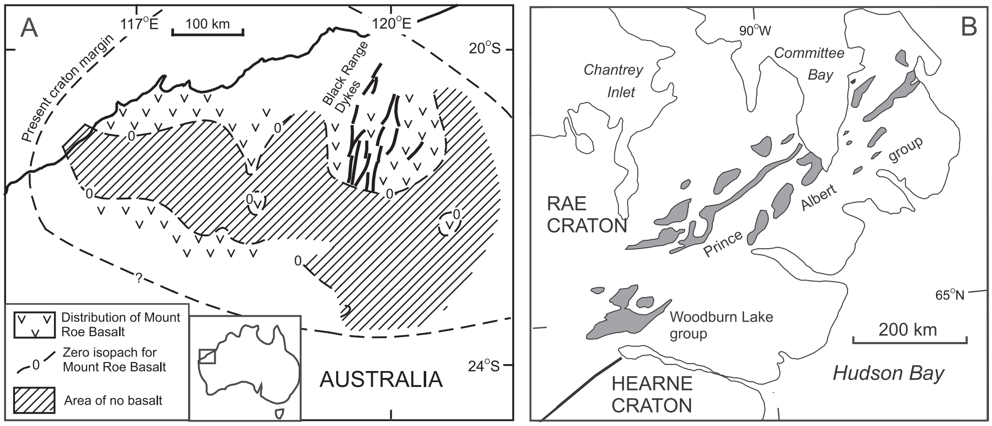

Fig. 4. Examples of Archean LIPs. (A) Part of Fortescue Group of Pilbara craton (Thorne and Trendall, 2001). Specifically shows the distribution of lowermost 2780 Ma Mount Roe sequence and associated Black Range feeder dykes. (B) 2730–2700 Ma Prince Albert and Woodburn Lake groups in the Rae Craton of northern Canada. Mary River Group located on Baffin Island (off the diagram to the northeast) is also probably correlative (Schau, 1997; Aspler et al., 1999; Zaleski et al., 2001; Skulski et al., 2003).

Second, White and McKenzie (1989) proposed that flood basalts are due to the build-up of plume material beneath continental lithosphere, followed by lithospheric extension associated with continental breakup. Flood volcanism was attributed to decompression melting associated with lithospheric extension. Other models of LIP formation involve lithospheric delamination (e.g., Elkins-Tanton and Hager, 2000; Şengör, 2001), a back-arc setting (e.g., Rivers and Corrigan, 2000; cf. Taylor, 1995), overriding of a spreading ridge by a continent (Gower and Krogh, 2002), enhanced mantle convection at the edge of the craton (edge-driven convection; King and Anderson, 1998; Anderson, 1995, 1998; Sheth, 1999; Foulger, 2002), lithospheric fracturing (e.g., Sheth, 1999) or meteorite impact (Boslough et al., 1996; Glikson, 1999; Jones et al., 2002; Abbott and Isley, 2002; Ingle and Coffin, 2004). The debate has intensified in recent years and there is a pressing need to set out clearly the predictions of various hypotheses and to apply appropriate tests to distinguish between these hypotheses.

LIPs are the product of anomalously high melt production rates. Three scenarios, high water content, decompression and anomalously high mantle temperatures, can lead to high melt production rates in the mantle (e.g., Campbell, 2001) Water in the mantle lowers the mantle solidus, and can potentially produce higher melt production rates. However, water ($\pm CO_2$) contents are low in the upper mantle, particularly in 'intraplate' settings away from subduction zones. Hence, water-induced melting is unlikely to be a general cause of LIPs. Decompression melting of the mantle, as seen in mid-ocean ridges, does produce large volumes of melt but melt rates are not regarded as anomalously high because, regardless of the rate of spreading, they produce normal oceanic crust with a thickness of 7 ± 1 km (White et al., 1992). 'Normal' oceanic crust is excluded from consideration as a LIP by definition (see Introduction). Although extension and associated decompression of typical upper mantle may contribute to high melt production rates, they are in themselves insufficient to form LIPs. We conclude that any successful hypothesis for the production of LIPs must include melting of anomalously hot mantle. An alternative view has been expressed that the upper mantle may have a considerable range in ambient temperatures (e.g., Anderson, 1995), which would make plumes unnecessary for the formation of LIPs. However, computer models of mantle convection consistently show that temperature variations within the upper mantle are small compared with variation associated with plumes and subduction zones (see Davies, 1999; Davies and Richards, 1992). Finally, we note that fertile zones within the mantle will produce more melt than refractory zones, and variations in mantle fertility may contribute to variations in melt production between LIPs. For instance, entrained eclogite streaks will tend to melt first (e.g., Gibson, 2002; Campbell, 1998).

Table 1
Frontiers in Large Igneous Province (LIP) research

Frontiers	Goal	Approach
Section 2.1.1: Investigating mantle plume and non-mantle plume origins: developing critical tests	To identify and test the various predictions of hypotheses for LIPs origin: (1) deep mantle plumes, (2) transition zone boundary plumes, (3) rifting, (4) 'edge-convection' from thick to adjacent thin lithosphere, and (5) meteorite impacts	Document diagnostic characteristics of LIPs such as presence/absence of uplift, extent and timing of uplift, length of volcanic margins, timing of rifting, giant radiating dyke swarms, presence/absence of shock features, and seismic character of underlying mantle
Section 2.2: Characterizing LIPs in terms of:		
Original volume and areal extent of combined intrusive and extrusive components	To determine the range in size of LIPs and explore implications for source models To search for variations through time	Identify additional units that belong to a LIP Reconstruct the original size of a LIP fragmented by plate tectonics
Melt production rates	To quantify rate of mass transfer to the crust through the course of a LIP event To search for variations through time	Date and determine the volume of LIP components precisely
Plumbing system geometry	To locate 'entry points' at which magma enters the crust/lithosphere from the astenospheric mantle To identify how and where magma is distributed in the crust	Map geometry of feeder dyke swarms, sills and layered intrusions
Nature of mantle source region	To quantify contributions from different geochemical reservoirs	Use incompatible element and isotope geochemistry
Link with ore deposits	Develop improved exploration model for PGE–Ni–Cu Apply to diamond exploration	Integrate ore deposit characteristics with improved understanding of LIP plumbing system
Section 2.3: Determining distribution of LIPs in space and time (from Archean to Present)	To assess proposed links between LIPs and supercontinent breakup, juvenile crust production, climatic excursions, extinction events, and magnetic superchrons. To evaluate periodicity To identify clusters of LIPs	High precision dating of LIPs Reconstruct LIPs fragmented by plate tectonics (to distinguish single fragmented LIPs from clusters of independent LIPs) Time series analysis of LIP record
Section 2.4: Comparing characteristics, origin and distribution of LIPs on Earth, Venus, Mars, Mercury and the Moon	To compare LIP record on plate tectonic planets (Earth, early Mars?) and non-plate tectonic planets (especially Venus and Mars) To better understand convective processes in the mantle	Map LIP components and associated features on the four terrestrial planets and the Moon Search Earth record for analogues to Venusian coronae

Criteria (adapted from Courtillot et al., 2003; Kerr et al., 2000; Ernst and Buchan, 2001a,b, 2003) for interpreting a deep mantle origin for a LIP include the following: evidence from mantle tomography for thermal anomalies extending into the deep mantle, a link with an age progressive hotspot track, link to uplift (marked in the young record by a large buoyancy flux), compositional geochemical arguments (including the presence of high Mg rocks, and high $^{3}He/^{4}He$) and the presence of a giant radiating dyke swarm. Each of these classes of criteria is discussed below.

2.1.1. Imaging with seismic tomography

Seismic methods offer the most direct method of testing whether the zones of anomalously hot mantle, predicted by the plume hypothesis, underlie present-day hotspots that can be backtracked to their initial LIP events (e.g., Zhao, 2001; Maguire et al., 2003; Montelli et al., 2004). Although it is has been difficult to image plume tails (e.g., Ritsema and Allen, 2003) recent technical improvements have enabled better resolution beneath hotspots (e.g., Montelli et al., 2004). Of the hotspots that can be backtracked to a LIP (see Table 1 of Courtillot et al., 2003), the following have a deep mantle origin (Montelli et al., 2004): Afar (linked with the Afar/Ethiopian LIP), Crozet (possibly linked with the Karoo LIP), Easter (linked with the Mid-Pacific Mountains LIP), Reunion (linked with the Deccan LIP), Kerguelen (linked with the Kerguelen and Rajmahal LIPs), and Louisville (linked with the Ontong Java LIP). Hotspots that have not been traced deep into the lower mantle include Iceland (linked with the North Atlantic Igneous Province LIP), and Galapagos (linked with the Caribbean–Colombian LIP). The Yellowstone hotspot (Columbia River LIP) has no discernable seismic anomaly beneath it. We note, however, that plumes must originate from a thermal boundary layer and there are unlikely to be two hot thermal boundary layers in the mantle (Griffiths and Turner, 1998). Seismic evidence suggests that slabs and most plumes pass through the 660-km seismic discontinuity, which makes it unlikely that plumes originate from a thermal boundary layer above the core. The failure of Montelli et al. (2004) to follow the Iceland and Galapagos plumes to the core–mantle boundary may reflect the limitation of the method. Receiver function analysis suggests that Iceland and a region to the southwest of Galapagos show transition zone anomalies (Li et al., 2003b). Other seismic analysis suggests that the Iceland plume does pass through the 660-km seismic discontinuity (Bijwaard and Spakman, 1999; Shen et al., 2002). One of the frontiers of LIP research is to obtain improved seismic images of plume tails and to follow them to their boundary layer source.

Plume tails, because of their small diameter, are difficult to resolve using standard seismic tomographic methods. Plume heads, which are larger and remain hotter than the adjacent mantle for up to 1000 Myr (Griffiths and Campbell, 1990), are an easier seismic target, and should be detectable below LIPs for at least 200 Myr. Fossil plume heads are interpreted to underlie the 65 Ma Deccan, 122 Ma Ontong Java, and 133 Ma Paraná LIPs (VanDecar et al., 1995; Kennett and Widiyantoro, 1999; Richardson et al., 2000; Klosko et al., 2001; Gomes and Okal, 2003). In the older Paleozoic or Precambrian record, underplated fossil plume heads can only be recognized using seismic methods if they exhibit a residual compositional anomaly.

2.1.2. Domal uplift

Broad domal uplift (0.5–2 km over a diameter of ~1000 km) provides an unambiguous test for the presence of a buoyant plume head below the lithosphere (e.g., Griffiths and Campbell, 1991; Şengör, 2001; Farnetani and Richards, 1994; He et al., 2003). The magnitude of uplift, prior to volcanism, is related to the temperature of the mantle source area, but uplift may also be affected by melting, although this effect is transitory (occurring only during the time the melt remains in the mantle). The lateral extent and shape of the uplift provides an estimate of the extent of buoyant underlying mantle and therefore the size and shape of the thermal anomaly in the mantle that produces the uplift (Griffiths and Campbell, 1991). The rate of uplift provides an estimate of the viscosity of the sublithospheric mantle (Campbell, 1998). It is important to note that the plume hypothesis only predicts magnitude and timing of uplift prior to volcanism, and that the pattern of uplift after the onset of volcanism can be affected by other factors. For instance, lateral redistribution of magma away from the central region can reduce or remove the prevolcanic uplift (Campbell, 2001).

The best studied LIPs are young (of Cenozoic–Mesozoic age). Cox (1989) identified domical uplift in the Paraná, Deccan and Karoo LIPs by the presence of a radiating pattern of river drainage. However, the uplift pattern is strongly affected by characteristics of the lithosphere (Monnereau et al., 1993) and associated rifting (e.g., Cox, 1989), and requires detailed mapping for proper assessment (Rainbird and Ernst, 2001; Ukstins Peate et al., 2003a). The Central Atlantic Magmatic Province was associated with uplift having a diameter of about 2000 km. This is inferred from the interruption of the sedimentation pattern in preexisting rift basins

(Hill, 1991; Rainbird and Ernst, 2001). An uplift diameter of about 1600 km is determined from stratigraphic patterns associated with the 258 Ma Emeishan event (He et al., 2003). In contrast, the Siberian Traps seem to lack associated uplift (Czamanske et al., 1998). However, this may be because the area selected for study was not close to the centre of the plume (e.g., Saunders et al., 2005), or because magma was redistributed away from the plume center (cf. Campbell, 2001). Uplift is first felt at the surface when the plume head is at a depth of ~1000 km and reaches a maximum when the top of the head reaches ~200 km (Griffiths and Campbell, 1991). The difference between the start of uplift and maximum uplift depends on the viscosity of the upper mantle but, for reasonable assumptions, is probably about 5 to 10 Myr (Campbell, 1998).

Given the first-order importance of large-scale domal uplift (1000 to 2000 km across) for paleo-drainage patterns, it is essential that more LIP events be investigated for domal uplift. In this regard, the most important sediments to study are those immediately underlying the base of the flood basalt pile (e.g., Xu et al., 2004). It is also crucial to obtain uplift information on a regional scale in order to ensure that rift flank uplift (e.g., Menzies et al., 2002b) and orogenic uplift (e.g., England and Molnar, 1990) can be distinguished from domal plume-generated uplift.

2.1.3. Giant radiating dyke swarm

Giant radiating dyke swarms (often >1000 km in radius) are associated with many continental LIP events, and are typically interpreted to result from the arrival at the base of the lithosphere of a mantle plume (e.g., May, 1971; Halls, 1982; Fahrig, 1987; Ernst and Buchan, 1997a,b; Ernst et al., 1995, 2001; Wilson and Head, 2002). Many giant radiating swarms converge toward an associated breakup margin, and it is typically concluded that the remaining segments of the radiating swarm are to be found on the cratonic blocks that rifted away. Support for primary radiating geometry of swarms comes from Venus and Mars (one-plate planets) where radiating patterns are more continuous and complete (Grosfils and Head, 1994; Mège and Masson, 1996; Ernst et al., 2001). In particular on Venus, most radiating systems fan over more than 270°.

Giant radiating swarms have only been observed on continents, but it is predicted that they also occur in an oceanic setting (Ernst and Buchan, 1997b). A possible oceanic example is observed in association with the Galapagos hotspot. Specifically, submarine volcanic chains radiate northward over an angle of more than 90° from the Galapagos hotspot (Fig. 9 in Sinton et al., 2003). The measured length of the volcanic chains ranges between 150 and 250 km, but these lengths are minimum since the volcanic chains terminate against the E–W Galapagos Spreading Centre [Ridge]. Various origins are offered for these volcanic chains, including the suggestion that they are fed by underlying dykes (p. 19 in Sinton et al., 2003).

2.1.4. Relative timing of rifting and volcanism

Most Mesozoic and Cenozoic LIPs are associated with continental rifting and breakup events (e.g., Burke and Dewey, 1973; Morgan, 1981; Hill, 1991; Courtillot et al., 1999; Şengör and Natal'in, 2001). They are recognized by volcanic rifted margins (e.g., Storey et al., 1992; Menzies et al., 2002a,b) that are associated with seaward dipping reflectors and thick high velocity lower crust marking underplated material. In contrast, nonvolcanic rifted volcanic margins (Wilson et al., 2001; Russell and Whitmarsh, 2003) represent the LIP-free transition to oceanic crust of normal thickness.

A similar relationship is also true for older Precambrian LIPs wherein the exposed feeder system (radiating dyke swarm) typically focuses on a cratonic breakup margin (Ernst and Buchan, 1997b). If volcanism precedes rifting, then a plume origin is required (Campbell and Griffiths, 1990; Campbell, 1998). However, this does not preclude a second pulse of volcanism associated with the onset of rifting. Other models, such as decompression melting during rifting above asthenosphere with elevated temperatures (White and McKenzie, 1989) would require that rifting precede LIP volcanism.

In most cases, initial LIP magmatism precedes rifting (Campbell, 1998; Courtillot et al., 1999; Fig. 3 in Menzies et al., 2002b), favouring a plume origin. However, there are cases where the relationship between rifting and volcanism is ambiguous (Menzies et al., 2002b). While the 200 Ma CAMP event precedes rifting and opening of the Central Atlantic (e.g., Hames et al., 2003), a precursor stage of rifting (forming the Newark basins of eastern North America) precedes CAMP magmatism by about 30 Myr. Formation of the Newark basins has been linked to the CAMP LIP (e.g., McHone, 2000). However, 30 Myr before the onset of volcanism the plume head would be deep in the lower mantle where it cannot influence surface topography (Griffiths and Campbell, 1991). Therefore, Hill (1991) considered formation of the Newark basins as an independent event.

Other models of LIP formation that involve convection between adjacent lithospheric plates of different thickness ("edge-driven convection"; Anderson, 1998), lithospheric delamination (Elkins-Tanton and Hager, 2000; Şengör, 2001) or bolide impact (e.g., Ingle and Coffin, 2004) are not directly tested by the relative timing of rifting and volcanism.

2.1.5. Shock features in basement to LIP

Major extinction events have been correlated with both LIPs (e.g., Courtillot et al., 1996; Courtillot and Renne, 2003) and impacts (e.g., Alvarez et al., 1980; Hildebrand et al., 1991; Koeberl and MacLeod, 2002), leading some authors (e.g., Stothers and Rampino, 1990; Glikson, 1999; Jones et al., 2002; Abbott and Isley, 2002) to suggest a link between impacts and LIPs. The most dramatic example is the recent proposal (Ingle and Coffin, 2004) of an impact origin for the world's largest LIP, the 58 Mkm^3 'greater Ontong Java' LIP (Fig. 2B) that includes the Ontong Java oceanic plateau as well as nearby coeval ocean basin flood basalts.

Theoretical arguments have been offered in favour of (Jones et al., 2002) and against (Ivanov and Melosh, 2003) an impact origin for LIPs. Conclusive evidence for an impact origin would be the recognition of shatter cones, shocked quartz and other features indicative of transient ultrahigh pressures. Only one magmatic province, the 1850 Ma Sudbury Complex of the Canadian Shield has been conclusively linked with an impact, in part based on the presence of shock features (e.g., Lightfoot and Naldrett, 1994), but this province is of sub-LIP scale. A novel endogenic explanation for shock features has recently been suggested; it is hypothesized that catastrophic CO_2 and SO_2 gas eruptions through the lithosphere ("Verneshots") associated with continental flood basalts may also induce shock features (Phipps Morgan et al., 2004).

2.1.6. Deep mantle geochemistry

It is difficult to recognize plume involvement based on geochemistry because (1) there are multiple mantle reservoirs that can contribute (see Section 2.2.4) and there is uncertainty about which of these (if any, Anderson, 1998, 2001) are located in the deep mantle. Recognition of plume origin is also complicated by the possible contamination of magmas on their ascent through the lithosphere and crust. Another significant factor is the possibility of secular changes in the composition of mantle source areas through time (e.g., Campbell and Griffiths, 1992; Campbell, 2002; Condie, 2003a).

Nevertheless, some generalizations are possible (e.g. recent reviews in Campbell, 1998, 2001; Condie, 2001; Ernst and Buchan, 2003). Uncontaminated plume-generated basaltic rocks will normally have flat REE patterns or LREE-enriched patterns and lack negative Nb, Ta and Ti anomalies. The presence of dry high-MgO magmas (picrites and komatiites) is considered diagnostic of plumes. Plots such as CaO/Al_2O_3 vs. Fe can be used to monitor depth and degree of melting in a plume. Contamination by continental crust or lithosphere can impart subduction-type signatures and lead to the misidentification of basalts as arc related. High $^3He/^4He>10\ R/R_A$ and high Os isotopic ratios are probably diagnostic as they imply either a primordial deep source or recycled oceanic crust. Furthermore, osmium isotopes can be used to distinguish between crustal and lithospheric contamination.

2.2. Characterization of LIPs

Very few LIPs have been fully characterized in terms of their size, the variation in melt production rate throughout the LIP event, the geochemical character and inferred distribution of mantle source areas, the plumbing system for emplacing and distributing magma in the crust, and the link with ore deposits. Yet, such information is fundamental to understanding the generation and geodynamic setting of LIPs and their effects, for instance, on climatic changes. In this section, we review our current understanding of some basic characteristics of LIPs.

2.2.1. Original volume and areal extent

Size is an important parameter in the study of LIPs. It has been used to draw conclusions about the flux of mafic magma from the mantle, the relationship between continental and oceanic LIPs, the origin of LIPs (e.g., plume vs. nonplume), the variation in plume size (largely controlled by the depth in the mantle at which the plume originates) and the climatic effects of LIPs (e.g., White and McKenzie, 1989; 1995; Davies, 1999; Campbell, 1998, 2001).

However, there are a number of uncertainties in estimating LIP size. In the older record, only remnants of the extrusive component have escaped erosion, and the LIP is typically fragmented by plate tectonic processes. Even in the younger record, where erosion and plate tectonic effects are less pronounced, the estimated size of a LIP may increase dramatically as additional components are recognized through better dating, modeling and stratigraphic correlation. A recent example is a doubling in the recognized size of the 250 Ma Siberian Traps based on dating of basalts recovered by drill core from beneath the West Siberian basin (Campbell et al., 1992; Reichow et al., 2002, 2005). An additional source of uncertainty in determining the volume of LIPs is the amount of basaltic material that has been intruded into or underplated beneath the crust (e.g., White and McKenzie, 1989; Cox, 1993; Trumbull et al., 2002). For oceanic LIPs, this can potentially be estimated if crustal thickness is known from seismic studies. For continental LIPs, there is the difficulty of distinguishing intrusive and underplated components from the host basement rocks. Better geophysical methods are required to solve this problem. Xenolith populations in kimberlites provide a way of dating deep crustal and subcrustal (lithospheric) components of LIPs and subcreted plumes (e.g., Davis, 1997; Griffin et al., 1999; Kopylova and Caro, 2004). Because the magma volume of LIPs is difficult to assess, their areal extent is often used instead to estimate their size (Coffin and Eldholm, 2001).

As noted above, most Mesozoic and Cenozoic LIPs cover >1 Mkm^2. One of the largest is the 200 Ma CAMP event which covers an area of 7 Mkm^2 in the Atlantic bordering continents in a pre-Atlantic ocean configuration. Most of this areal extent is represented by the associated giant radiating dyke swarm (e.g., May, 1971; Marzoli et al., 1999). The Ontong Java plateau covers an area of 2 Mkm^2, but is about 33 km thick and represents a magmatic volume of about 45 Mkm^3 (Coffin and Eldholm, 1994; Neal et al., 1997; Gladczenko et al., 1997). An even larger size estimate of 55 Mkm^3 is obtained when nearby ocean floor basalts are included (Ingle and Coffin, 2004).

The estimation of the maximum size is more problematic for the Proterozoic and Archean LIPs due to erosion. Many LIPs are known only from their feeder dykes, sills and layered intrusions, and many are fragmented by plate tectonics. The 1270 Ma Mackenzie giant radiating dyke swarm and associated Coppermine volcanics and Muskox layered intrusion cover an area of 2.7 Mkm^2 (Fahrig, 1987; LeCheminant and Heaman, 1989; Baragar et al., 1996). The Mackenzie swarm fans over 100° and its focal point on Victoria Island (Canadian Arctic) likely marks the centre of a plume. The remaining segments of the Mackenzie radiating swarm are likely to be found on the blocks that subsequently rifted away from northern Canada during formation of a northern Ocean.

In the Archean, the identification of LIPs and an estimate of their size is even more complicated. Apart from rare intact flood basalts [Fortescue (Fig. 4A) of Australia and Ventersdorp of South Africa; e.g., Eriksson et al., 2002], most Archean magmatism is preserved as deformed and metamorphosed fragments that are termed greenstone belts. As mentioned above, the best candidates for Archean LIPs are the greenstone belts containing thick tholeiite sequences with minor komatiites, which are strong analogues for the packages of dominant tholeiites and minor picrites in young flood basalts (e.g., Gibson, 2002). The presence of komatiites precludes an arc origin, and also suggests a hotter source than for normal Archean MORB (e.g., Campbell and Griffiths, 1992; Arndt, 2003). Thus, komatiite-bearing greenstone belts satisfy part of the LIP definition; the only remaining criterion is the emplacement of large volumes of magma. Because of extensive deformation and fragmentation, correlation within and between greenstone belts has been difficult. Nevertheless, LIP-scale magmatic events have been identified by detailed geochronological and stratigraphic work in the Abitibi greenstone belt of the Superior Province, the Prince Albert group and related units in the Rae Province (Fig. 4B), and the Kam Group of the Slave Province, all in the Canadian Shield, and also the Bababudan belt of the Dharwar craton in India (Bleeker, 2002; Ayer et al., 2002; Skulski et al., 2003; Chadwick et al., 1985a,b).

An important event occurred in the late Archean. Tholeiitic–komatiitic greenstone belts with ages of 2730–2700 Ma are recognized from many cratons around the world (e.g., de Wit and Ashwal, 1997; Tomlinson and Condie, 2001; Ernst and Buchan, 2001a; Bleeker, 2003). It will require continental reconstructions (Section 2.3.3) to test whether these coeval sequences originally belonged to a single LIP, or to a cluster of coeval but independent magmatic events (Ernst and Buchan, 2002).

2.2.2. Melt production rate

One of the most important parameters associated with LIP events is the melt production rate, including its peak value and its variation through the course of an event. Melt production increases with increasing size of the underlying region of hot sublithospheric mantle. For example, mantle plume models predict that a plume head has greater melt production than a plume tail. Melt production also increases with decreasing lithospheric thickness (e.g., in a rift zone; White and McKenzie, 1995; Campbell and Davies, in press). In addition, melt production in oceanic LIPs is substantially greater for those associated with a spreading ridge axis than those located away from a ridge axis (e.g., Ito et al., 2003; Campbell and Davies, in press).

Many events, such as the Columbia River event, feature a single pulse of magmatism, followed by a protracted period of magmatism at a much lower rate that is linked to a plume tail (Tolan et al., 1989). Other events may exhibit two pulses of high volume activity, a first phase that can be linked to a mantle plume head and a second if the plume head is drawn into a rift zone during continental breakup (Campbell, 1998). Events exhibiting two pulses of significant activity include the Keweenawan, CAMP and North Atlantic Igneous Province (NAIP) LIPs. The initial peak of the Keweenawan event occurred at 1108–1105 Ma and was followed at 1102–1094 Ma by a second peak which may correspond to the initiation of rifting (Ojakangas et al., 2001; Fig. 2 in Hanson et al., 2004). Volcanism continued at a reduced rate for at least the next 15 Myr. The CAMP event began at 200 Ma and was followed by the emplacement of seaward dipping reflectors about 10–25 Myr later (Benson, 2003) or perhaps only 5 Myr later (Sahabi et al., 2004) in association with rifting and breakup. The two pulses of the NAIP event occurred at about 61 and 56 Ma, only a few million years apart (e.g., Saunders et al., 1997). An even more complicated pattern appears to apply in some Archean examples where more than two pulses are observed. For instance, in the Archean Abitibi greenstone belt, tholeiite–komatiite sequences include the Pacaud (2750–2735 Ma), Stoughton–Roquemaure (2725–2720 Ma), Kidd–Munro (2718–2710 Ma) and Tisdale (2710–2703 Ma) assemblages (Ayer et al., 2002). Sproule et al. (2002) interpret these to represent either multiple plumes or a single plume with a prolonged and punctuated history.

In some LIPs, there is evidence for minor magmatism a few million years in advance of the main pulse. In such cases, the precurser phase of magmatism may be related to plume arrival, while the main pulse occurs after sufficient time has elapsed for the plume head to ascend into the lithosphere by the mechanism of lithospheric delamination or erosion. An example, based on Ar–Ar dating, is the magmatism associated with the Paraná–Etendeka LIP in which a minor phase of magmatism of 138–135 Ma precedes the main pulse of activity at 134–129 Ma (Peate, 1997). Another example is from the NAIP where minor seamount magmatism at 70 Ma may have preceded the initial main pulse at 62 Ma (O'Connor et al., 2000).

After the burst(s) of initial magmatism, LIP magmatism can continue for perhaps 100 Myr or more, but at a much reduced rate. This process is recorded in ocean basins as seamount chains (and submarine ridges), which exhibit progressively younger ages along the seamount chain in a direction away from the LIP. However, continental plume–tail magmatism is more difficult to recognize.

Unfortunately, current estimates of melt production rate for most LIPs are very imprecise owing to the uncertainties related to event size (see Section 2.2.1), and the small number of precise ages available for most events (Campbell and Davies, in press). There is a need to generate curves of melt production rate vs.

time for LIPs with a range of ages, as well as for LIPs in different tectonic settings (e.g., oceanic vs. continental; rift vs. nonrift settings; thickened lithosphere vs. 'thinspots'). The duration of oceanic LIP events can be particularly difficult to determine because sampling is often restricted to the top 100 m, which may only represent the waning stage of volcanism. Deep drill holes are required to obtain representative samples of oceanic plateaus and aseismic ridges. A simpler approach is to study obducted sections like those associated with oceanic plateaus, such as the Caribbean–Colombian LIP (Kerr et al., 1998; Révillon et al., 2000). Dating distal ash layers on the seafloor can sometimes be used to establish the duration of LIP events. For example, a strong constraint on the duration of the Ontong Java event comes from dating ocean sediment tuff horizons to the north of the plateau (Tarduno et al., 1991). In another example, dating and trace element geochemistry of ash from the Ethiopian LIP, which was carried into the Indian Ocean, has allowed correlation of individual ash layers with specific Afar flood basalt units (Ukstins Peate et al., 2003b).

2.2.3. Plumbing system of LIPs

Two important questions are related to the plumbing system for LIPs: (1) How does magma enter the lithosphere and crust from asthenospheric mantle source areas? (2) How is magma subsequently redistributed within the crust (as dykes, sills and layered intrusions), and onto the surface (as volcanics)? Several end-member processes are apparent.

Some LIPs, particularly those associated with giant radiating dyke swarms, appear to have entered the lithosphere through a single restricted 'entry zone'. For instance, the 1270 Ma Mackenzie event (Fig. 3A), has a localized centre defined by the focal region of a giant radiating dyke swarm (Fahrig, 1987; Ernst et al., 1995; Baragar et al., 1996). Magma is inferred to have entered the lithosphere from the mantle over a relatively small area (perhaps 500 km in diameter) at the focal point of the swarm (Ernst and Buchan, 1997a). It then rose to form shallow magma chambers from which magma was distributed laterally through the crust via the radiating dyke swarm out to distances of >2000 km.

Laterally propagating dykes can feed lava flows and emplace sills at great distances from the presumed lithospheric entry zone at the focus of the swarm (Ernst et al., 1995; Ernst and Buchan, 1997a). For example, it has been proposed on the basis of age correlations and matching geochemistry that the ca. 2220 Ma giant radiating Ungava dyke swarm of the Canadian Shield fed the Nipissing sill province in the Huronian basin 1500–1800 km away from the swarm centre (Buchan et al., 1998).

Some LIPs appear to have multiple lithospheric entry zones. Such LIPs may be derived from a single mantle source (e.g., a plume) or a cluster of coeval mantle sources (see Section 2.3). For example, paleomagnetism and dating demonstrate a short duration for the 182 Ma Karoo–Ferrar event, which is widespread throughout southern Africa and Antarctica with a felsic equivalent in South America (Hargraves et al., 1997; Storey et al., 2001), but geochemistry and the distribution of dykes and rifts zones indicate multiple entry zones (e.g., Ernst and Buchan, 2002). Specifically, the distinct geochemical differences between Karoo- and Ferrar-type magmas favour derivation from distinct source regions in the mantle (e.g., Elliot and Fleming, 2000). The geometry of rifts and radiating dyke swarms requires two lithospheric entry zones, the Nuanetsi and Zambesi (Burke and Dewey, 1973; Ernst et al., 1995). The Weddell Sea triple rift junction, with associated volcanic margins, may represent a third entry zone (e.g., Storey et al., 2001) and the large Dufek–Forrestal layered mafic–ultramafic intrusion with its newly discovered dyke swarms, a fourth (Ferris et al., 2003).

According to a third scenario, mantle magmatism is emplaced into the crust along a linear zone. This case includes rift margin packages with seaward dipping reflectors (Menzies et al., 2002b) and linear rift-parallel dyke swarms (e.g., Fahrig, 1987; Ernst and Buchan, 1997b), large volume magmatism found in a back-arc setting (e.g., Rivers and Corrigan, 2000), hot-lines or sheets (e.g., White, 1992; Meyers et al., 1998, Al-Kindi et al., 2003) and lines of coronae (on Venus; see Section 2.4). In this regard, it is important to note that no physical mechanism has been suggested to produce a linear heat source in the mantle. 'Line plumes' are unstable in a fluid that has a strong temperature-dependent viscosity, such as the mantle and break up into axisymmetric (cylindrical) plumes (Loper and Stacey, 1983). Lines of axisymmetric plumes are possible in the mantle, line plumes are not. The most realistic explanation for linear belts of intraplate magmatism is decompression melting produced during extension (White and McKenzie, 1989).

Another aspect of the plumbing system relates to sublithospheric movement of source (e.g., plume) material (Sleep, 2002). Ebinger and Sleep (1998) proposed that plume material from the Afar centre was distributed to other centres in western Africa by channeling along sublithospheric 'valleys' generated by previous rifting events. Northeast-directed channeling for the 200 Ma CAMP event was suggested by Wilson (1997). A more general scenario of plume-material moving along sublithospheric topography toward lithospheric thin-spots was described by Thompson and Gibson (1991).

Recent studies of the North Atlantic Igneous Province (NAIP) reveal some other aspects of the plumbing system. A magnetic fabric study from the NAIP on eastern Greenland suggests that discrete high-level magmatic centres developed along the rift and that these local magma chambers fed the passive margin volcanics via lateral injection along dyke swarms (e.g., Callott et al., 2001). These spaced magma chambers presumably locate spaced zones of magma entry into the crust. Another aspect related to the plumbing system is the role of dyke swarms in causing segmentation along rifted margins, and in causing magmatic extension at the continent–ocean boundary (Ebinger and Casey, 2001; Klausen and Larsen, 2002).

2.2.4. Nature of mantle source region (geochemistry)

LIP geochemistry can be used to characterize mantle sources, provided it is not overprinted by lithosphere or crustal contamination, a common problem with continental LIPs (Arndt and Christensen, 1992). The involvement of known mantle reservoirs, depleted MORB mantle (DMM), enriched mantle I (EMI), enriched mantle II (EMII), mantle with a high time-integrated U–Pb ratio (HIMU) and focus zone (FOZO), can be assessed using isotopes and trace elements (e.g., Hart et al., 1992; Blichert-Toft and White, 2001; Condie, 2001; van Keken et al., 2002; Ernst and Buchan, 2003; Workman et al., 2004). DMM represents an upper mantle reservoir, the source for mid-ocean ridge basalts. EMI, EMII and HIMU are enriched components that may represent recycled oceanic lithosphere ± a small contribution from the continental crust. FOZO is a depleted reservoir that also has elevated He and Os anomalies. There is debate as to whether it represents an ancient and deep depleted source (e.g., van Keken et al., 2002; Workman et al., 2004) or is linked to shallow mantle processes (e.g., Meibom et al., 2003). The role of recycled crustal material is increasingly being recognized as an important component in the mantle. Of particular current interest is the role of eclogite in the source area (Campbell, 1998; Takahashi et al., 1998; Cordery et al., 1997), which is derived from fossil subducted slabs. Using trace elements and isotopes, it should prove possible to model the cycling of components (Campbell, 2002), and progressive change in mantle reservoirs (e.g., Condie, 2003a). Osmium isotopes are especially valuable in identifying mantle source regions that have a high concentration of eclogite representing subducted oceanic crust (e.g., Hauri, 2002; van Keken et al., 2002). A likely scenario is that all of the mantle has been processed through mid-oceanic ridges so that it now consists of eclogite and depleted harzburgite together with minor amounts of sediments, mixed in different proportions to produce the observed geochemical end-members (DMM, HIMU, EMI, EMII and FOZO; e.g., Hart et al., 1992; Hofmann, 1997; van Keken et al., 2002). Geochemistry can also be used to assess the depth of partial melting (White and McKenzie, 1995), and in particular whether melting has taken place beneath a lithospheric lid (i.e., thick overlying lithosphere; Ellam, 1992; Lassiter and DePaolo, 1997). However, these generalizations only apply if the mantle is assumed to be homogeneous. If, at the scale of melting, the mantle is a heterogeneous mixture of eclogite and depleted harzburgite (e.g., Kerr et al., 1998; Révillon et al., 2002), then existing melting models are difficult to apply.

Data from high-Mg rocks are particularly useful for characterizing mantle sources because their high degree of partial melting more faithfully preserves the trace element ratios in the source area (Campbell, 1998). High-Mg magmas also come from the hottest part of the plume, a portion of the plume that has not entrained cooler mantle and is likely to be most representative of the plume boundary layer source.

The origin of high-Ti and low-Ti magmas (e.g., p. 162 in Condie, 2001) is an important but unresolved problem. Both types are found in continental flood basalts and their distribution has been attributed to melting of continental mantle lithosphere of different composition (Gibson et al., 1996), control by lithospheric thickness (e.g., Arndt et al., 1993) or different degrees of partial melting correlated with proximity to the hotter plume centre (Fodor, 1987; Xu et al., 2004). In contrast, low-Ti basalts are unknown from OIBs, but are found in oceanic plateaus, such as Ontong Java and the Caribbean–Colombian Plateau, which suggests that the control may be a plume head vs. a plume tail rather than a continental vs. an oceanic setting.

2.2.5. Links with ore deposits

There are strong links between LIPs and Ni–Cu–PGE (platinum group element) ore deposits (e.g., Barnes et al., 1997; Naldrett, 1999; Pirajno, 2000, 2004; Schissel and Smail, 2001; Diakov et al., 2002; Crocket, 2002; Maier et al., 2003; Ernst and Hulbert, 2003). Prominent examples are the Noril'sk deposits (of the 250 Ma Siberian Trap event) which produce 70% of the world's palladium and the 2060 Ma Bushveld intrusion which is the largest known mafic–ultramafic intrusion and the worlds most important producer of platinum and chrome (Naldrett, 1997, 1999). Archean greenstone belts which contain komatiites (and are considered as LIP candidates in this paper) are an important source of Ni–Cu–PGE ores (Lesher and Keays, 2002). An important frontier issue is the development of better exploration models which integrate the characteristics of Ni–Cu–PGE deposits with an improving understanding of LIP plumbing systems.

LIPs have not been linked to kimberlites. However, some authors (e.g., Haggerty, 1999; Pirajno, 2004) have proposed a deep mantle origin for kimberlites, suggesting a link with mantle plumes. In addition, others (e.g., Heaman and Kjarsgaard, 2000; Schissel and Smail, 2001) have linked kimberlites with hotspot tracks in both time and space.

2.3. Distribution of LIPs in space and time

As the LIP database improves, it should become possible to look at the temporal and spatial record of LIPs to answer some fundamental geodynamic questions. Is there a periodicity in the record that links with exogenic or terrestrial forcing processes? Are LIPs emplaced in clusters either geographically or temporally? Is there a systematic spacing or distribution of LIPs along incipient rift zones. Specific aspects are discussed below.

2.3.1. Time series analysis of LIP record

The distribution of LIPs through time is still poorly understood, since the Proterozoic and Archean record is incomplete. However, using the well-defined LIP record for the past 150 Myr, Coffin and Eldholm (2001) obtained a rate of about 1 continental LIP per 10 Myr. This record includes oceanic LIPs which may be lost during subduction associated with subsequent ocean closure (Cloos, 1993) and hence are poorly preserved in the older record. If only continental LIPs are counted, a rate of about 1 LIP per 20 Myr applies for this young record. On a broader time scale, the preliminary database of Ernst and Buchan (2001a, 2002) can be used to infer a similar emplacement rate of 1 LIP per 20 Myr throughout the Proterozoic and Phanerozoic. The record is relatively continuous, although a few significant gaps are observed. It is not clear if these gaps are real or merely artifacts of an incomplete database. There is also some clustering of events (see “LIP clusters” in Section 2.3.3).

There have been numerous attempts to identify cycles and trends in the spectrum of LIPs through time (Rampino and Caldeira, 1993; Yale and Carpenter, 1998; Isley and Abbott, 1999, 2002; Ernst and Buchan, 2001a, 2002; Prokoph et al., 2004). Prokoph et al. (2004) suggested cycles of 730–550, 330, 170, 100 and 30 Myr which are broadly similar to those in Isley and Abbott (2002). However, given the broadband nature and weak persistence of most of these cycles, their significance and link with forcing functions (e.g., Isley and Abbott, 2002 and references therein), such as a ~30 Myr cometary impact cycle, a 270 Myr galactic year, a 500–300 Myr supercontinent cycle and ~800 Myr resonance between tidal and free oscillations of the core, remains unclear. Future time series analyses on improved versions of the LIP database (and appropriate subsets) will be required to test the robustness of proposed cycles, and to identify underlying forcing functions.

It is also important to compare the LIP spectrum with possible aperiodic phenomena. For instance, the link between deep-sourced plumes and core processes can be monitored by the timing of magnetic superchrons and more subtle variations in magnetic reversal frequency (Larson and Olson, 1991; Johnson et al., 1995; Condie, 2001, p. 240) and also by variations in paleointensity (Macouin et al., 2003). There is also the potential of looking for links between the timing of plume generation and mantle avalanche events (Schubert and Tackley, 1995; Condie, 1998; Schubert et al., 2001) and mantle overturn events (Stein and Hofmann, 1994).

2.3.2. Links to climatic changes and extinction events

One of the most exciting frontiers in LIP research is testing the potential effect of LIPs on climate, and in particular, proposed links with extinction events.

There is a widespread view that at least some extinction events are due to the climatic effects of meteorite impacts. In particular, the end-Cretaceous extinction has been linked with the Chicxulub Crater on the Yucatan Peninsula, Mexico (Alvarez et al., 1980; Hildebrand et al., 1991). Other extinction events have been linked more tenuously to impacts, such as the end-Triassic (Olsen et al., 2002) and end-Permian (Becker et al., 2004). Other authors (e.g., Courtillot et al., 1996; Wignall, 2001; Courtillot and Renne, 2003) have proposed that extinctions are caused by the climatic effects of LIPs, because most extinction events during the last 350 Myr have ages that correlate within a few million years or less with those of major LIP events. For example, the Siberian Traps, CAMP and Deccan Traps are of the same age as the end-Permian, end-Triassic and end-Cretaceous extinction events, respectively. However, the mechanisms linking LIPs, extreme climate changes and extinctions are complex and not well modelled (e.g., Wignall, 2001). It has also been suggested that the end-Permian, end-Triassic and end-Cretaceous extinction events result from the superimposed environmental effects of coeval impact and LIP events (White and Saunders, 2005).

The study of LIPs and extinction events is part of a broader inquiry into the climatic effects of LIPs. Emplacement of a LIP may release massive amounts of SO_2 into the atmosphere, causing global cooling and acid rain, and CO_2, which has a strong greenhouse effect (e.g., Veevers, 1990; Campbell et al., 1992; Kerr, 1998; Wignall, 2001; Condie, 2001; Ernst and Buchan, 2003). Furthermore, a minor temperature increase can potentially trigger a massive gas hydrate melting and thus a LIP event can have an effect far greater than its direct contribution to climate change (Wignall, 2001; Jahren, 2002). Paradoxically, a continental LIP can also lead to a temperature decrease due to increased weathering and fixing of CO_2 which leads to a drop in atmospheric CO_2 (Goddéris et al., 2003). Oceanic LIPs can interrupt ocean circulation patterns, and cause displacement of water onto continental shelves (e.g., Kerr, 1998; Wignall, 2001). Furthermore, distinguishing environmental effects related to LIPs from other causes (such as plate tectonics, variation in solar cycle, obliquity) is difficult. Assessing the environmental effects of LIPs is therefore complicated and represents an important frontier.

An important approach is to compare LIP timing (Ernst and Buchan, 2001a) with seawater isotopic chemistry which is recorded by marine carbonates (Veizer et al., 1999; Shields and Veizer, 2002). A preliminary comparison suggests a weak correlation between the timing of LIPs and negative excursions of $^{87}Sr/^{86}Sr$ (Prokoph et al., 2004, and unpublished data). This presumably reflects an increased contribution to seawater of 'mantle-like' $^{87}Sr/^{86}Sr$ through weathering of LIPs. Such comparisons will become more meaningful as the biostratigraphic and absolute timescales become better linked (Gradstein et al., 2004).

2.3.3. LIP clusters ("superplume events"), and links with supercontinent breakup and juvenile crust production

The recognition of clusters of LIPs in time and (or) in space has important geodynamic implications (e.g., Ernst and Buchan, 2002). Spatial clusters of LIPs have been linked with supercontinent breakup (Storey, 1995; Li et al., 2003a). Specifically, at least 5 LIPs have been linked with the progressive breakup of Gondwana (Storey, 1995), and several are linked with breakup of the proposed Rodinia supercontinent (Li et al., 2003a). In addition, bursts of juvenile crustal production have been suggested to reflect increases in oceanic LIP production (Condie, 2001). An example might be the coveal Keweenawan and Umkondo magmatism of North America and southern Africa, respectively (Hanson et al., 2004). These events are coeval but are emplaced while an ocean was closing and the Grenville orogen was forming between them. Do the Keweenawan and Umkondo events represent a single LIP resulting from a single deep mantle plume or do they represent separate LIPs?

In a broader sense, at about 30 times since 3.5 Ga, coeval mafic magmatism is recognized on more than one continental block (Ernst and Buchan, 2001a, 2002). Reliable continental reconstructions are required in order to assess which of these represent single fragmented LIPs and which represent clusters of independent LIPs. A variety of reconstructions have been proposed through this time period (e.g., Rogers, 1996; Bleeker, 2003). Those reconstructions which are younger than ~400 Myr are relatively well constrained on the basis of paleomagnetism, and geological considerations and, in the past 200 Myr, by analysis of seafloor spreading anomalies. In the earlier record, however, reconstructions are much more speculative and controversial. For example, numerous versions of the ca. 1.3–0.7 Ga supercontinent Rodinia have been proposed (e.g., Hoffman, 1991; Moores, 1991; Dalziel, 1991; Meert and Van der Voo, 2001; Wingate et al., 2002; Condie, 2003b; Pisarevsky et al., 2001, 2003; Weil et al., 2004). Prior to Rodinia, in the early Mesoproterozoic, an even more speculative supercontinent has been termed Nuna (e.g., Hofmann, 1997; Bleeker, 2003) or Colombia (Rogers and Santosh, 2002; Zhao et al., 2002). Still earlier, a variety of late Archean supercontinents have been described (e.g., Williams et al., 1991; Rogers, 1996; Aspler and Chiarenzelli, 1998; Bleeker, 2003). Testing such reconstructions requires a rigorous application of integrated paleomagnetism and high-precision geochronology (Buchan et al., 2000). To date, such tests are only available for a few specific cratonic blocks at specific times (Buchan et al., 2001; Wingate and Giddings, 2000).

2.4. Comparing characteristics, origin and distribution of LIPs on Earth, Venus and Mars

Mars, Venus, and possibly the Moon and Mercury have a robust history of LIP-scale basaltic volcanism (Head and Coffin, 1997). However, unlike Earth, they appear to lack plate tectonics, except perhaps in the early history of Mars. Therefore, both Venus and Mars provide a natural laboratory to study intraplate magmatism in the absence of plate boundary effects, and should contribute to our understanding of LIPs on Earth (Ernst and Desnoyers, 2004). However, the lessons from Mars and Venus for LIPs on Earth may not be direct. The presence of a 'stagnant lithospheric lid' on Mars and Venus, and the early loss of water on Venus impose a different mantle viscosity structure which may result in systematic differences between intraplate magmatism on Venus and Mars from that on Earth (e.g., Schubert et al., 2001; Ernst and Desnoyers, 2004).

Examples of LIPs on Mars include the massive volcanic edifices of the Tharsis Montes region, the largest of which is Olympus Mons (Head and Coffin, 1997). Individual martian flows can be 1800 km long (Fuller and Head, 2003). On Venus, LIPs are represented by large flow fields averaging 0.2 Mkm^2 (Magee and Head, 2001), and large volcanic edifices hundreds of kilometers across (e.g., Hansen et al., 1997; Crumpler and Aubele, 2000; Ernst and Desnoyers, 2004). In addition, an important class of LIPs on Venus is associated with widespread annular structures (diameter 50–2600 km) termed coronae, which are interpreted to result from mantle diapirs, and are often distributed along rift zones (e.g., Hansen et al., 1997). Coronae also appear to be present on Mars (Wattters and Janes, 1995), but their identification on Earth is speculative (Herrick, 1999; Jellinek et al., 2002; Johnson and Richards, 2003; Ernst and Desnoyers, 2004 and references therein).

3. LIPs—a suggested research agenda

The research frontiers that are described above can only be fully addressed with an expanded LIP record and much better continental reconstructions. LIP compilations (Isley and Abbott, 1999; Tomlinson and Condie, 2001; Ernst and Buchan, 2001a; Prokoph et al., 2004) contain many major events that remain undated and/or poorly characterized, especially in the Precambrian and early Paleozoic. Such basic characteristics as LIP size, spatial distribution and relationship to plate boundaries are unclear in the absence of reconstruction information.

The Large Igneous Provinces (LIPs) Commission (www.largeigneousprovinces.org) is proposing a 5-year research agenda to improve the pre-Mesozoic LIP database through a concerted programme of high-precision (U–Pb and Ar–Ar) geochronology integrated with paleomagnetism and geochemistry. Such studies, to be carried out on a large number of mafic units (dyke swarms, sills, layered intrusions and volcanics), and related felsic units, would allow a much better understanding of the age, duration and extent of LIP events and, in particular, the correlation of events between continental blocks through improved reconstructions. Without robust reconstructions, many of the frontier LIP issues discussed in this paper cannot be adequately addressed.

The integration of precise dating and paleomagnetism of LIPs is a key component of the research being undertaken by the Tectonics Special Research Centre, based at the University of Western Australia (www.tsrc.uwa.edu.au), in its mission of "discovering the supercontinents of which Australia has been part in the past 3000 million years". Recently, a campaign of similar scope has been proposed for Canada (Bleeker, in press).

Additionally, an improved understanding of the continental reconstruction record would allow topical ideas such as global glaciation associated with 'Snowball Earth' (Hoffman and Schrag, 2002), true polar wander (Evans, 2003) and the use of LIP event ages as boundary markers ('golden spikes') in the Precambrian time scale (Okulitch, 2002; Gradstein et al., 2004) to be better assessed.

Key elements of the suggested research agenda are summarised below. More details will be available on the LIPs Commission website (www.largeigneousprovinces.org).

3.1. Objectives

(1) To survey the global record of mafic magmatism in order to identify key LIP components that require precise dating, paleomagnetism, geochemistry and other studies. Emphasis will be placed on events that:

(a) are widespread on a given continent and may, if precisely dated, contribute to the establishment of 'bar codes' for each of the ~35 Archean cratonic fragments that have been identified around the world (Bleeker, 2003). For example, preliminary 'bar codes' for the Superior and Slave provinces of the Canadian Shield suggest that these blocks were not in proximity to one another prior to ca. 1850 Ma (Bleeker, 2003).
(b) should be present on other blocks (because of proximity to a cratonic margin). These represent key targets because of the potential for contributing to defining continental reconstructions. Examples include radiating swarms such as the 1270 Ma Mackenzie swarm of the Laurentian craton (Fig. 3A).
(c) occur on different continental fragments and are coeval. They can be used to test continental reconstructions using paleomagnetism. These events should be unmetamorphosed so that their magnetic remanence is likely to be primary (see key paleopole concept of Buchan et al., 2000). An example is a reconstruction based on the comparison of the precisely dated paleopoles of the Mackenzie event of Laurentia and the coeval Central Scandinavian Dolerite Complex of Baltica (Buchan et al., 2000).
(d) occur at pivotal points in Earth history. An example is the mafic events that occur at the Phanerozoic/Proterozoic boundary associated with the opening of the Iapetus Ocean (e.g., Puffer, 2002; event nos. 52–55 in Ernst and Buchan, 2001a).
(e) should be the target of widespread dating in order to provide essential information on the rate of melt emplacement as a function of time. This in turn may relate to the origin and setting of LIPs. A representative set might be chosen to include Mesozoic/Cenozoic, Proterozoic and Archean continental LIPs, as well as a young oceanic LIP and an older Archean analogue.

(2) To develop guidelines for the efficient use of the various methodologies (e.g., geochronology, paleomagnetism and geochemistry) that will be employed during the project. For example, to reduce ambiguity in interpretation, sampling for geochronology, paleomagnetism and geochemistry should be integrated at the site level (e.g., so that paleopoles and geochemistry are obtained for the dating sites). Other aspects to be considered include ongoing improvements in baddeleyite separation techniques for U–Pb dating (e.g., Söderlund and Johansson, 2002), intercalibration between Ar–Ar and the U–Pb systems (e.g., Renne et al., 1998; Min et al., 2000) and increased use of high precision trace element geochemistry (now routinely available using ICP-MS technology) as fingerprints for LIP events.

(3) To establish regional/national LIP working groups. This is particularly important for regions where

LIPs are poorly dated and characterized, such as parts of Africa, South America and Asia.

(4) To obtain significant new data on LIPs. Possible targets include: 500 to 1000 new precise ages, 50 new paleomagnetic poles (including tests for primary remanence) from precisely dated LIPs in order to evaluate reconstructions and full suites of geochemical analyses from 100 LIPs.

(5) To promote networking among the LIP researchers in order to increase the effectiveness of fundraising in support of the expanded LIP research.

(6) To coordinate the distribution of information and results by ongoing conference sessions, publications and on the website of the LIPs Commission (www.largeigneousprovinces.org).

Acknowledgments

This paper has been inspired by the work of the LIP community, and represents some suggested foci for the activities of the Large Igneous Provinces (LIPs) Commission of the International Association of Volcanology and Chemistry of the Earth's Interior (IAVCEI), with its current theme of "Large Igneous Provinces in time and space" (www.largeigneousprovinces.org). This is Geological Survey of Canada contribution no. 2004119. We appreciate the detailed comments from reviewers Martin Menzies, Nick Arndt, and Wouter Bleeker, and volume editor Andrew Kerr, and discussions with Jian Lin, and Franco Pirajno.

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