LARGE IGNEOUS PROVINCES

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Introduction

Large igneous provinces (LIPs) are massive crustal emplacements of predominantly iron- and magnesium-rich (mafic) rock that form by processes other than normal seafloor spreading; they are the dominant form of near-surface magmatism on the terrestrial planets and moons of our solar system. On the Earth's surface, LIP rocks are readily distinguishable from the products of the two other major types of magmatism – mid-ocean ridge magmatism and arc magmatism – on the basis of petrologic, geochemical, geochronological, geophysical, and physical volcanological data. LIPs occur both on the continents and in the oceans, and include continental flood basalts, volcanic passive margins, oceanic plateaus, submarine ridges, seamount chains, and ocean-basin flood basalts (**Figure 1** and **Table 1**). LIPs and their contemporary small-scale analogues, hotspot volcanoes, are commonly attributed to decompression melting of hot low-density mantle material ascending from the Earth's interior in mantle plumes, and thus provide a window onto mantle processes. This type of magmatism currently accounts for about 10% of the mass and energy flux from the Earth's deep interior to its crust. The flux may have been higher in the past, but is episodic over geological time, in contrast to the relatively steady-state activity at seafloor spreading centres. Such episodicity reveals dynamic non-steady-state circulation within the Earth's mantle, perhaps

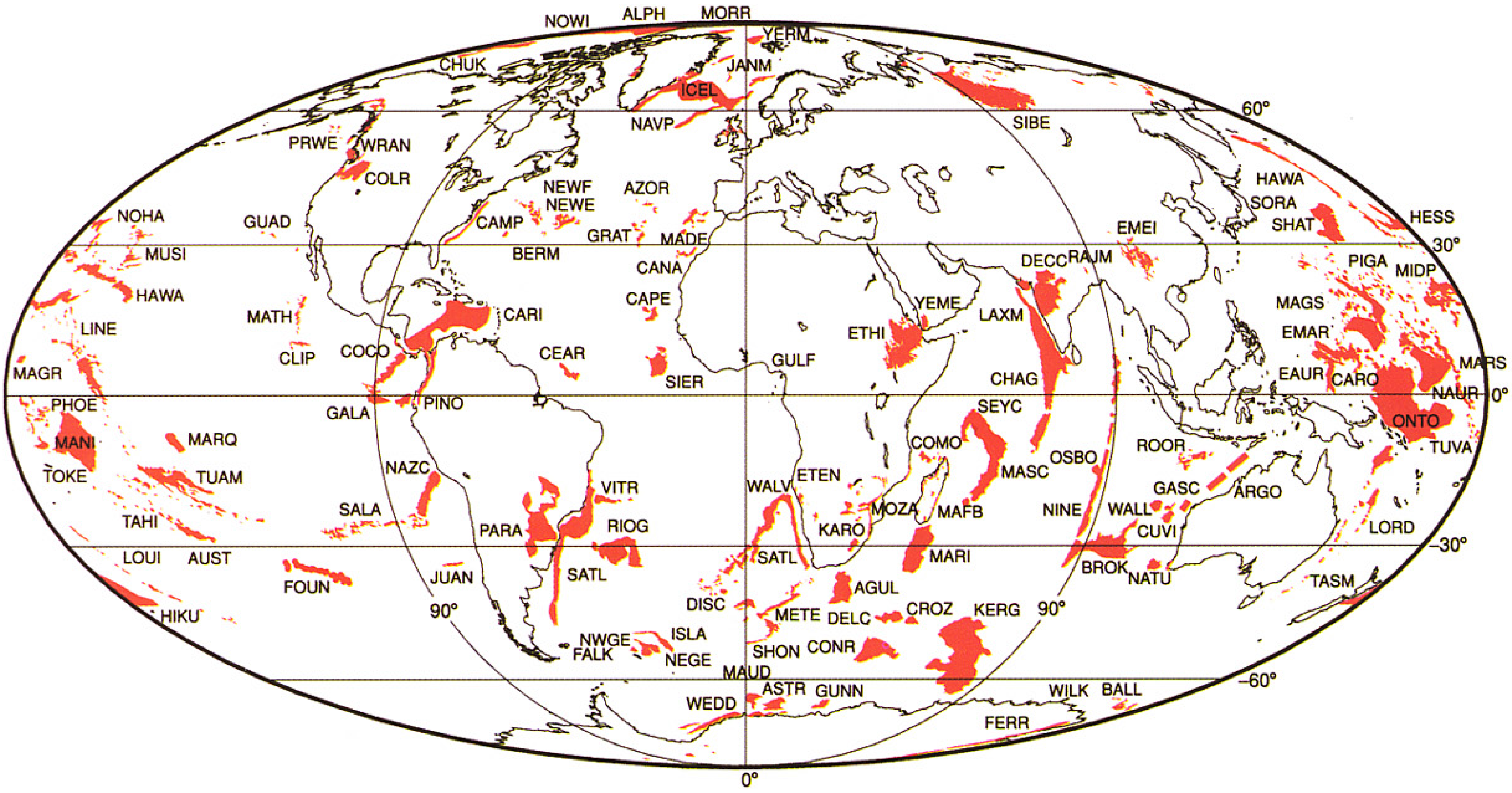

Figure 1 Phanerozoic global LIP distribution (red), with LIPs labelled (*see* **Table 1**).

Table 1 Large igneous provinces

Large igneous province	*Abbreviation (see* **Figure 1***)*	*Type*
Agulhas Ridge	AGUL	SR
Alpha-Mendeleyev Ridge	ALPH	SR/OP
Argo Basin	ARGO	VM
Astrid Ridge	ASTR	VM
Austral Seamounts	AUST	SMT
Azores	AZOR	SMT
Balleny Islands	BALL	SMT
Bermuda Rise	BERM	OP
Broken Ridge	BROK	OP
Canary Islands	CANA	SMT
Cape Verde Rise	CAPE	OP
Caribbean Flood Basalt	CARI	OBFB (partly accreted)
Caroline Seamounts	CARO	SMT
Ceara Rise	CEAR	OP
Central Atlantic Magmatic Province (VM only)	CAMP	CFB/VM
Chagos–Laccadive Ridge	CHAG	SR
Chukchi Plateau	CHUK	VM
Clipperton Seamounts	CLIP	SMT
Cocos Ridge	COCO	SR
Columbia River Basalt	COLR	CFB
Comores Archipelago	COMO	SMT
Conrad Rise	CONR	OP
Crozet Plateau	CROZ	OP
Cuvier (Wallaby) Plateau	CUVI	VM
Deccan Traps	DECC	CFB/VM
Del Caño Rise	DELC	OP
Discovery Seamounts	DISC	SMT
East Mariana Basin	EMAR	OBFB
Eauripik Rise	EAUR	OP
Emeishan Basalts	EMEI	CFB
Etendeka	ETEN	CFB
Ethiopian Flood Basalt	ETHI	CFB
Falkland Plateau	FALK	VM
Ferrar Basalts	FERR	CFB
Foundation Seamounts	FOUN	SMT
Galapagos–Carnegie Ridge	GALA	SMT/SR
Gascoyne Margin	GASC	VM
Great Meteor–Atlantis Seamounts	GRAT	SMT
Guadelupe Seamount Chain	GUAD	SMT
Gulf of Guinea	GULF	VM
Gunnerus Ridge	GUNN	VM
Hawaiian–Emperor Seamounts	HAWA	SMT
Hess Rise	HESS	OP
Hikurangi Plateau	HIKU	OP
Iceland–Greenland–Scotland Ridge	ICEL	OP/SR
Islas Orcadas Rise	ISLA	SR
Jan Mayen Ridge	JANM	VM
Juan Fernandez Archipelago	JUAN	SMT
Karoo	KARO	CFB
Kerguelen Plateau	KERG	OP/VM
Laxmi Ridge	LAXM	VM
Line Islands	LINE	SMT
Lord Howe Rise Seamounts	LORD	SMT
Louisville Ridge	LOUI	SMT
Madagascar Flood Basalts	MAFB	CFB
Madagascar Ridge	MARI	SR/VM?
Madeira Rise	MADE	OP
Magellan Rise	MAGR	OP
Magellan Seamounts	MAGS	SMT
Manihiki Plateau	MANI	OP
Marquesas Islands	MARQ	SMT
Marshall Gilbert Seamounts	MARS	SMT
Mascarene Plateau	MASC	OP
Mathematicians Seamounts	MATH	SMT
Maud Rise	MAUD	OP
Meteor Rise	METE	SR
Mid-Pacific Mountains	MIDP	SMT
Morris Jesup Rise	MORR	VM
Mozambique Basin	MOZA	VM
Musicians Seamounts	MUSI	SMT
Naturaliste Plateau	NATU	VM
Nauru Basin	NAUR	OBFB
Nazca Ridge	NAZC	SR
New England Seamounts	NEWE	SMT
Newfoundland Ridge	NEWF	VM
Ninetyeast Ridge	NINE	SR
North Atlantic Volcanic Province	NAVP	CFB
North-east Georgia Rise	NEGE	OP
North-west Georgia Rise	NWGE	OP
North-west Hawaiian Ridge	NOHA	SR/SMT
Northwind Ridge	NOWI	SR
Ontong Java Plateau	ONTO	OP (partly accreted)
Osborn Knoll	OSBO	OP
Paraná	PARA	CFB
Phoenix Seamounts	PHOE	SMT
Pigafetta Basin	PIGA	OBFB
Piñón Formation (Ecuador)	PINO	OP (accreted)
Pratt–Welker Seamounts	PRWE	SMT
Rajmahal Traps	RAJM	CFB
Rio Grande Rise	RIOG	OP
Roo Rise	ROOR	OP
Sala y Gomez Ridge	SALA	SR
Seychelles Bank	SEYC	VM
Shatsky Rise	SHAT	OP
Shona Ridge	SHON	SR
Siberian Traps	SIBE	CFB
Sierra Leone Rise	SIER	OP
Sorachi Plateau (Japan)	SORA	OP (accreted)
South Atlantic Margins	SATL	VM
Tahiti	TAHI	SMT
Tasmantid Seamounts	TASM	SMT
Tokelau Seamounts	TOKE	SMT
Tuamotu Archipelago	TUAM	SMT
Tuvalu Seamounts	TUVA	SMT
Vitória–Trindade Ridge	VITR	SR/SMT
Wallaby Plateau (Zenith Seamount)	WALL	OP
Walvis Ridge	WALV	SR
Weddell Sea	WEDD	VM
Wilkes Land Margin	WILK	VM
Wrangellia	WRAN	OP (accreted)
Yemen Plateau Basalts	YEME	CFB
Yermak Plateau	YERM	VM

CFB, continental flood basalt; OBFB, ocean-basin flood basalt; OP, oceanic plateau; SMT, seamount; SR, submarine ridge; VM, volcanic margin.

extending far back into Earth history, and suggests a strong potential for LIP emplacements to contribute to, if not instigate, major environmental changes.

Composition, Physical Volcanology, Crustal Structure, and Mantle Roots

LIPs are defined by the characteristics of their dominantly iron- and magnesium-rich (mafic) extrusive rocks; these typically consist of subhorizontal subaerial basalt flows. Individual flows can extend for hundreds of kilometres, be tens to hundreds of metres thick, and have volumes as great as 10^4–$10^5\,km^3$. Silica-rich rocks also occur as lavas and intrusive rocks and are usually associated with the initial and late stages of LIP magmatic activity. Relative to mid-ocean-ridge basalts, LIPs include higher MgO lavas, basalts with more diverse major-element compositions, rocks with more common fractionated components, both alkalic and tholeiitic differentiates, basalts with predominantly flat light-rare-earth-element patterns, and lavas erupted in both subaerial and submarine settings.

As the extrusive component of LIPs is the most accessible for study, nearly all of our knowledge of LIPs is derived from the lavas forming their uppermost crusts. The extrusive layer may exceed 10 km in thickness. On the basis of geophysical, predominantly seismic, data from LIPs and from comparisons with normal oceanic crust, LIP crust beneath the extrusive layer is believed to consist of an intrusive layer and a lower crustal body, characterized by P-wave velocities of 7.0–7.6 km s^{-1}, at the base of the crust (**Figure 2**). Beneath continental crust this body may be considered as a magmatically underplated layer. Seismic-wave velocities suggest an intrusive layer that is probably gabbroic and a lower crust that is ultramafic. If the LIP forms on pre-existing continental or oceanic crust or along a divergent plate boundary, dikes and sills are probably common in the middle and upper crust. The maximum crustal thickness, including extrusive and intrusive layers and the lower crustal body, of an oceanic LIP is about 35 km, as determined from seismic and gravity studies of the Ontong Java Plateau (**Figure 1** and **Table 1**).

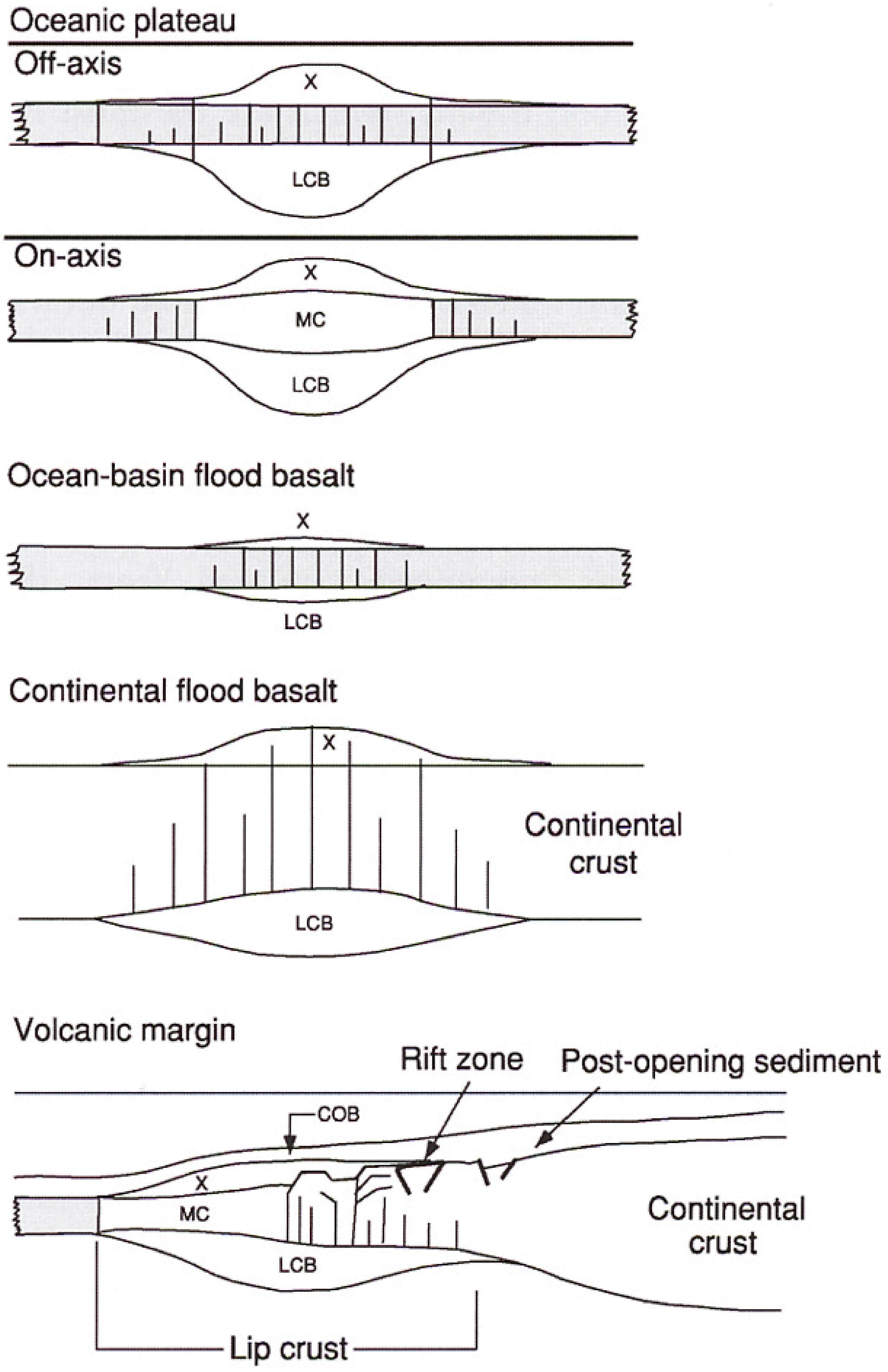

Figure 2 Schematic LIP plate-tectonic settings and gross crustal structure. The LIP crustal components are extrusive cover (X), middle crust (MC), and lower crustal body (LCB). Normal oceanic crust is shown in grey, and intrusives are denoted by vertical lines. The continent-ocean boundary (COB) is indicated for the volcanic margin.

Low-velocity zones have recently been observed in the mantle beneath the oceanic Ontong Java Plateau and the continental Deccan Traps and Paraná flood basalts (**Figure 1** and **Table 1**). Interpreted as lithospheric roots or keels, these zones can extend at least 500–600 km into the mantle. In contrast to the high-velocity roots beneath most continental areas and the absence of lithospheric keels in most oceanic areas, the low-velocity zones beneath LIPs apparently reflect primarily residual chemical, and perhaps some thermal, effects of mantle-plume activity. High-buoyancy roots extending well into the mantle beneath oceanic LIPs would suggest a significant role for LIPs in continental growth via accretion of oceanic LIPs to the edges of continents.

Distribution, Tectonic Setting, and Types

LIPs occur worldwide, in both continental and oceanic crust, in purely intraplate settings and along present and former plate boundaries (**Figure 1** and **Table 1**), although the tectonic setting at the time of formation is unknown for many features. If a LIP forms at a plate boundary, the entire crustal section is LIP crust (**Figure 2**). Conversely, if one forms in an intraplate setting, the pre-existing crust must be intruded and sandwiched by LIP magmas, albeit to an extent that is not resolvable by current geological or geophysical techniques.

Continental flood basalts, which are the most intensively studied LIPs owing to their exposure, are erupted from fissures in continental crust (**Figure 1** and **Table 1**). Most continental flood basalts overlie sedimentary basins that formed via extension, but it is not clear what happened first, the magmatism or the extension. Volcanic passive margins form as a result of excessive magmatism during continental breakup along the trailing rifted edges of continents. In the deep ocean basins, four types of LIPs are found. Oceanic plateaus, commonly isolated from the major continents, are broad typically flat-topped features generally lying 2000 m or more above the surrounding seafloor. They can form at triple junctions (e.g. the Shatsky Rise), at mid-ocean ridges (e.g. Iceland), or in intraplate settings (e.g. the northern Kerguelen Plateau). Submarine ridges are elongated steep-sided elevations of the seafloor. Some form along transform plate boundaries (e.g. the Ninetyeast Ridge). In the oceanic realm, oceanic plateaus and submarine ridges are the most enigmatic LIPs with respect to the tectonic setting in which they formed. Seamounts, which are closely related to submarine ridges, are local elevations of the seafloor; they may be discrete, form linear or random groups, or be connected along their bases and aligned along a ridge or rise (*see* **Seamounts**). They commonly form in intraplate regions (e.g. Hawaii). Ocean-basin flood basalts (e.g. the Nauru Basin and the Caribbean province) are the least-studied type of LIP and consist of extensive submarine flows and sills lying above and postdating the normal oceanic crust.

Ages

Age control for all LIPs apart from continental flood basalts is poor owing to their relative inaccessibility, but the ^{40}Ar/^{39}Ar dating technique is having a particularly strong impact on studies of LIP volcanism. Geochronological studies of continental flood basalts (e.g. the Siberian Traps, the Karoo, the Ferrar Basalts, the Deccan Traps, and the Columbia River Basalt; **Figure 1**) suggest that most LIPs result from mantle plumes, which initially transfer huge volumes (*ca.* 10^5–10^7 km^3) of mafic rock into localized regions of the crust over short intervals (*ca.* 10^5–10^6 years) but which subsequently transfer mass at a far lesser rate, albeit over significantly longer intervals (10^7–10^8 years). Transient magmatism during LIP formation is commonly attributed to mantle-plume 'heads' reaching the crust following transit through all or part of the Earth's mantle, whereas persistent magmatism is considered to result from steady-state mantle-plume 'tails' penetrating the lithosphere, which is moving relative to the plume (**Figure 3**). However, not all LIPs have obvious connections with mantle plumes or even hotspot tracks, suggesting that more than one source model may be required to explain all LIPs.

LIPs are not distributed uniformly over time. For example, many LIPs formed between 50 Ma and 150 Ma, whereas few have formed during the past 50 Myr (**Figure 4**). Such episodicity probably reflects variations in rates of mantle circulation, and this is supported by high rates of seafloor spreading during a portion of the 50–150 Ma interval, specifically during the long Cretaceous Normal Superchron

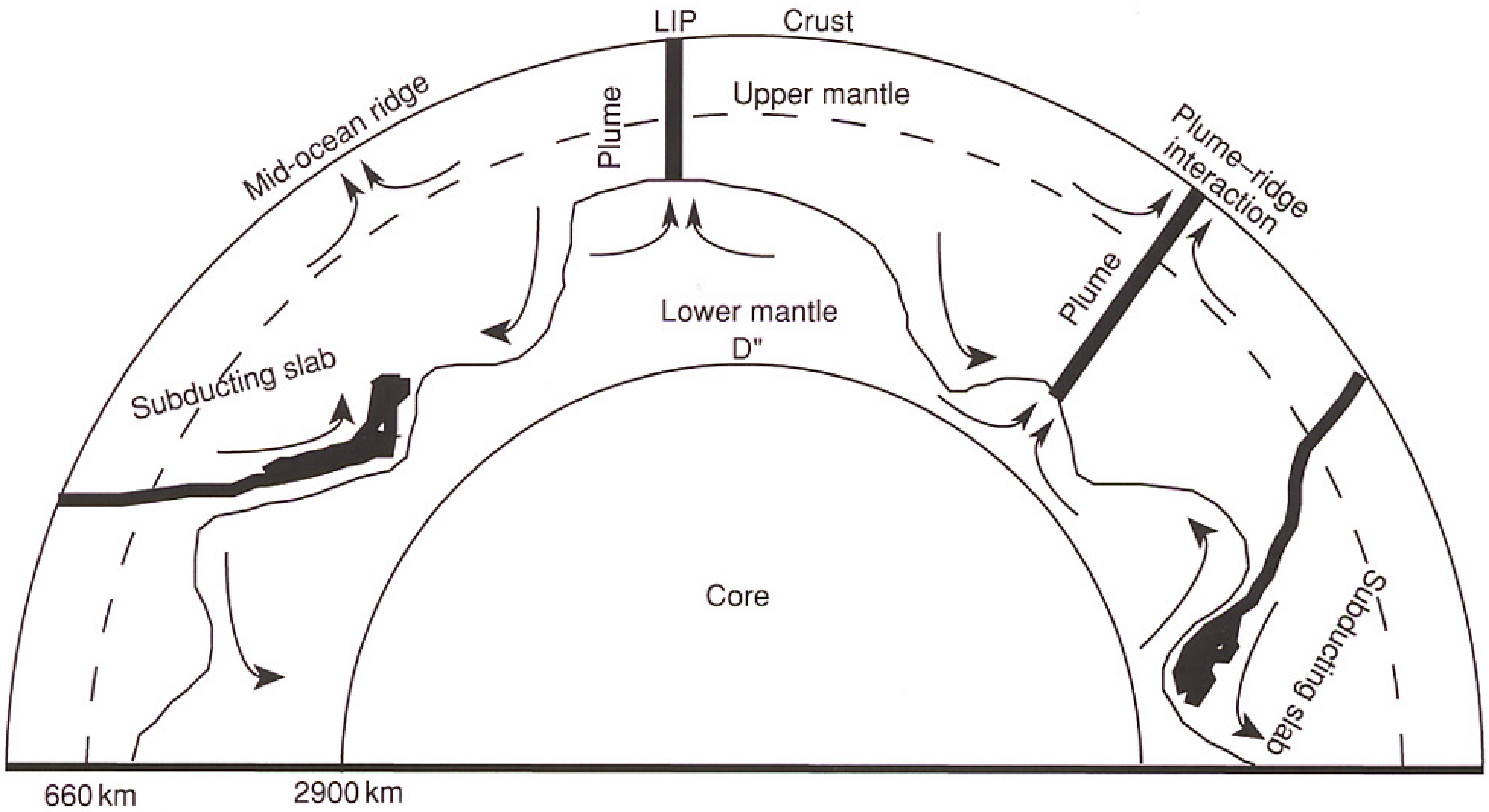

Figure 3 Diagram of the Earth's interior, showing plumes (tails), subducting slabs, and two mantle layers that move in complex patterns but never mix.

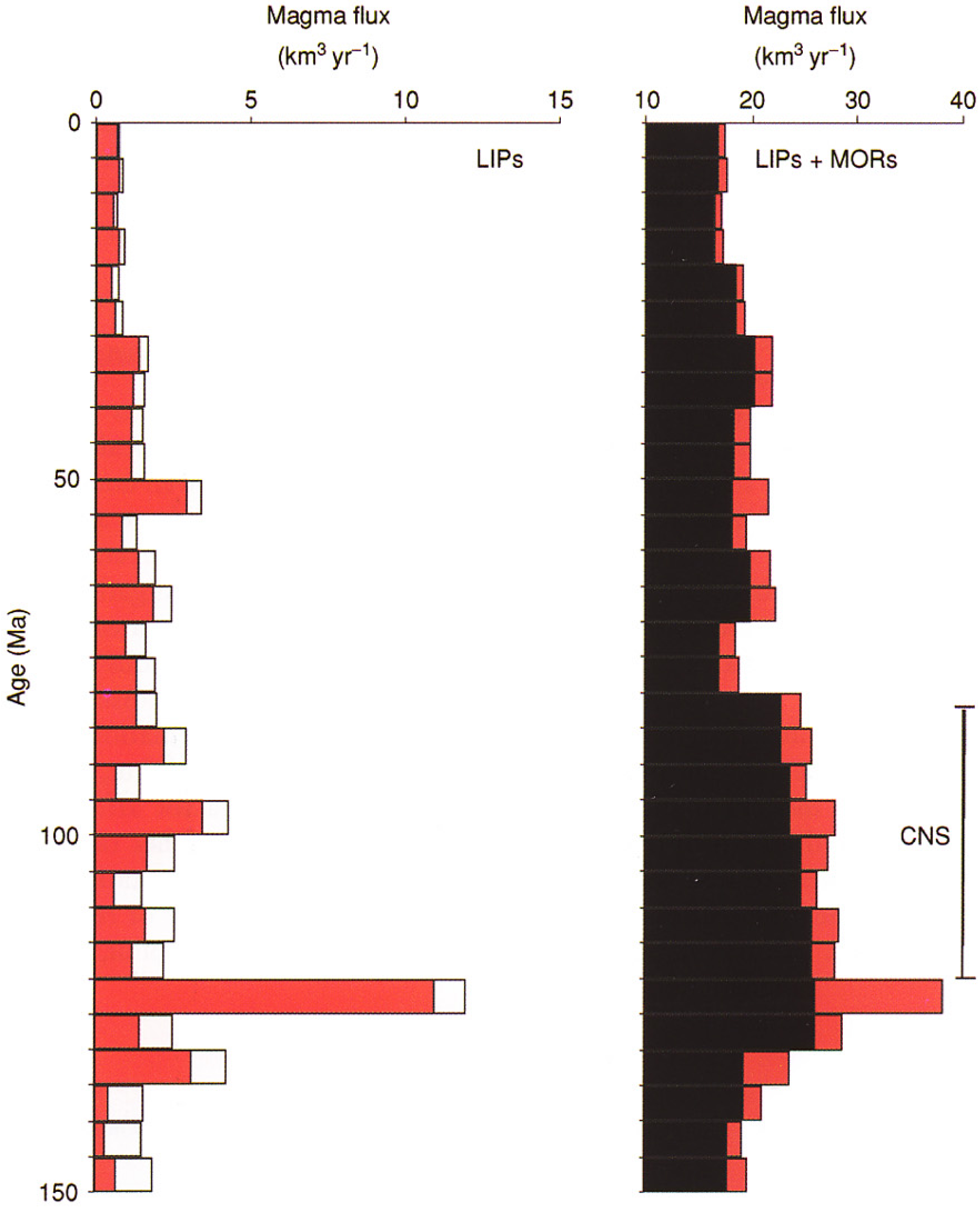

Figure 4 LIP, corrected for subduction (left; preserved in red and subduction correction in white), and summed LIP (red) and mid-ocean ridge (MOR; black) (right) magma production since 150 Ma. Note the difference in the *x*-axis scales. CNS, Cretaceous Normal Superchron.

(*ca.* 120–80 Ma), a time during which the Earth's magnetic field was of normal polarity. Thus, although LIPs manifest types of mantle processes distinct from those resulting in seafloor spreading, the waxing and waning rates of overall mantle circulation probably affect both sets of processes. A major question that emerges from studies of the global LIP production rate is whether the mantle is circulating less vigorously as the Earth ages.

LIPs and Mantle Dynamics

The formation of various sizes of LIP in a variety of tectonic settings on both continental and oceanic lithosphere suggests that a variety of thermal anomalies in the mantle give rise to LIPs and that the lithosphere strongly controls their formation. Equivalent thermal anomalies beneath continental and oceanic lithosphere should produce more magmatism in the latter setting, as oceanic lithosphere is thinner, allowing more decompression melting. Similarly, equivalent thermal anomalies beneath an intraplate region (e.g. Hawaii) and a divergent plate boundary (e.g. Iceland) (**Figure 1** and **Table 1**) will produce more magmatism in the latter setting, again because decompression melting is enhanced. Recent seismic tomographic images of mantle-velocity (a proxy for temperature) structure beneath Iceland and Hawaii show significant differences between the two.

Only recently, seismic tomography has revealed that slabs of subducting lithosphere can penetrate the entire mantle to the D″ layer just above the boundary between the mantle and the core, at a depth of approximately 2900 km (**Figure 3**). If we assume that the volume of the Earth's mantle has remained roughly constant throughout geological time, then the mass of crustal material fluxing into the mantle must be balanced by an equivalent mass of material fluxing from the mantle to the crust. Most, if not all, of the magmatism associated with the plate-tectonic processes of seafloor spreading and subduction is believed, on the basis of geochemistry and seismic tomography, to be derived from the upper mantle (above *ca.* 660 km depth). It is reasonable to assume that the lithospheric material that enters the lower mantle is eventually recycled, in some part contributing to the emplacement of LIPs at the Earth's surface.

Although LIPs are commonly believed to have originated from mantle plumes generated solely by solid Earth processes, alternative mechanisms have also been proposed. The spatial, if not temporal, association of flood basalts and impact craters on the Moon, as well as limited evidence on Earth, suggests that massive decompression melting of the mantle or at least significant crustal thinning and fracturing forming conduits for mantle material to reach the surface of a terrestrial planet could account for the emplacement of some LIPs. Such a mechanism has been proposed as an alternative to the plume hypothesis for the Siberian Traps, the Ontong Java Plateau, and the Deccan Traps. Other LIPs may originate as a result of a combination of plate divergence or fracturing and co-located underlying thermally anomalous mantle. Thus, multiple mechanisms may be required to explain all LIPs, both on Earth and elsewhere in our solar system.

LIPs and the Environment

The formation of LIPs has had documented environmental effects both locally and regionally. The global effects are less well understood, but the formation of some LIPs may have affected the global environment, particularly when conditions were at or near a threshold state. Investigations of volcanic passive margins and oceanic plateaus have demonstrated widespread and voluminous subaerial basaltic eruptions. The eruption of enormous volumes of basaltic magma during LIP formation releases volatiles such as carbon dioxide, sulphur, chlorine, and fluorine (**Figure 5**). A key factor affecting the magnitude of volatile release is whether the eruptions are subaerial or submarine; hydrostatic pressure inhibits vesiculation and degassing of relatively soluble volatile components (water, sulphur, chlorine, and fluorine)

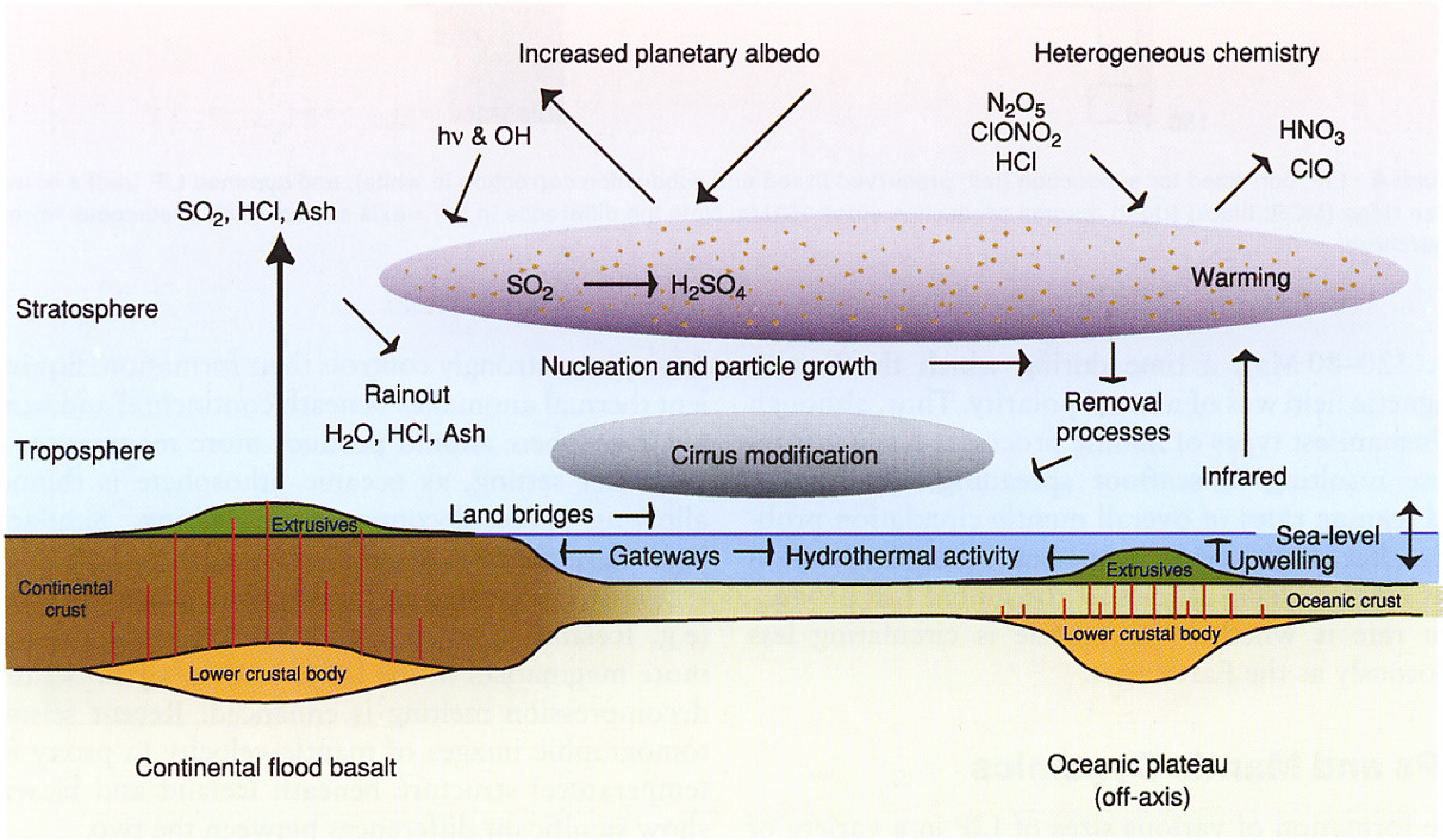

Figure 5 Environmental effects of LIP formation. LIP eruptions can perturb the Earth–ocean–atmosphere system significantly. Note that many oceanic plateaus form, at least in part, subaerially. Energy from solar radiation is hv, where h = Planck's constant, and v = frequency of electromagnetic wave of solar radiation.

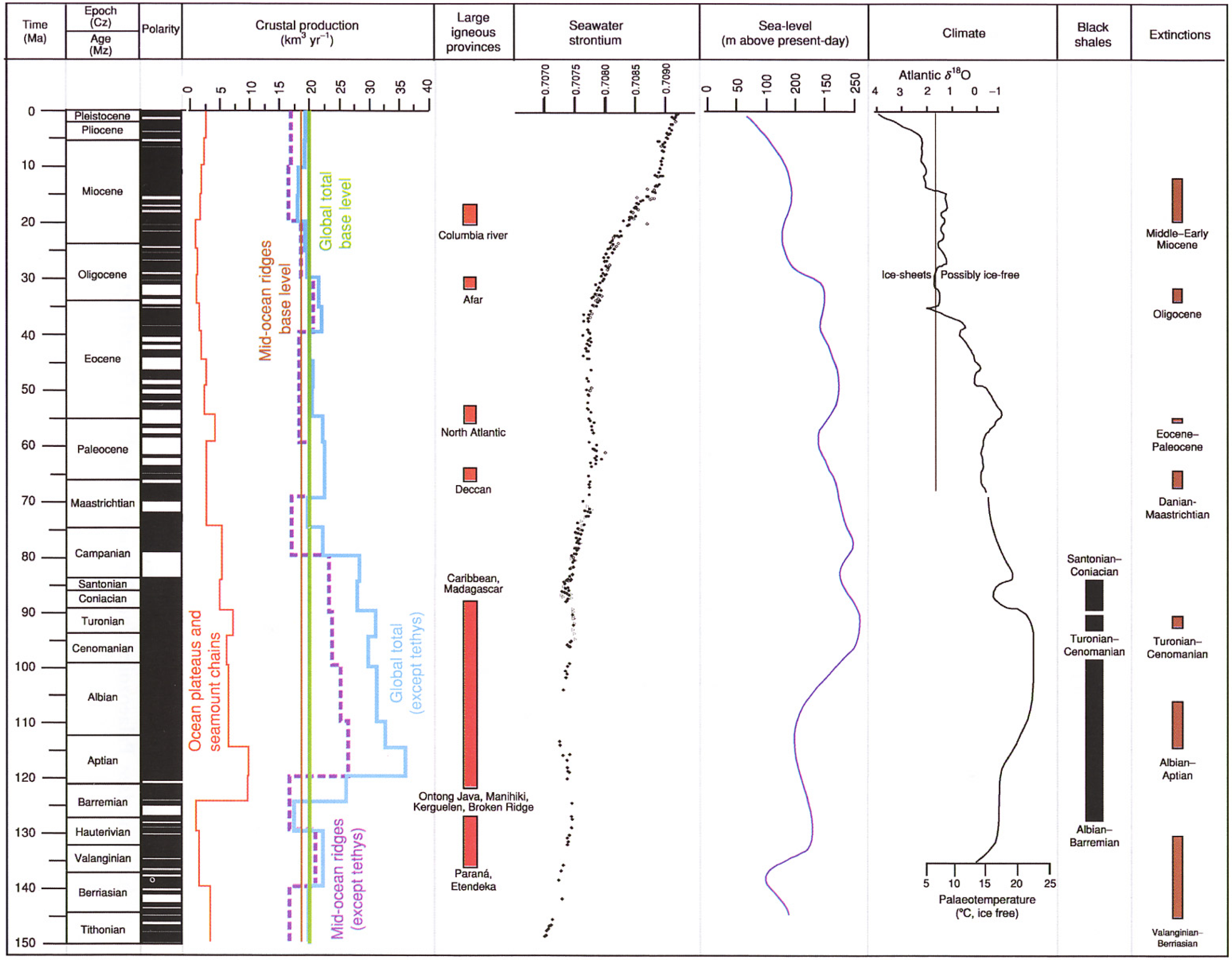

Figure 6 Temporal correlations among geomagnetic polarity, crustal production rates, LIPs, seawater strontium, sea-level, climate, black shales, and extinctions.

during deep-water submarine eruptions, although low-solubility components (carbon dioxide and noble gases) are mostly degassed even at abyssal depths.

Another important factor affecting the environmental impact of LIP volcanism is the latitude at which the LIP forms. In most basaltic eruptions, released volatiles remain in the troposphere. However, at high latitudes, the tropopause is relatively low, allowing large-mass-flux basaltic fissure eruption plumes to transport SO_2 and other volatiles into the stratosphere. Sulphuric acid aerosol particles that form in the stratosphere after such eruptions have a longer residence time and greater global dispersal than if the SO_2 remains in the troposphere; therefore, the effects on climate and atmospheric chemistry are greater. The large volume of volatiles released, over relatively brief geological intervals, by the subaerial flood basalts of high-latitude LIPs would contribute to potential global environmental effects.

Highly explosive felsic eruptions, such as those documented from volcanic passive margins, an oceanic plateau (Kerguelen; **Figure 1** and **Table 1**), and continental flood basalt provinces, can also inject both particulate material and volatiles (SO_2 and CO_2) directly into the stratosphere. The total volume of felsic volcanic rocks in LIPs is poorly constrained, but they may account for a small, but not negligible, fraction of the volcanic deposits in LIPs. Significant volumes of explosive felsic volcanism would further contribute to the effects of predominantly mafic LIP volcanism on the global environment.

Between about 145 Ma and 50 Ma, the global oceans were characterized by variations in chemistry, relatively high temperatures, high relative sea-level, episodic deposition of black shales, high production of hydrocarbons, mass extinctions of marine organisms, and radiations of marine flora and fauna (**Figure 6**). Temporal correlations between the intense pulses of igneous activity associated with LIP formation and environmental changes suggest a causal relationship. Perhaps the most dramatic example is the eruption of the Siberian Traps (**Figure 1** and **Table 1**) at approximately 250 Ma, coinciding with the largest extinction of plants and animals in the geological record. It is estimated that 90% of all species became extinct at that time (*see* **Palaeozoic:** End Permian Extinctions). On Iceland, the 1783–1784 eruption of Laki provides the only human experience of the type of volcanism that constructs LIPs. Although Laki produced a basaltic lava flow representing approximately 1% of the volume of a typical (10^3 km^3) LIP flow, the eruption's environmental impact resulted in the deaths of 75% of Iceland's livestock and 25% of its population from starvation.

Conclusions

Oceanic plateaus, volcanic passive margins, submarine ridges, seamount chains, ocean-basin flood basalts, and continental flood basalts share geological and geophysical characteristics that indicate an origin distinct from that of igneous rocks formed at mid-ocean ridges and arcs. These characteristics include

- a broad areal extent (in excess of 10^4 km^2) of iron- and magnesium-rich lavas;
- massive transient basaltic volcanism occurring over 10^5–10^6 years;
- persistent basaltic volcanism from the same source lasting 10^7–10^8 years;
- lower crustal bodies characterized by P-wave velocities of 7.0–7.6 km s^{-1};
- a component of more silica-rich volcanic rocks;
- higher MgO lavas, basalts with more diverse major-element compositions, rocks with more common fractionated components, both alkalic and tholeiitic differentiates, and basalts with predominantly flat light-rare-earth-element patterns, relative to mid-ocean-ridge basalts;
- thick (tens to hundreds of metres) individual basalt flows;
- long (up to 750 km) single basalt flows; and
- lavas erupted in both subaerial and submarine settings.

There is strong evidence that many LIPs both manifest a fundamental mode of mantle circulation, commonly distinct from that which characterizes plate tectonics, and contribute episodically, at times catastrophically, to global environmental change. Nevertheless, it is important to bear in mind that we have literally only scratched the surface of oceanic and continental LIPs, and that LIPs on other terrestrial planets await investigation.

See Also

Earth: Mantle. **Igneous Processes**. **Lava**. **Mantle Plumes and Hot Spots**. **Palaeozoic:** End Permian Extinctions. **Plate Tectonics**. **Seamounts**. **Tectonics:** Mid-Ocean Ridges.

Further Reading

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