LARGE IGNEOUS PROVINCES: CRUSTAL STRUCTURE, DIMENSIONS, AND EXTERNAL CONSEQUENCES

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Abstract. Large igneous provinces (LIPs) are a continuum of voluminous iron and magnesium rich rock emplacements which include continental flood basalts and associated intrusive rocks, volcanic passive margins, oceanic plateaus, submarine ridges, seamount groups, and ocean basin flood basalts. Such provinces do not originate at "normal" seafloor spreading centers. We compile all known in situ LIPs younger than 250 Ma and analyze dimensions, crustal structures, ages, and emplacement rates of representatives of the three major LIP categories: Ontong Java and Kerguelen-Broken Ridge oceanic plateaus, North Atlantic volcanic passive margins, and Deccan and Columbia River continental flood basalts. Crustal thicknesses range from 20 to 40 km, and the lower crust is characterized by high (7.0–7.6 km s^{-1}) compressional wave velocities. Volumes and emplacement rates derived for the two giant oceanic plateaus, Ontong Java and Kerguelen, reveal short-lived pulses of increased global production; Ontong Java's rate of emplacement may have exceeded the contemporaneous global production rate of the entire mid-ocean ridge system. The major part of the North Atlantic volcanic province lies offshore and demonstrates that volcanic passive margins belong in the global LIP inventory. Deep crustal intrusive companions to continental flood volcanism represent volumetrically significant contributions to the crust. We envision a complex mantle circulation which must account for a variety of LIP sizes, the largest originating in the lower mantle and smaller ones developing in the upper mantle. This circulation coexists with convection associated with plate tectonics, a complicated thermal structure, and at least four distinct geochemical/isotopic reservoirs. LIPs episodically alter ocean basin, continental margin, and continental geometries and affect the chemistry and physics of the oceans and atmosphere with enormous potential environmental impact. Despite the importance of LIPs in studies of mantle dynamics and global environment, scarce age and deep crustal data necessitate intensified efforts in seismic imaging and scientific drilling in a range of such features.

1. INTRODUCTION

Large igneous provinces (LIPs) are massive crustal emplacements of predominantly mafic (Mg and Fe rich) extrusive and intrusive rock which originate via processes other than "normal" seafloor spreading. As physical manifestations of mantle processes, these global phenomena include continental flood basalts, volcanic passive margins, oceanic plateaus, submarine ridges, seamount groups, and ocean basin flood basalts (Figure 1 and Tables 1 and 2). Study of LIP-forming processes has intensified over the past five years as new information has become available from seismic tomography, geochemistry, marine geophysics, petrology, and geodynamic modeling, among other geoscientific disciplines [e.g., *Coffin and Eldholm*, 1991, 1992, 1993]. Furthermore, recent research has documented important temporal, spatial, and compositional similarities among various LIPs. These new analyses have produced fascinating, speculative interpretations, headlined by bold terms such as superswell

[*McNutt and Fischer*, 1987], megaswell [*Vogt*, 1991], superplume [*Larson*, 1991*a*, *b*], rolling thunder [*Pringle*, 1991], and supergreenhouse [*Arthur et al.*, 1991].

LIPs are globally significant. After basalt and associated intrusive rock emplaced at spreading centers, LIPs are the most significant accumulations of mafic material on the Earth's surface. They are commonly attributed to mantle plumes or hotspots [*Wilson*, 1963; *Morgan*, 1971, 1981] and presently account for between 5% and 10% of the heat and magma expelled from the mantle [*Davies*, 1988; *Sleep*, 1990]. However, these fluxes are not distributed evenly in space and time; their episodicity punctuates the relatively steady state production of crust at seafloor spreading centers (Figure 2) [*Larson*, 1991*b*]. LIPs therefore provide windows into those regions of the mantle which do not generate normal mid-ocean ridge basalt. Seismic tomography suggests that velocity structure beneath mid-ocean ridges is different than that beneath hotspots [e.g., *Dziewonski and Woodhouse*, 1987], implying shallower, passive upwelling beneath ridges and
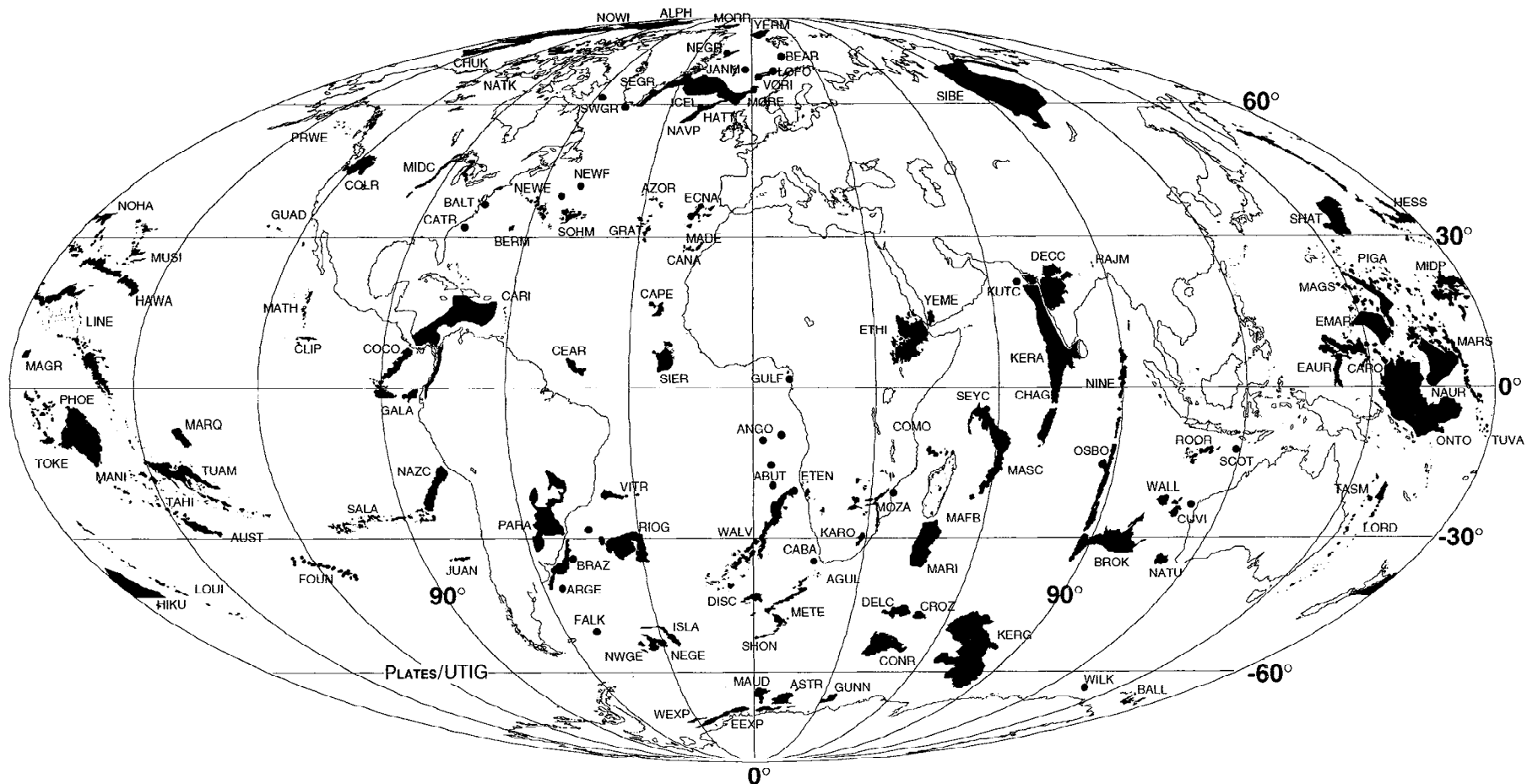
Figure 1. Global large igneous provinces (LIPs), including continental flood basalt and associated intrusive provinces, volcanic passive margins, oceanic plateaus, submarine ridges, ocean basin flood basalts, and seamount groups (see Table 1 for key to abbreviations). Solid circles indicate volcanic passive margins where seaward dipping reflector sequences have been recognized (see Table 2 for key to abbreviations). Modified from the work by *Coffin and Eldholm* [1991] with additional data from K. Hinz (personal communication, 1992). Digital map courtesy of the Plates project (Institute for Geophysics, University of Texas, Austin).

TABLE 1. Large Igneous Provinces, Excluding Volcanic Passive Margins

LIP	Abbreviations in Figure 1	Type	Reference
Agulhas Ridge	AGUL	SR	IHO/IOC/CHS [1984]
Alpha-Mendeleyev Ridge	ALPH	SR/OP*	Weber and Sweeney [1990]
Angola Plain	ANGO	ASSC	Hinz and Block [1991]
Austral seamounts	AUST	SMT	Crough [1978]
Azores	AZOR	SMT	G. R. Davies et al. [1989]
Balleny Islands	BALL	SMT	Lanyon et al. [1993]
Bermuda Rise	BERM	OP	Detrick et al. [1986]
Broken Ridge	BROK	SR	MacKenzie [1984]
Canary Islands	CANA	SMT	Schmincke [1982]
Cape Verde Rise	CAPE	OP	Courtney and White [1986]
Caribbean flood basalt	CARI	CFB/OBFB	Bowland and Rosencrantz [1988]
Caroline seamounts	CARO	SMT	Mattey [1982]
Ceara Rise	CEAR	OP	Supko and Perch-Nielsen [1977]
Chagos-Laccadive Ridge	CHAG	SR	Duncan [1990]
Chukchi Plateau	CHUK	OP	Hall [1990]
Clipperton seamounts	CLIP	SMT	Mammerickx [1989]
Cocos Ridge	COCO	SR	Batiza [1989]
Columbia River Basalt	COLR	CFB	Reidel and Hooper [1989]
Comores Archipelago	COMO	SMT	Emerick and Duncan [1982]
Conrad Rise	CONR	OP	Diament and Goslin [1986]
Crozet Plateau	CROZ	OP	Goslin and Diament [1987]
Cuvier Plateau	CUVI	OP	Larson et al. [1979]
Deccan traps	DECC	CFB	Mahoney [1988]
Del Caño Rise	DELC	OP	Goslin and Diament [1987]
Discovery seamounts	DISC	SMT	IHO/IOC/CHS [1984]
Eauripik Rise	EAUR	OP	Den et al. [1971]
Eastern central North Atlantic	ECNA	ASSC	K. Hinz (personal communication, 1991)
East Mariana Basin	EMAR	OBFB	Abrams et al. [1993]
Emeishan basalts†		CFB	Chung and Jahn [1993]
Etendeka	ETEN	CFB	Cox [1988]
Ethiopian flood basalt	ETHI	CFB	Mohr and Zanettin [1988]
Foundation seamounts	FOUN	SMT	Mammerickx [1992]
Galapagos/Carnegie Ridge	GALA	SMT/SR	Christie et al. [1992]
Great Meteor-Atlantis seamounts	GRAT	SMT	Tucholke and Smoot [1990]
Guadelupe seamount chain	GUAD	SMT	Batiza [1989]
Hawaiian-Emperor seamounts	HAWA	SMT	Clague and Dalrymple [1989]
Hess Rise	HESS	OP	Vallier et al. [1983]
Hikurangi Plateau	HIKU	OP	Wood and Davey [1994]
Iceland/Greenland-Scotland Ridge	ICEL	OP/SR	Vogt [1974]
Islas Orcadas Rise	ISLA	SR	LaBrecque and Hayes [1979]
Juan Fernandez Archipelago	JUAN	SMT	IHO/IOC/CHS [1984]
Karoo	KARO	CFB	Cox [1988]
Kerguelen Plateau	KERG	OP	Houtz et al. [1977]
Line Islands	LINE	SMT	Sandwell and Renkin [1988]
Lord Howe Rise seamounts	LORD	SMT	Wellman and McDougall [1974]
Louisville Ridge	LOUI	SMT	Lonsdale [1988]
Madagascar flood basalts	MAFB	CFB	Mahoney et al. [1991]
Madagascar Ridge	MARI	SR	Sinha et al. [1981]
Madeira Rise	MADE	OP	Peirce and Barton [1991]
Magellan Rise	MAGR	OP	Winterer et al. [1973]
Magellan seamounts	MAGS	SMT	IHO/IOC/CHS [1984]
Manihiki Plateau	MANI	OP	Winterer et al. [1974]
Marquesas Islands	MARQ	SMT	Fischer et al. [1987]
Marshall Gilbert seamounts	MARS	SMT	Schlanger et al. [1981]
Mascarene Plateau	MASC	OP	Duncan [1990]
Mathematicians seamounts	MATH	SMT	Mammerickx [1989]
Maud Rise	MAUD	OP	Barker et al. [1988]
Meteor Rise	METE	SR	LaBrecque and Hayes [1979]
Midcontinent rift volcanics (Keweenawan)	MIDC	CFB	Van Schmus [1992]
Mid-Pacific Mountains	MIDP	SMT	Winterer and Metzler [1984]
Musicians seamounts	MUSI	SMT	Batiza [1989]
Naturaliste Plateau	NATU	OP	Coleman et al. [1982]
Natkusiak flood basalt	NATK	CFB	Rainbird [1993]
Nauru Basin	NAUR	OBFB	Shipley et al [1993]

TABLE 1. (continued)

LIP	Abbreviations in Figure 1	Type	Reference
Nazca Ridge	NAZC	SR	*Pilger and Handschumacher* [1981]
New England seamounts	NEWE	SMT	*Duncan* [1984]
Ninetyeast Ridge	NINE	SR	*Peirce et al.* [1989]
North Atlantic volcanic province	NAVP	CFB	*Upton* [1988]; *Dickin* [1988]
Northeast Georgia Rise	NEGE	OP	*LaBrecque and Hayes* [1979]
Northwest Georgia Rise	NWGE	OP	*Brenner and LaBrecque* [1988]
Northwest Hawaiian Ridge	NOHA	SR/SMT	*Mammerickx* [1989]
Northwind Ridge	NOWI	SR	*Hall* [1990]
Ontong Java Plateau	ONTO	OP	*Hussong et al.* [1979]
Osborn Knoll	OSBO	OP	*Sclater and Fisher* [1974]
Paraná flood volcanism	PARA	CFB	*Piccirillo et al.* [1988]
Phoenix seamounts	PHOE	SMT	*IHO/IOC/CHS* [1984]
Pigafetta Basin	PIGA	OBFB	*Abrams et al.* [1993]
Pratt-Welker seamounts	PRWE	SMT	*Batiza* [1989]
Rajmahal traps	RAJM	CFB	*Mahoney et al.* [1983]
Rio Grande Rise	RIOG	OP	*Gamboa and Rabinowitz* [1984]
Roo Rise	ROOR	OP	*IHO/IOC/CHS* [1984]
Sala y Gomez Ridge	SALA	SR	*Pilger and Handschumacher* [1981]
Shatsky Rise	SHAT	OP	*Den et al.* [1969]
Shona Ridge	SHON	SR	*IHO/IOC/CHS* [1984]
Siberian traps	SIBE	CFB	*Zolotukhin and Al'Mukhamedov* [1988]
Sierra Leone Rise	SIER	OP	*Kumar* [1979]
Sohm Abyssal Plain	SOHM	ASSC	*Hinz and Popovici* [1989]
Tahiti	TAHI	SMT	*Duncan and McDougall* [1976]
Tasmantid seamounts	TASM	SMT	*McDougall and Duncan* [1988]
Tokelau seamounts	TOKE	SMT	*IHO/IOC/CHS* [1984]
Tuamotu Archipelago	TUAM	SMT	*Duncan and Clague* [1985]
Tuvalu seamounts	TUVA	SMT	*IHO/IOC/CHS* [1984]
Vitória-Trindade Ridge	VITR	SR/SMT	*Francisconi and Kowsmann* [1976]
Wallaby Plateau	WALL	OP	*Symonds and Cameron* [1977]
Walvis Ridge	WALV	SR	*Rabinowitz and LaBrecque* [1979]
Wrangellia†		OP	*Richards et al.* [1991]
Yemen Plateau basalts	YEME	CFB	*Mohr and Zanettin* [1988]

ASSC, anomalous seafloor spreading crust; CFB, continental flood basalt; OBFB, ocean basin flood basalt; OP, oceanic plateau; SMT, seamount; SR submarine ridge.

*Referred to in the literature as both a submarine ridge and an oceanic plateau.

†Not mapped in Figure 1.

deeper, dynamic upwelling beneath hotspots [*Zhang and Tanimoto*, 1992]. Furthermore, thermal structure of the upper mantle is heterogeneous, suggesting that extensive magmatism requires a combination of hot upper mantle and suitable lithospheric conditions [*Anderson et al.*, 1992a, b]. Despite a wealth of data and results on hotspots and mantle plumes (for reviews and syntheses, see, for example, *White and McKenzie* [1989], *Duncan and Richards* [1991], *Hill et al.* [1992], and *Sleep* [1992]), not one of the available models satisfactorily explains the genesis of all LIPs or even individual categories of LIPs.

In this study we will describe and characterize transient LIPs, that is, those LIPs which form by voluminous, but probably relatively short-lived, magmatic episodes and review their crustal structure. Because dimensions and emplacement rates for most LIPs are poorly known but are of fundamental importance both for internal processes and external implications, we calculate LIP dimensions and rates in order to evaluate and summarize their genesis, evolution, and pos-

sible environmental consequences. To do so, we use primarily geophysical and geological data and published results, focusing on five major mafic emplacements: the Ontong Java and Kerguelen/Broken Ridge oceanic plateaus, the North Atlantic volcanic passive margins, and the Deccan and Columbia River continental flood basalts, which represent the three main categories of transient LIPs. Of these, the two oceanic plateaus are the largest igneous provinces and may represent the highest rates of planetary crustal production on timescales of 10^6 years.

2. COMPOSITION, CATEGORIES, AND TECTONIC SETTING

LIPs are dominated by mafic rocks. Because continental flood basalts are far more accessible to sampling than submarine volcanic passive margin and oceanic plateau igneous rocks, we know much more about the petrology and geochemistry of the former. The

**TABLE 2. Volcanic Passive Margins, Including Marginal Plateaus**

Continent (Subcontinent/Microcontinent)	Location	Abbreviations in Figure 1	Reference
Africa	Abutment Plateau	ABUT	*Hinz* [1981]
	Cape Basin	CABA	*Hinz* [1981]
	Gulf of Guinea	GULF	*Rosendahl et al.* [1991]
	Mozambique Basin	MOZA	*DeBuyl and Flores* [1986]
Antarctica	Astrid Ridge	ASTR	*Roeser et al.* [1990]
	Eastern Explora wedge	EEXP	*Hinz and Krause* [1982]
	Western Explora wedge	WEXP	*Kristoffersen and Hinz* [1991]
	Gunnerus Ridge	GUNN	*Roeser et al.* [1990]
	Wilkes Land	WILK	*Eittrem et al.* [1985]
Australia	Cuvier Plateau	CUVI	*Hopper et al.* [1992]
	Scott Plateau	SCOT	*Hinz* [1981]
Greenland	Jan Mayen Ridge	JANM	*Skogseid and Eldholm* [1987]
	Morris Jesup Rise	MORR	*Feden et al.* [1979]
	Northeast Greenland	NEGR	*Hinz et al.* [1987]
	Southeast Greenland	SEGR	*Larsen and Jakobsdottir* [1988]
	Southwest Greenland	SWGR	*Chalmers* [1991]
India	Kerala Basin	KERA	*Hinz* [1981]
	Kutch Basin	KUTC	*Hinz* [1981]
North America	Baltimore Canyon trough	BALT	*Sheridan et al.* [1993]
	Carolina trough	CATR	*Austin et al.* [1990]
	Newfoundland Ridge	NEWF	*Grant* [1977]
Northwest Europe	Bear Island margin	BEAR	*Faleide et al.* [1988]
	Hatton Bank	HATT	*Roberts et al.* [1984]
	Lofoten margin	LOFO	*Eldholm et al.* [1979]
	Møre margin	MORE	*Smythe* [1983]
	Vøring margin	VORI	*Hinz and Weber* [1976]
	Yermak Plateau	YERM	*Feden et al.* [1979]
Seychelles	Seychelles Bank	SEYC	*Devey and Stephens* [1992]
South America	Argentine margin	ARGE	*Hinz* [1990]
	Brazilian margin	BRAZ	*Chang et al.* [1992]
	Falkland Plateau	FALK	*Lorenzo and Mutter* [1988]

most abundant rock types are typically subhorizontal, subaerial basalt (mostly tholeiitic) flows, some of which extend for many hundred kilometers and have volumes as great as several thousand cubic kilometers [e.g., *Tolan et al.*, 1989]. Felsic and intermediate rocks also occur as lavas and intrusives and are mostly associated with initial and late emplacement stages, while flood basalts largely uncontaminated by crustal components represent the most intense, transient eruption period. Relative to mid-ocean ridge basalt (MORB), major element compositions of continental flood basalts are much more diverse [*Macdougall*, 1988a]. Fractionated components are more common, and both alkalic and tholeiitic differentiates occur.

Petrologic and geochemical studies of rocks from LIP environments suggest lower to upper mantle sources for the magma. A consensus has developed on the basis of isotopic taxonomy that at least four different source reservoirs persist in the mantle: depleted MORB mantle (DMM), high U/Pb ratio (HIMU) and enriched mantle 1 and 2 (EM1 and EM2) [*Zindler and Hart*, 1986; *Hart et al.*, 1992]. In chemical and isotopic character, some continental flood basalts mimic intraoceanic hotspot basalts along the same plume trace [*Carlson*, 1991]. Other continental flood basalts resem-

ble MORB lavas derived from depleted mantle. Many continental flood basalts have chemistries characteristic of lithospheric fractionation and isotopic signatures more closely resembling those of continental crust than of convecting mantle. Differences among LIP chemistry and isotopes may be explained by varying degrees of crustal or lithospheric-mantle contamination and different subcontinental lithospheric and asthenospheric source compositions.

Intermediate and mafic sills are well developed in continental flood basalt provinces that overlie sedimentary sequences and along the offshore Norwegian and British volcanic margins and may contribute a significant fraction of the total igneous volume. Off Norway, flood basalts of a dipping reflector wedge are locally underlain by dacitic-andesitic lavas interpreted as having been emplaced during the late rift stage [*Eldholm et al.*, 1989] by partial melting of a component of continental material in a shallow crustal magma chamber [*Parson et al.*, 1989]. Dacite also underlies basalt in the northern Rockall Trough off Great Britain [*Morton et al.*, 1988]. Some provinces also contain significant amounts of rhyolite, probably erupted as high-temperature (~1100°C) lavas. Whether dacite and/or rhyolite is present in purely
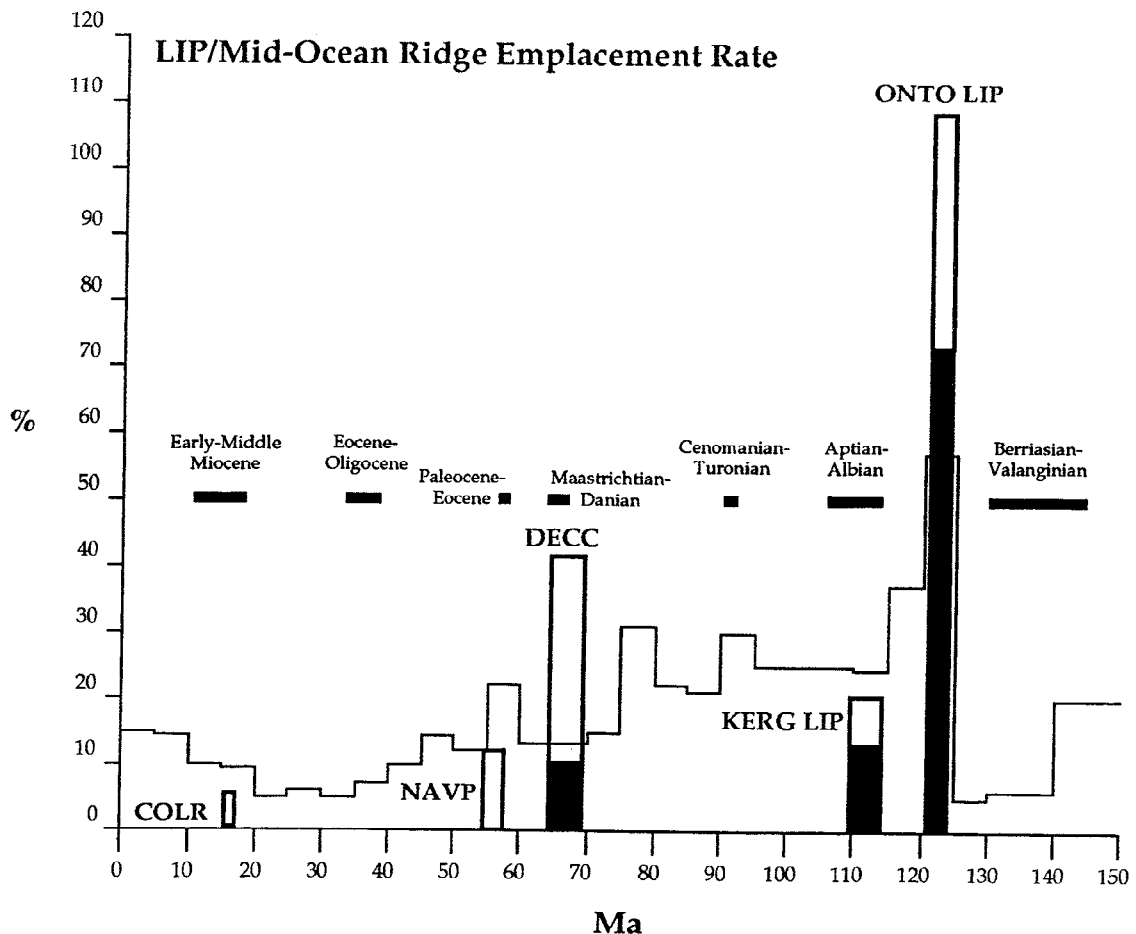
Figure 2. LIPs as a percentage of mid-ocean ridge crustal production versus time. Thin line indicates curve derived from the work by *Larson* [1991b]. Thick line depicts results of *Coffin and Eldholm* [1993]. The solid oceanic plateau (Kerguelen large igneous province (KERG LIP) and Ontong Java (ONTO) LIP) columns indicate off-ridge emplacement; solid and open parts depict on-ridge emplacement. Solid part of continental flood basalt DECC and COLR) columns indicates minimum rate; solid and open part indicates maximum rates. Maximum and minimum rates were calculated using 1.5- and 11.5-m.y. emplacement periods for Columbia River basalts (COLR) and 1.0 and 4.0 m.y. for Deccan traps (DECC). Compare with Figure 6. Solid horizontal bars indicate Cenozoic and Cretaceous mass extinctions [after *Rampino and Stothers*, 1988; *Thomas*, 1992]; note the correlation of COLR, DECC, and Kerguelen Plateau (KERG) LIP with extinctions. ONTO indicates Ontong Java Plateau only; ONTO LIP includes Ontong Java and Manihiki plateaus, and Mariana, Nauru, and Pigafetta basin flood basalts. KERG indicates Kerguelen Plateau only; KERG LIP includes Kerguelen Plateau, Elan Bank, and Broken Ridge.

mafic oceanic plateaus is uncertain. No intrusive rocks have been recovered from LIPs; these deeper rocks, presumably gabbros and ultramafics, are probably at least as voluminous as the extrusive components.

LIPs are distributed worldwide, occurring on both continental and oceanic crust in purely intraplate settings, on present and former plate boundaries, and along the edges of continents [*Coffin and Eldholm*, 1992] (Figure 1 and Tables 1 and 2). Continental flood basalts, the most intensively studied LIPs, are predominantly tholeiitic magmas erupted on continental crust over timescales of 10^5–10^6 years. Generally believed to be the product of fissure eruptions, they

consist primarily of horizontal and subhorizontal flows. Volcanic passive margins, situated on the trailing, rifted edges of continents, are characterized by excessive volcanism and by uplift and/or lack of rapid initial subsidence during continental breakup [e.g., *Roberts et al.*, 1984; *Eldholm*, 1991]. Margin formation is associated with both intrusive and extrusive activity [*Larsen and Marcussen*, 1992; *Skogseid et al.*, 1992a]. The oldest oceanic crust is thicker than adjacent "normal" oceanic crust, and the lower crust beneath the extrusives is commonly characterized by bodies with high seismic velocities [e.g., *Mutter and Zehnder*, 1988; *White and McKenzie*, 1989; *Eldholm and Grue*, 1993].

In deep ocean basins, the largest LIPs are oceanic plateaus: broad, more or less flat-topped features which generally lie 2000 m or more above the surrounding seafloor. They are formed by mafic volcanism and associated intrusive activity; their extrusive cover may be emplaced either in subaerial (e.g., Kerguelen Plateau) [Coffin, 1992] or submarine (e.g., Ontong Java Plateau) [Tarduno, 1992] environments. Oceanic plateaus are commonly isolated from major continents. Their crustal thickness is anomalously greater than that of adjacent oceanic crust, and their age may or may not be similar to that of adjacent crust. Another category of oceanic LIP is submarine ridges, which are elongated, steep-sided elevations of the seafloor. They may be of continental or oceanic origin, and we only consider the latter herein. Submarine ridges are commonly characterized by varying topography, and those of basaltic nature may be created either on or off the axes of spreading centers. As with oceanic plateaus, their volcanic upper crust may be emplaced either subaerially or under water [Coffin, 1992]. Closely related to submarine ridges are seamounts, which are local elevations of the seafloor. Seamounts are flat-topped (guyot) or peaked mafic volcanoes whose last stages of construction were in a subaerial or a submarine environment, respectively. They may be discrete, arranged in a linear or random grouping, or connected at their bases and aligned along a ridge or rise. The least studied LIPs are ocean basin flood basalts, which are thick, extensive submarine flows and sills lying above and postdating oceanic igneous basement. Recently, K. Hinz (personal communication, 1992) has discovered thickened oceanic crust and dipping intrabasement reflectors in deep ocean basins with linear seafloor spreading-type magnetic anomalies. We term this crustal type anomalous seafloor spreading crust.

Surface manifestations of the LIP types above record important spatial and temporal characteristics related to magma production. Many LIPs result from long-lived magmatic sources in the mantle, sources which initially transfer huge volumes of mafic rock into the crust over short intervals (~1 m.y.) but which later transfer material at a far lesser rate, albeit over long intervals (10–100 m.y.).

Transient magmatism during LIP formation is most clearly documented for continental flood basalts and volcanic passive margins (Figure 1 and Tables 1 and 2). Numerous ^{40}Ar-^{39}Ar dates, for example, show that most basalts of the Deccan traps [e.g., Duncan and Richards, 1991] and Siberian traps [e.g., Campbell et al., 1992] were erupted during ~1-m.y. periods indistinguishable from the Cretaceous-Tertiary and Permian-Triassic boundaries, respectively. Similarly, the North Atlantic volcanic province and other volcanic margins were emplaced during and immediately subsequent to continental breakup, but little or no volcanism, aside from that associated with seafloor spread-

ing, persisted thereafter. Such transient events, which also include oceanic plateaus, have been commonly attributed to plume "heads" manifesting themselves in the crust after arriving from the lower mantle [e.g., Richards et al., 1989].

Persistent magmatism resulting in LIP formation is in most cases related to lithosphere moving over more or less steady state mantle plumes or plume "tails" in the terminology of Richards et al. [1989]. Through Early Cretaceous and younger plate reconstructions [e.g., Duncan and Richards, 1991], each of several mantle plume sources, or hotspots, can be tied to a group of LIPs, including (Figure 1) (1) the Iceland source (North Atlantic volcanic province including North Atlantic volcanic margins and the Greenland-Scotland ridge), (2) the Kerguelen source (Bunbury basalt (Australia), Naturaliste Plateau, Rajmahal traps, Kerguelen Plateau/Broken Ridge, Ninetyeast Ridge, and northernmost Kerguelen Plateau), (3) the Réunion source (Deccan traps, western Indian volcanic margins, Chagos-Laccadive Ridge, Mascarene Plateau, and Mauritius), and (4) the Tristan da Cunha source (Paraná and Etendeka basalts, South Atlantic volcanic margins, Rio Grande Rise, Walvis Ridge). The Iceland source has been active for at least 60 m.y. [e.g., Upton, 1988], the Kerguelen source for 135 m.y. [e.g., H. L. Davies et al., 1989], the Réunion source for 65 m.y. [e.g., Courtillot et al., 1988], and the Tristan da Cunha source for 120 m.y. [e.g., O'Connor and le Roex, 1992]. Each of these hotspots was at some point during its evolution temporally and spatially associated with continental breakup.

Not all LIPs, however, have obvious connections to hotspots. In particular, some volcanic margins, for example, the U.S. East Coast [Sheridan et al., 1993] and northwest Australian Cuvier [Hopper et al., 1992] margins, appear to have formed away from known hotspots. These observations, together with well-documented transient and persistent LIP-forming magma sources, suggest that more than one source model may be required to explain LIPs in general and volcanic margins in particular.

Mesozoic/Cenozoic LIP emplacement peaked near the middle of Cretaceous time (Figure 2) [e.g., Larson, 1991b]. To date, there has been no detailed explanation as to why this peak occurred, although recent speculation has centered on heightened activity at the core-mantle boundary (D" layer) preceding crustal emplacement of numerous LIPs in the Pacific Ocean [Larson, 1991a, b], including the largest known igneous province, the Ontong Java Plateau. Crustal production at mid-ocean ridges has varied far less than LIP crustal production over the past 150 m.y. [Larson, 1991b], suggesting that LIP production and seafloor spreading reflect fundamentally different mantle processes.

LIPs are found in a variety of tectonic settings, including the axes of seafloor spreading centers (e.g., Iceland), triple junctions (e.g., Shatsky Rise) [Nakan-

ishi et al., 1989], old oceanic lithosphere (e.g., Hawaii), passive margins (e.g., North and South Atlantic volcanic margins), and cratons (e.g., Siberian traps). Those situated at spreading centers, such as Iceland, Azores, and Galapagos, are readily distinguishable from "normal" oceanic crust by their anomalous volumes and geochemical signatures [e.g., *Zindler and Hart*, 1986; *Hart et al.*, 1992]. Recent discussion has focused on whether mantle plumes and plate tectonics operate largely independently [e.g., *Hill et al.*, 1992] or are intimately linked [e.g., *Anderson et al.*, 1992a, b]; no consensus has been reached.

LIPs that are indisputably emplaced at plate boundaries include those now observed to lie on active spreading centers, those which are reliably dated to be the same age as adjacent "normal" oceanic crust, and volcanic passive margins. As higher-quality seismic reflection data become available on passive margins, an increasing number appear to be "volcanic" [*Coffin and Eldholm*, 1992]; thus massive volcanism may be quite common when continents separate.

Many LIPs, however, are difficult to associate with plate separation. Part of the reason is that most oceanic crust is Cretaceous or younger; nearly all older oceanic crust with accompanying LIPs has been subducted. This makes tying older LIPs to younger LIPs or to presently active hotspots difficult. The paucity of hotspot tracks on continents, however, could suggest that plate separations produce favorable conditions for thermally anomalous mantle to upwell. Alternatively, continental lithosphere might be robust enough to deflect all but the most vigorous mantle upwellings to oceanic areas [*Thompson and Gibson*, 1991].

Given the potential characteristics required for a mantle upwelling to reach the surface and form a LIP (the size of the mantle upwelling, its temperature, and its temporal variability, as well as the heterogeneous vulnerability of the lithosphere), it is remarkable that hotspot trails are in some cases traceable for more than 100 m.y. It further implies that the long-lived tracks, including those emanating from Hawaii, Kerguelen, Réunion, Iceland, and Tristan da Cunha, are among the most stable and vigorous.

3. DIMENSIONS, CRUSTAL STRUCTURE, AGES, AND EMPLACEMENT: A STUDY OF FIVE LIPS

Evaluation of how LIP emplacements affect Earth history depends on key parameters, especially size, composition, emplacement rate, and emplacement environment. Information on these parameters in different provinces is highly variable, and each LIP category offers particular advantages in understanding LIPs as a whole. Continental flood basalts provide by far the most data on composition and age. The areal extent of continental LIPs is easier to measure than that of submarine LIPs, but subaerial erosion of basalt

is much more common than marine erosion. Thus uncertainty in the areas originally covered by continental flood basalts increases with increasing age of the province. The other two main categories of LIPs, volcanic passive margins and oceanic plateaus, suffer from severe undersampling of igneous basement and overlying sediment. For example, volcanic margins of the North Atlantic, the best studied of any worldwide, have been penetrated by fewer than ten drill holes of which only two have sampled more than 100 m of basement. Igneous basement of the two largest oceanic plateaus, Ontong Java and Kerguelen, which together cover an area nearly half the size of the conterminous United States, has been sampled at only seven sites, of which only one was drilled more than 100 m. Yet seismic refraction, reflection, and gravity data, necessary for investigating crustal structure and hence volumes of LIPs, are easier to obtain and thus more prevalent from oceanic than from continental LIPs. Furthermore, the simpler lithospheric setting of oceanic plateaus permits less ambiguous evaluation of crustal structure than of continental flood basalts or volcanic passive margins. To study temporal and spatial aspects of LIP development, we examine characteristic examples which are among the best known LIPs (Figures 1 and 3): Ontong Java and Kerguelen-Broken Ridge oceanic plateaus, North Atlantic rifted volcanic margins, and Deccan and Columbia River continental flood basalts.

LIP Parameters: Methodology

On the basis of existing data and new calculations [*Coffin and Eldholm*, 1993] we determine the magnitude and range of key LIP parameters. Seismic (e.g., Figure 4), outcrop, and borehole data from LIPs show that they are major constructions of extrusive igneous rock underlain by intrusive rock with crustal thickness ranging from 20 to 40 km. Despite scarce crustal data, an expanded crustal thickness and a lower crustal body characterized by high velocities ($7.0–7.6$ km s^{-1}) appear typical for many LIPs (Figure 5). These features exist on Hawaii, the best known mid-oceanic LIP. Lower crustal bodies are believed to be added to the crust as part of the magmatic event, which also results in massive extrusive activity [*Watts and ten Brink*, 1989; *White and McKenzie*, 1989].

Surface areas of the five LIPs examined are either calculated or taken from published work (Figures 3 and 6 and Table 3). For the giant oceanic plateaus we determine LIP boundaries from bathymetry. For onshore and offshore portions of the North Atlantic volcanic province, we employ published values based on drilling results and seismic data. Areas of continental flood basalts are also taken from the literature. These estimates do not take into account any rock that may have been eroded. We compile both present and preerosional estimates for the Deccan traps and present estimates for the Columbia River flood basalts.
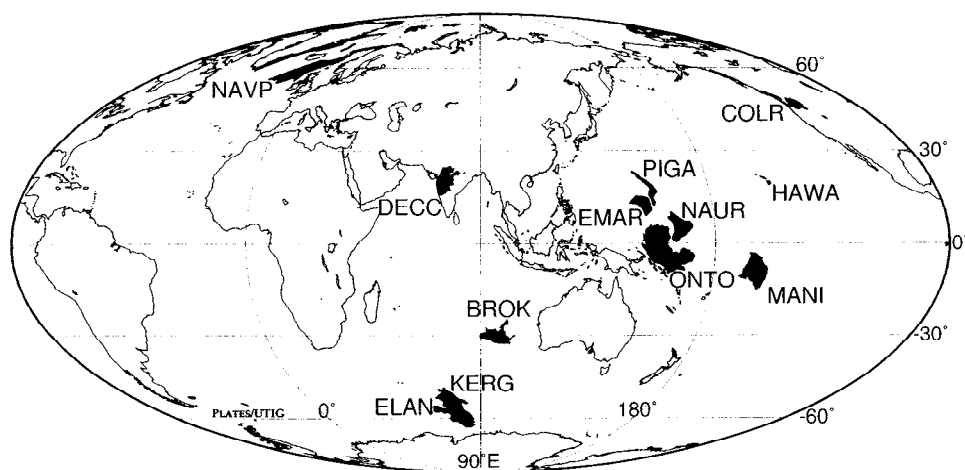
Figure 3. Locations of Broken Ridge (BROK), Columbia River basalts (COLR), Deccan traps (DECC), East Mariana Basin flood basalts (EMAR), Elan Bank (ELAN), Hawaiian Islands (HAWA), Kerguelen Plateau (KERG), Manihiki Plateau (MANI), Nauru Basin flood basalts (NAUR), North Atlantic volcanic province (NAVP), Ontong Java Plateau (ONTO), and Pigafetta Basin flood basalts (PIGA). Digital map courtesy of the Plates project (Institute for Geophysics, University of Texas, Austin).

LIP volumes are calculated from areal and vertical dimensions, the latter obtained primarily through seismic reflection and refraction investigations of LIP crustal structure. For the upper crust these geophysical data are commonly correlated with outcrop and/or drilling results to arrive at structural and stratigraphic interpretations. Absence of such data from the deeper crust forces geological interpretations to rely more on ideas and forward models. Given that drilling has not yet penetrated an entire crustal section [e.g., *Dick et al.*, 1992], even where it is relatively thin, for example, along the flank of the East Pacific Rise, our knowledge of the crust of much thicker LIPs is likely to rely solely on seismic data and interpretations for quite some time. It is important to bear in mind that only the uppermost extrusive components of all of these LIPs have been sampled and dated; intrusive material forms a large proportion of volcanic passive margins and oceanic plateaus, if not continental flood basalts, and remains unsampled and undated.

Another approach for determining oceanic LIP dimensions has been to use global topographic data and assumptions of Airy isostasy [*Schubert and Sandwell*, 1989]. The latter method, however, is unable to distinguish different age provinces within an individual LIP, to uniquely sense the presence of lower crustal bodies,
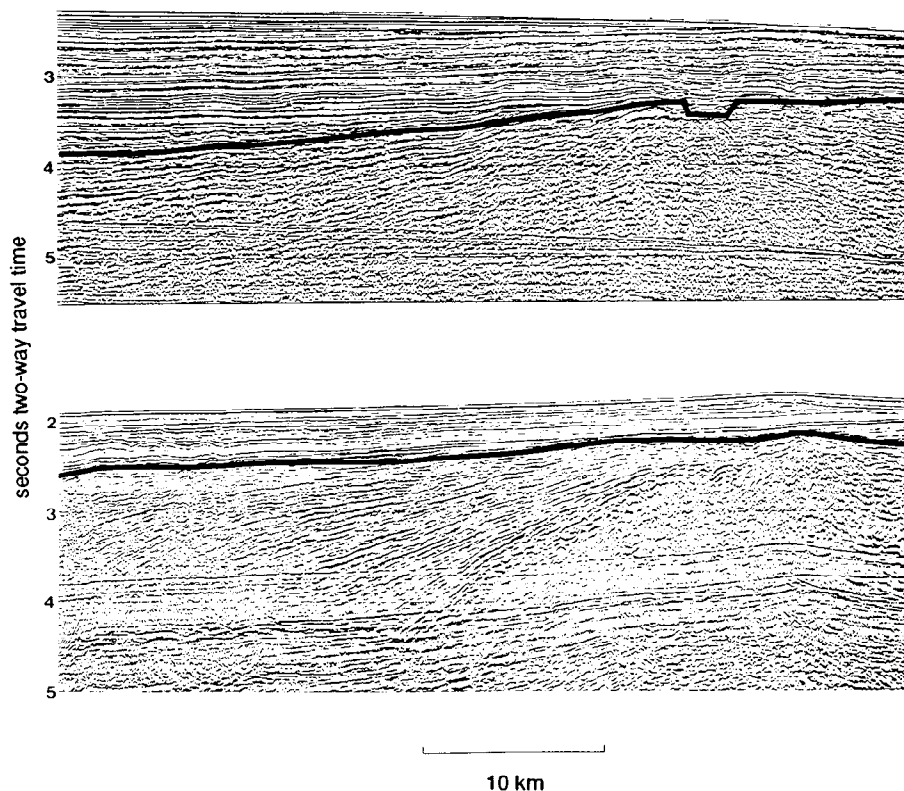
Figure 4. Dipping reflector sequences (below thick black lines) consisting primarily of basalt flows on (top) the Kerguelen oceanic plateau and (bottom) the Norwegian volcanic passive margin. The dipping wedge off Norway was sampled at Ocean Drilling Project site 642; no dipping reflectors were drilled on Kerguelen.

10 km
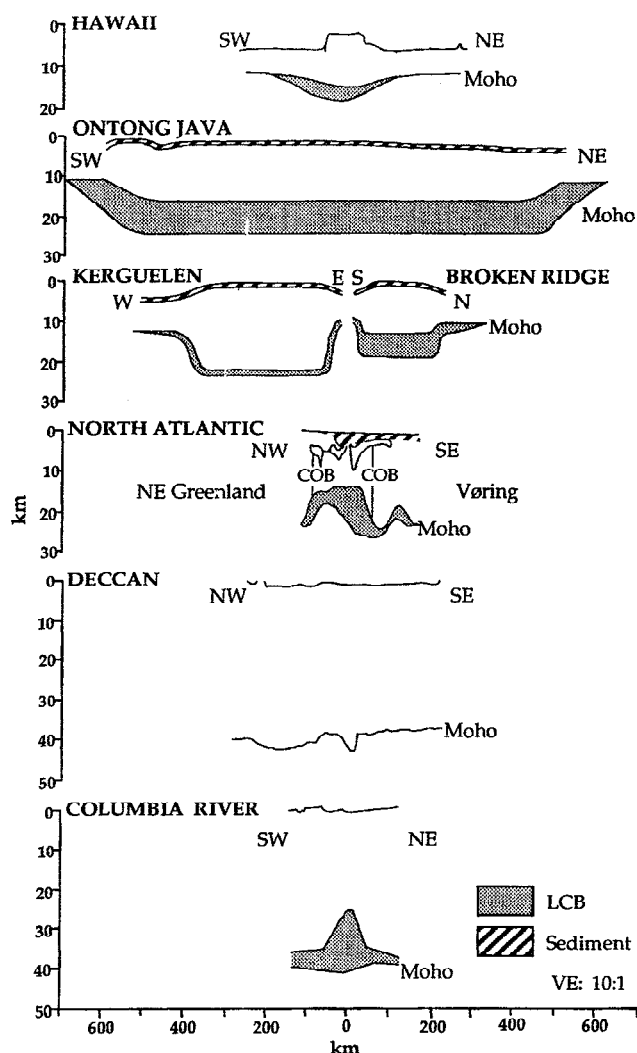
Figure 5. Simplified crustal structures of five LIPs examined plus Hawaii. LCB indicates the lower crustal body characterized by high (7.0–7.6 km s^{-1}) seismic velocities. Open areas above lower crustal bodies indicate other igneous crust and possible sedimentary and metamorphic rocks beneath Deccan traps and Columbia River basalts. For Ontong Java Plateau, crustal thickness is maximum Airy compensation depth [Sandwell and Renkin, 1988]. For North Atlantic volcanic province, dipping reflector wedges are outlined, and continent-ocean boundaries (COB) are indicated.

or to calculate volumes of ocean basin flood basalts. It also relies on more assumptions than the method combining seismic, topographic, and sampling data, which we prefer to use on the five representative LIPs (Figure 6 and Table 3).

For oceanic plateaus we calculate crustal volumes from available refraction and geoid studies (Figures 5 and 6 and Table 3), employing Hawaii as a template in determining geometries of the lower crustal bodies and Moho. Volume calculations for oceanic plateaus depend strongly on whether the features formed off or on ridge; therefore we calculate for both cases, assuming

7.1 km thick [White et al., 1992] and no preexisting oceanic crust, respectively. Furthermore, transient events may have generated excess crust outside the plateaus proper. Where adjacent coeval features exist, we include them in our calculations (Figures 3 and 6).

Seismic refraction data show a lower crustal body beneath the Columbia River basalts [Catchings and Mooney, 1988]. To calculate crustal volume beneath the extrusive cover, we assume an identical configuration beneath the entire province (Figure 5). A similar procedure is applied to the Deccan traps. Crisp [1984] examined ratios of intrusive to extrusive mafic rock volumes worldwide and documented a loosely constrained range of 3:1 to 16:1. We note that our calculated volumes for the intrusive components of the two continental flood basalt provinces fit within this range (Table 3). Finally, we use the volumes of Eldholm and Grue [1993] for the conjugate North Atlantic volcanic margins.

We employ published data and results in assessing the ages of the five LIPs examined. Radiometric, biostratigraphic, magnetostratigraphic, and marine magnetic anomaly methods have all been employed; the advent of the ^{40}Ar/^{39}Ar mineral separate technique has begun to have a particularly large impact on the dating of LIPs. More precise and reliable dates [e.g., Renne et al., 1992] will allow more accurate estimates of emplacement durations of LIPs and the correlation of their emplacement with other geological, biological, oceanographic, and environmental events. At present, however, dating of submarine LIPs is limited by a dearth of samples rather than a lack of appropriate techniques.

Emplacement rates are calculated using our volume determinations and published age information. Given existing data, we believe our estimated dimensions of the five LIPs are the best now available, despite significant uncertainties. Nonetheless, absolute values will undoubtedly change with new data.

Below we describe and discuss parameters of relatively well-sampled LIPs representing each of the three major categories: oceanic plateaus, volcanic margins, and continental flood basalts.

Giant Oceanic Plateaus

The Ontong Java Plateau (Figure 3) is bounded to the southwest by the north Solomon trough and elsewhere is defined by the 4000-m contour [International Hydrographic Organization/Intergovernmental Oceanographic Commission/Canadian Hydrographic Service (IHO/IOC/CHS), 1984]. It includes only that part of the Caroline seamount chain which impinges on the northwestern plateau. The Ontong Java Plateau sensu stricto encompasses 1.86 × 10^6 km^2, roughly one-third of the conterminous United States (Figure 3 and Table 3). Datable igneous rock has only been recovered from three scientific drill sites on the Ontong Java Plateau (289, 803, and 807), one in the Nauru Basin (462), one
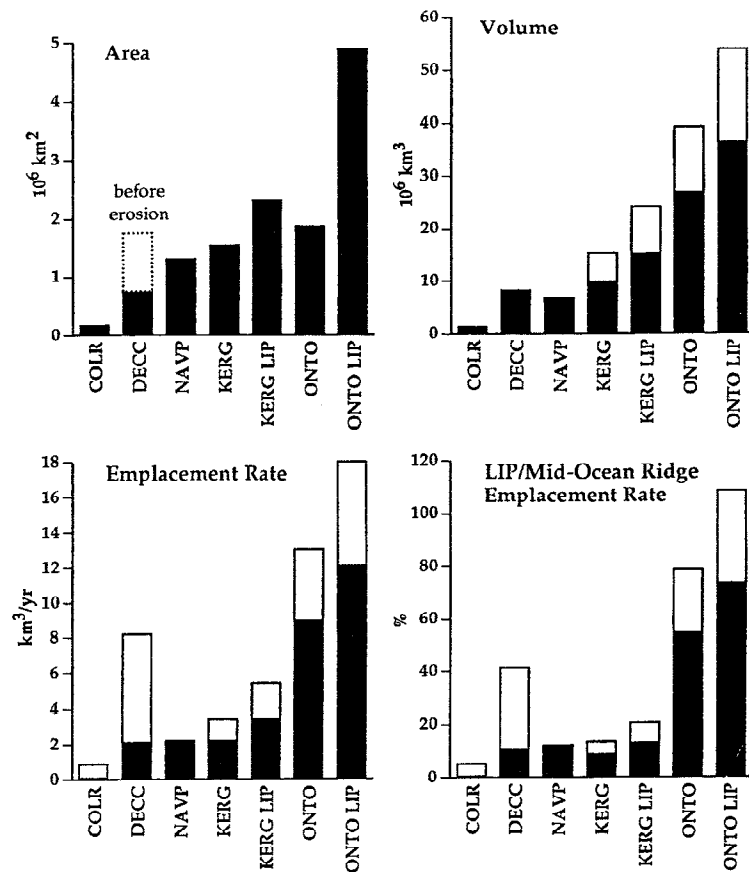
Figure 6. LIP areas, volumes, crustal emplacement rates, and emplacement rates relative to mid-ocean ridge system (last values from *Larson* [1991*a*]). Minimum values are indicated by solid columns; solid and open parts depict maximum values. Key to shading and abbreviations as in Figures 2 and 3.

in the East Mariana Basin (802), two in the Pigafetta Basin (800 and 801), and one on the Manihiki Plateau (317). *Mahoney et al.* [1993] report that ^{40}Ar/^{39}Ar dates of igneous basement at Deep Sea Drilling Project (DSDP) site 289, Ocean Drilling Program (ODP) site 807, and on Malaita Island are all ~122 Ma. Igneous basement at DSDP site 803, however, yielded a date of ~90 Ma, suggesting that some volcanism continued for 30 m.y. or more. *Tarduno et al.* [1991] have argued that the plateau is the same age (early Aptian) as Nauru Basin flood basalts and the Manihiki Plateau based on radiometric, magnetostratigraphic, and biostratigraphic dating of sediment and basement. Acknowledging uncertainty in absolute ages, they preferred emplacement between 121 and 124 Ma (timescale of *Harland et al.* [1990]), although they raised the possibility of emplacement between 117.7 and 118.2 Ma (timescale of *Harland et al.* [1982]) (Table 3). Although ^{40}Ar/^{39}Ar dates of ODP samples in the Pigafetta and East Mariana basins [*Pringle*, 1992] differ from those of *Tarduno et al.* [1991] by ≤7 m.y., their proximity and similarity to Nauru basalts suggest that they may belong to the same LIP, which would affect 4.9 × 10^6 km^2, or close to 1% of the Earth's surface (Figures 1, 3, and 6 and Table 3). Recently, *Winterer* [1992] has reported a date of ~121 Ma for Resolution guyot in the western Mid-Pacific Moun-

tains (Figure 1), which suggests that the Ontong Java LIP (sensu lato) may be larger still.

Furumoto et al. [1976] determined crust as thick as 42.7 km on the central Ontong Java Plateau from refraction data. Maximum average thickness is 39 km including a ~16-km-thick lower crustal body (7.4–7.7 km s^{-1}). They noted, however, that the free-air gravity value of +15 mGal differs by ~200 mGal from that expected from the crustal structure. *Sandwell and Renkin* [1988] examined satellite geoid data and topography and determined a maximum Airy compensation (Moho) depth of 25 km. Basalt flows and sills have been identified in the adjacent Nauru [*Shipley et al.*, 1993], East Mariana, and Pigafetta [*Abrams et al.*, 1993] basins, but no deep crustal data are available from them. Refraction Moho beneath the Manihiki Plateau has not been determined [*Sutton et al.*, 1971]; studies of global topography assuming Airy isostasy have estimated an average crustal thickness of only 10.8 km [*Schubert and Sandwell*, 1989].

Crustal thickness is thus uncertain because of differing results from seismic refraction experiments (≤42.7 km) and gravity studies (~25 km). Because of this, we have calculated volumes for both thicknesses (Table 3). Neither of these estimates include crust that may have been subducted. For comparison with other LIPs, however, we employ the conservative value

TABLE 3. Dimensions, Age Ranges, and Emplacement Rates of Five Large Igneous Provinces

Large Igneous Province	Area, 10^6 km^2	Reference	Volume, 10^6 km^3	Reference	Age Range, Ma	Reference
Ontong Java						
Ontong Java Plateau[b]	1.86	1	49.0–61.3[c]	1, 8	117.7–118.2[d]; 121–124[e]	11
Ontong Java Plateau[g]	1.86	1	26.9–39.2[c]	1, 9	117.7–118.2[d]; 121–124[e]	11
Ontong Java Plateau[h]	1.86	1	8.4	4	117.7–118.2[d]; 121–124[e]	11
Nauru Basin	1.75	2	0.856	2	117.7–118.2[d]; 121–124[e]	11
Manihiki Plateau	0.77	1	8.8–13.6[c]	1	117.7–118.2[d]; 121–124[e]	11
Pigafetta and East Mariana basins	0.50	3	0.25	3	114.6–126.1	12
Total	4.88	1	36.4–76.0[c]	1	117.7–118.2[d]; 121–124[e]	11
Kerguelen (Cretaceous)						
Kerguelen Plateau	1.54	1	9.9–15.4[c]	1	109.5–114	13
Kerguelen Plateau[h]	1.54	1	7.7	4	109.5–114	13
Elan Bank	0.24	1	1.2–1.9[c]	1	109.5–114	13
Broken Ridge	0.51	1	4.1–6.9[c]	1	88.0–89.2	14
Total	2.30	1	15.2–24.1[c]	1	109.5–114	13
North Atlantic volcanic Province	>1.3	4	6.63	4	54.5–57.5	4
North Atlantic volcanic Province[h]	>1.3	4	1.8	4	54.5–57.5	4
Deccan						
Deccan traps	0.5–1.5	5	8.2	1, 5	64.5–65.5; 65–69	15
Deccan traps[h]	0.5–1.5	5	>1.5	5	64.5–65.5; 65–69	15
Seychelles	0.25	6	?	6	64.5–65.5; 65–69	6
Total	0.75–1.75	1	8.2	1	64.5–65.5; 65–69	15
Columbia River basalts	0.1637	7	1.3	1, 7, 10	6–17.5; 15.7–17.2	16

Timescales of *Berggren et al.* [1985] and *Kent and Gradstein* [1985] are used except where noted. Note 1, this paper; 2, *Shipley et al.* [1993]; 3, *Abr*
5, *Mahoney* [1988], *Verma and Banerjee* [1992]; 6, *Devey and Stephens* [1992]; 7, *Tolan et al.* [1989]; 8, *Furumoto et al.* [1976]; 9, *Sandwell and Ren*
11, *Tarduno et al.* [1991]; 12, *Pringle* [1992]; 13, *Leclaire et al.* [1987]; *Whitechurch et al.* [1992]; 14, *Duncan* [1991]; 15, *Baksi* [1990]; *Vandamme* *
16, *Hooper* [1988]; *Baksi* [1989].
[a] At 5–30% partial melting.
[b] Assumes crustal thickness from seismic refraction experiments.
[c] Minimum and maximum volumes assume off-and on-ridge emplacement, respectively.
[d] Timescale of *Harland et al.* [1982].
[e] Timescale of *Harland et al.* [1990].
[f] First range, minimum volume (preexisting crust), maximum and minimum age range; second range, maximum volume (no preexisting crust), n
[g] Assumes crustal thickness from satellite altimetry data.
[h] Extrusive component.
[i] Assuming 0.5-m.y. eruption period for two-thirds of the basal.
[j] For 80% of the lavas.

(Figure 6). For the Manihiki Plateau, the same age as Ontong Java [*Tarduno et al.*, 1991], we assumed a maximum value of 25 km (Table 3), identical to the lower value for Ontong Java. Volume estimates of basalt flows and sills are available for the adjacent Nauru [*Shipley et al.*, 1993], Pigafetta, and East Mariana basins [*Abrams et al.*, 1993], but it is not possible to estimate volumes of any deeper crustal components. For the entire Ontong Java LIP, calculated crustal accretion rates are 12.1 and 18.0 km^3 yr^{-1} for off- and on-ridge emplacement, respectively (Figure 6 and Table 3). Using refraction, Moho for the Ontong Java Plateau would increase these rates by ~60 and ~40%, respectively.

Three drill sites on the Ontong Java Plateau recovered tholeiitic basalt presumed to represent basement. The basalt appears to record high degrees of partial melting (~30%), and isotopic signatures indicate a hotspot-like mantle source [*Mahoney et al.*, 1993]. No shallow water sediment overlies basement at these drill sites [*Tarduno*, 1992], and at each site, basalt apparently erupted in a submarine environment [*Berger et al.*, 1992]. Seismic data do not indicate subaerial exposure and erosion of basement [*Kroenke et al.*, 1991], but subsidence analysis indicates that at least part of the Ontong Java Plateau was emplaced near sea level [*Tarduno*, 1992].

Kerguelen Plateau and Broken Ridge are conjugate, Early Cretaceous LIPs in the Indian Ocean (Figures 1 and 3) which separated in Eocene time [*Houtz et al.*, 1977; *Mutter and Cande*, 1983]. Cretaceous parts of the features are difficult to distinguish on the basis of bathymetry (*Schlich et al.* [1987] for Kerguelen and *IHO/IOC/CHS* [1984] for Broken Ridge) alone, because they are continuous with younger parts. However, using dates from rock samples (Table 3), plate reconstructions [e.g., *Royer and Sandwell*, 1989; *Royer and Coffin*, 1992], and changes in bathymetry, we interpreted the break between Mesozoic and Cenozoic parts to be just south of the Kerguelen Islands and between Broken and Ninetyeast ridges. Elan Bank, a distinct structural entity of unknown age adjacent to the southern Kerguelen Plateau [*Coffin et al.*, 1986], was treated independently but assumed to be part of the Early Cretaceous Kerguelen Plateau. Together, the two conjugate Cretaceous features cover 2.3 × 10^6 km^2 (Figures 3 and 6 and Table 3).

Whitechurch et al. [1992] employed ^{40}Ar-^{39}Ar techniques on whole rock ODP basement samples and obtained a wide variety of dates but concluded that the bulk of the plateau formed at ~110 Ma (Table 3). Their most reliable date, 109.5 Ma, and the 114-Ma K-Ar basement date of *Leclaire et al.* [1987] have been used to bracket the igneous event. *Duncan* [1991] determined two ^{40}Ar-^{39}Ar age ranges for Broken Ridge dredge samples, 83–89 and 62–63 Ma, and proposed that the ridge formed during the older interval, while the younger dates mark early rifting between Broken

Ridge and Kerguelen Plateau. Turonian (88.5–91 Ma) sediment recovered from Broken Ridge [*Peirce et al.*, 1989; *Resiwati*, 1991], however, implies older basement. Furthermore, *Royer and Sandwell*'s [1989] plate model suggests that rifting between Broken Ridge and Kerguelen Plateau continued from Early Cretaceous to Eocene time; we conclude that both age intervals [*Duncan*, 1991] represent rifting events and that the basement age of Broken Ridge is similar to that of Kerguelen Plateau.

Seismic refraction experiments on the Cretaceous Kerguelen Plateau [*Charvis and Operto*, 1992] indicate 23- to 25-km-thick crust (Figure 5), including a variable thickness lower crustal body (7.3 km s^{-1}). Refraction crustal thickness concurs with results of gravity modeling, which indicate 20- to 23-km-thick crust [*Houtz et al.*, 1977]. We do not consider refraction results from the Kerguelen Islands, because Quaternary volcanism suggests a different thermal and, probably, a different seismic crustal structure. Refraction data from Broken Ridge [*Francis and Raitt*, 1967] reveal a 20.5-km-thick crust, including a 5.9-km-thick lower crustal body (7.3 km s^{-1}). The difference in our on-ridge volume (Figure 6) and the ~57 × 10^6 km^3 value of *Schubert and Sandwell* [1989] occurs mainly because we considered only the Cretaceous portions of the features. A 4.5-m.y. igneous event results in emplacement rates for the entire LIP of 3.4 and 5.4 km^3 yr^{-1}, depending on whether it was constructed off- or on-ridge (Figure 6).

Four drill sites on the southern Kerguelen Plateau (738, 747, 749, and 750) recovered basalts, mostly tholeiites, interpreted as basement. A lava flow drilled at ODP site 748 is an alkali basalt. The rocks show large geochemical and isotopic variations; however, with the exception of site 738 basalts they appear hotspot related [*Storey et al.*, 1992]. Extreme isotopic compositions of a site 738 tholeiite may indicate subcontinental lithospheric contamination [*Alibert*, 1991]. Terrestrial, terrigenous, and/or shallow-water sediment overlies basement at two of the drill sites (738 and 750), and at the other two sites (747 and 749) a significant hiatus separates basement and sediment [*Barron et al.*, 1989; *Schlich et al.*, 1989]. Weathering characteristics of the basalt at all four sites and subsidence analysis suggest subaerial or shallow water emplacement [*Coffin*, 1992] in a near-tropical to temperate climate with high orographic rainfall [*Holmes*, 1992]. Angular unconformities observed on seismic reflection data (Figure 4) support subaerial exposure and erosion of basement [*Coffin et al.*, 1990], which may have lasted as long as 50 m.y. following emplacement [*Coffin*, 1992].

North Atlantic Volcanic Margins

Widespread occurrence of early Tertiary continental flood basalts comprising the North Atlantic volcanic province (Figures 1 and 3) ranks it as a major continental flood basalt [*Dickin*, 1988]. Nonetheless,
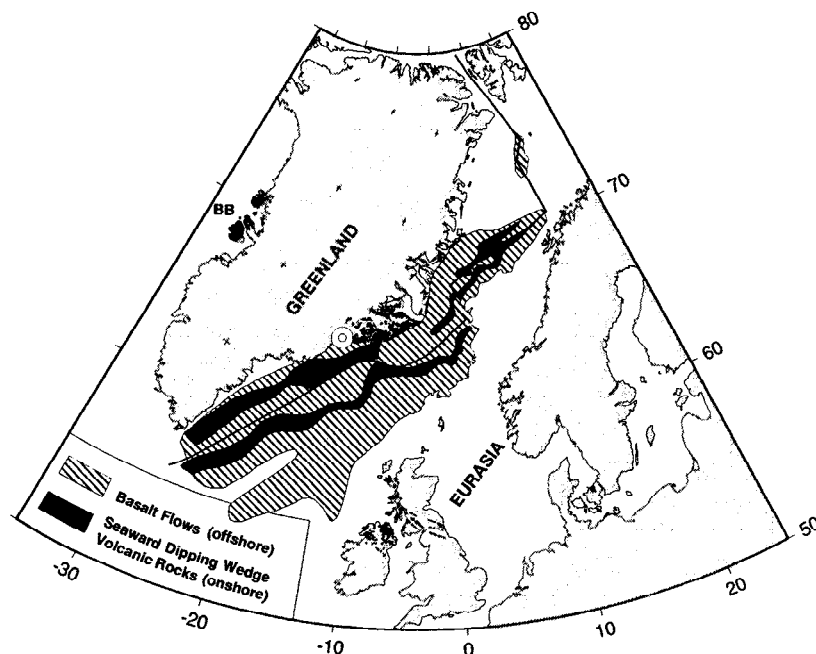
Figure 7. Extent of early Tertiary basalts plotted on a reconstruction at ~3 m.y. after breakup [after *Eldholm and Grue*, 1993]. Double circle indicates initial Iceland hotspot location interpreted by *White and McKenzie* [1989]. BB indicates Baffin Bay basalts.

its area is smaller than that of the Columbia River basalts, and its volume is only ~2.3 × 10^5 km^3. Drilling and seismic data show that subsequent to Late Cretaceous–Paleocene uplift coeval with rifting (synrift), continental breakup was accompanied by massive volcanic activity which formed oceanic crust above or just below sea level along the almost 3000-km-long rifted plate boundary (Figure 7) [e.g., *Eldholm et al.*, 1989].

Calculations of North Atlantic volcanic province dimensions, including the coeval volcanic margins, show its main part lies offshore [*Eldholm and Grue*, 1993]. By including continental crust covered by flood basalt and oceanic crust emplaced during the transient event, the area presently covered by extrusives is >1.3 × 10^6 km^2 (Figures 6 and 7 and Table 3). The total crustal volume of ~6.6 × 10^6 km^3 is based on velocity profiles along margin transects (e.g., Figure 5) and includes expanded oceanic crust, parts of the extrusive cover and the lower crustal body extending landward of the defined continent-ocean transition, and onshore flood basalts. Our values are minima, because the amount of melt intruded into continental crust after breakup cannot yet be calculated. Nevertheless, the North Atlantic volcanic province ranks among the world's most prominent volcanic margin LIPs.

Sediment-covered lavas on the margin form a smooth acoustic basement locally underlain by intra-basement reflectors, including the series of seaward dipping reflectors typical of volcanic margins (Figure 4) [*Hinz*, 1981; *Coffin and Eldholm*, 1992]. The igneous event also included sills and dikes in the continental crust, as well as subextrusive thickened crust which in most places contains a lower crustal body (Figure 5). We note that igneous crust associated with volcanic passive margins in particular and LIPs in general, need

not be characterized by dipping and/or layered reflections; these wedges are only the most prominent and recognizable component of the magmatic events.

Temporally, the igneous activity has two components. Along an almost 3000-km-long rifted plate boundary, lavas were extruded largely subaerially between 57.5 and 54.5 Ma (timescale of *Berggren et al.* [1985]), with most lavas having erupted during chron 24r. This transient component contrasts with the central region, where conjugate continental flood basalts are connected by the broad submarine ridge on which Iceland lies (Figures 1 and 3). The ridge reflects the persistent subaerial volcanism over ~60 m.y. that is commonly attributed to the Iceland plume.

On the Norwegian margin, site 642 on the inner dipping wedge (e.g., Figure 4) and 643 on crust formed ~3 m.y. after breakup have provided key information about early Tertiary environments [*Eldholm et al.*, 1989] by penetrating an entire (800 m) seaward-dipping wedge below early Eocene basal sediment. This sediment includes continental soils formed under a hot, seasonably humid climate overlain by restricted marine facies [*Thiede et al.*, 1989]. Volcanism abated 2–3 m.y. after breakup, and the injection center submerged with sporadic but waning volcanism. Subsequently, a deepwater injection center developed, leaving an isostatically compensated igneous complex trailing behind new oceanic crust accreted at a normal spreading axis. As the ocean basin matured, the high subsided at rates similar to normal oceanic crust, maintaining the difference in basement relief.

The wedge is composed of transitional mid-ocean ridge basalts and altered, interbedded, basaltic vitric tuffs. Many flows have reddened tops, while some flows and sediments indicate shallow marine deposi-

tional environments. Subaerial and neritic environments are inferred during and subsequent to breakup when the wedge was constructed by an intense phase of explosive, subaerial volcanism. The basalt rests on dacitic lavas, emplaced by infrequent eruptions during the late rift stage, and interbedded sediments indicating fluvial or shallow-water deposition.

Continental Flood Basalts

The Deccan traps, covering ~5 × 10^5 km² in western India (Figure 3), may have covered >1.5 × 10^6 km² prior to erosion (Figure 6 and Table 3) [Mahoney, 1988]. These estimates do not include any offshore component of Deccan magmatism, which is suggested by seaward-dipping reflectors on the Indian margin [Hinz, 1981]. At the time of eruption, the Seychelles were attached to India; the coeval basalt there is estimated to cover 2.5 × 10^5 km² [Devey and Stephens, 1992]. Apparently, the Deccan volcanic event covered a much larger area than reported in the literature, although a paucity of data offshore India and on the Mascarene Plateau preclude estimating both the area and volume of the total event.

The Deccan traps are predominantly tholeiitic basalts, although alkalic, felsic, and ultramafic intrusive suites form lithologically distinct but volumetrically negligible components [Mahoney, 1988]. Trace elements and isotopic ratios vary considerably within the province, and bulk compositions change as well, but most evidence suggests that the depleted source of most tholeiites is homogeneous, arguing that continental lithosphere is not the major basalt source. According to Biswas and Thomas [1992], most basalt erupted between 65 and 69 Ma, while Baksi [1990] and Vandamme et al. [1991] suggest a period of <1 m.y. (Table 3). An ^{40}Ar-^{39}Ar date of 64 Ma has been reported from Mascarene Plateau basalts ~500 km southwest of the Seychelles (Figure 1 and Table 1) [Duncan and Hargraves, 1990], implying that a significant portion of that feature may have been created by the transient Deccan event.

Verma and Banerjee [1992] have reviewed deep seismic reflection data and modeled gravity data across the Deccan traps. Interval velocities show lower crustal bodies (7.0–7.3 km s^{-1}), but thicknesses are not well defined. Gravity modeling suggests an anomalously high density crustal body beneath the traps, but geometry is poorly constrained. We estimated a lower crustal body volume by applying the ratio of surface area to deep crustal volume for the Columbia River basalts to the original surface area of the Deccan flood basalts, excluding the Seychelles and the Mascarene Plateau (Figure 6). The total Deccan traps volume of 8.2 × 10^6 km³ conservatively includes only extrusive rocks and lower crustal bodies. If the Deccan traps formed during 1- and 4-m.y. intervals, emplacement rates were 8.2 and 2.1 km³ yr^{-1}, respectively (Figure 6 and Table 3).

Tolan et al. [1989] have calculated that the Colum-

bia River basalts (Figures 1 and 3) once covered 1.64 × 10^5 km² (Figure 6 and Table 3) with an extrusive volume of 1.74 × 10^5 km³. A seismic refraction profile across the Columbia Plateau shows a lower crustal body (7.5 km s^{-1}) (Figure 5) interpreted as a "rift pillow" [Catchings and Mooney, 1988]. To calculate crustal volume beneath the extrusive cover, we assumed an identical configuration beneath the entire province. The total LIP volume amounts to 1.3 × 10^6 km³.

The great majority of Columbia River basalts are tholeiites, although some flows are characterized by transitional to alkali basalt mineralogy [Hooper, 1988]. Variable isotopic values and trace element ratios require a variable source, either in the form of a heterogeneous mantle or different degrees of crustal contamination or both. Basalts range in age from 6 to 17.5 Ma, but most erupted between 15.7 and 17.2 Ma [Baksi, 1989]. Corresponding emplacement rates are 0.1–0.9 km³ yr^{-1} (Figure 6).

Discussion

Areas and volumes of transient LIPs (Figure 6), as well as previous estimates for other continental flood basalts [e.g., White and McKenzie, 1989], reveal variations over 1 to 2 orders of magnitude. Crustal structures of the three main LIP types are similar, comprising an extrusive upper crust (Figure 4) and a lower crust (Figure 5) characterized by high seismic velocities (7.0–7.6 km s^{-1}), and differ from "normal" oceanic or continental crust. Thus these LIPs may share a common genesis. Construction of the largest igneous provinces during transient magmatic episodes significantly increases global crustal production rates (Figure 2). The North Atlantic volcanic province and recent studies of the U.S. East Coast continental margin [Holbrook and Kelemen, 1993] show that volcanic margins must be included in the LIP inventory.

Compressional wave velocities of 7.0–7.6 km s^{-1} in lower crustal bodies are not those of normal oceanic or continental crust, although they are found in some continental settings, mainly beneath rifts [e.g., Catchings and Mooney, 1988]. Composition of lower crustal bodies on volcanic margins is unknown; gabbroic, strongly mafic, and hot ultramafic rocks are possibilities [Meissner and Köpnick, 1988]. Some lower crustal bodies have been explained as magmatic underplating by accumulating mantle-derived material below the original crust [Large Aperture Seismic Experiments (LASE) Study Group, 1986; White et al., 1987]. Underplating suggests emplacement of mafic magma at the base of continental crust [Furlong and Fountain, 1986]. Density contrasts between the old crust and the melt determine whether melts are ponded below the crust or intruded into the crust [Herzberg et al., 1983]. For example, White and McKenzie [1989] suggested that decompressional melting of hot asthenospheric mantle would produce 7.1- to 7.2-km s^{-1} velocities in ponded basaltic melt because of increased MgO con-

tent. Thus underplating should only occur beneath continental crust, contradicting observed lower crustal body continuity into oceanic crust in the North Atlantic [Eldholm and Grue, 1993] and off the U.S. East Coast [LASE Study Group, 1986; Tréhu et al., 1989].

The underplating hypothesis suggests that melts are trapped by a crustal density filter during rifting. Following breakup, no such filter exists in oceanic crust, although similar anomalous velocity-density structures are found beneath oceanic crust adjacent to volcanic passive margins. Clearly, the underplating hypothesis does not adequately explain the occurrence of lower crustal bodies beneath oceanic crust. It is also arguable whether magmatic underplating takes place when the crust is strongly attenuated, for example, during the late rift stage, because crustal strength rapidly becomes negligible when massive mafic upwelling reaches the brittle crust, causing the assumption of conductive heat transfer to break down [Dixon et al., 1989]. Although the lower crustal body is most likely emplaced during breakup, we consider the term "underplating" a misnomer.

Uncertainty in absolute dimensions, however, is dramatically illustrated by the biggest LIP, Ontong Java. We estimate a conservative volume based on Airy isostasy, off-ridge emplacement, and a 3-m.y. magmatic event leading to an emplacement rate of 12.1 km^3 yr^{-1} (Figure 6). On-ridge emplacement and refraction Moho might increase the rate by a factor of 3 to ~36 km^3 yr^{-1}. In addition, emplacement rate is particularly sensitive to construction duration. For example, the alternate 0.5-m.y. duration considered by Tarduno et al. [1991] results in an off-ridge emplacement rate of 72.8 km^3 yr^{-1}. We stress, however, that well-dated basement at four sites over an area greater than half the size of the conterminous United States and penetrating, at most, the upper $\sim0.5\%$ of the total igneous section does not rule out a somewhat longer emplacement period, which would result in values resembling those of other LIPs.

Comparison of LIP crustal production with that of the global mid-ocean ridge system over the past 150 m.y. [Larson, 1991a] (Figures 3 and 6) shows that the largest LIPs had the potential to significantly alter the hydrosphere and atmosphere through their mass, heat, chemical, fluid, and particulate fluxes during isolated time periods [e.g., Larson, 1991a, b; Eldholm, 1990]. In this context, the extrusive crustal component is a key factor [e.g., Officer et al., 1987], and extrusive eruption rates for Deccan traps, North Atlantic volcanic margins, and the two giant plateaus show less variability (0.3–2.6 km^3 yr^{-1}) than do total crustal production rates [Eldholm and Grue, 1994]. This observation, and the temporal correspondence of Deccan and North Atlantic eruptions with major environmental change, may imply that LIPs smaller than Ontong Java have also contributed to these changes.

4. MANTLE DYNAMICS

LIPs are crustal manifestations of dynamic processes in the Earth's mantle; hence LIP parameters may be applied as boundary conditions to invert such processes. Presently, no consensus has developed concerning the nature and dimensions of mantle circulation required to emplace LIPs. In general, however, more effort has been devoted to development of mantle models that may produce transient and/or persistent magmatism than to verification of such models through tests employing geological data. Despite the fragmented nature of the LIPs global database, data exist that directly or indirectly relate LIPs to mantle processes. Particularly relevant are (1) dimensions and emplacement rates (see section 3), (2) composition of the igneous complexes (see sections 2 and 3), (3) spatial location relative to known hotspots (see section 2), and (4) geological setting prior to and during emplacement (in particular, the relationship of LIPs to rifting and other types of lithospheric deformation) (see section 2).

Transient LIPs exist on many scales, reflecting a wide size range of mantle melt anomalies. Recognition of this dimensional variability is fundamental in evaluating mantle processes leading to LIP emplacement. Volumetric variations among LIPs (Figure 6 and Table 3) are influenced by at least three factors: mantle source intensity, vulnerability of the lithosphere to asthenospheric penetration, and lithospheric plate speed over the source. Seismic tomography provides a broad, three-dimensional measure of source intensity, that is, volumes and relative magnitudes of asthenospheric thermal anomalies. The method outlines such features as mid-ocean ridges, dead slabs, and cratons. Global studies indicate that the upper mantle is dominated by subduction, whereas upwelling appears to dominate circulation in the lower mantle. It appears that the main mode of mantle convection is flow controlled and scaled by plates [e.g., Dziewonski and Woodhouse, 1987; Anderson et al., 1992a; Davies and Richards, 1992].

Lithospheric vulnerability is somewhat difficult to assess; it was once thought that younger lithosphere should be more vulnerable than older lithosphere to asthenosphere-derived volcanism [Gass et al., 1978; Pollack et al., 1981; Vogt, 1981], but recent analyses of present-day heat flux from oceanic lithosphere suggest rather a correlation between plate speed and vulnerability [Sleep, 1990; McNutt, 1990]. Absolute global plate motions in the hotspot reference frame are well constrained for the past ~90 m.y. [e.g., Müller et al., 1993]; plate speed over the source does not appear to influence the volumes of transient features. However, lithospheric structures, for example, fracture zones and rifts, may act as preferential conduits for asthenospheric melts [Sykes, 1978; Okal and Batiza, 1987; McNutt et al., 1989]. In most intraplate settings, melting of the plume head is inhibited by the >125-km-thick mechanical boundary layer. Therefore Kent

et al. [1992] proposed a plume incubation period to thin and partly remove the boundary layer prior to massive continental flood basalt production.

Mantle Convection and Plumes

The original hotspot/mantle plume concept [*Wilson*, 1963; *Morgan*, 1971] has recently been developed further. Although models for structure and temporal evolution of mantle plumes vary considerably, a commonly observed feature is the capability of a plume to generate large quantities of melt by decompression of upwelling, thermally anomalous mantle. Where the thermal anomaly is associated with continental breakup, it may induce the formation of a volcanic margin and adjacent continental flood basalts distinguished by transient magmatism over a wide region. If the plume initially surfaces through oceanic lithosphere, an oceanic plateau may form, and as the plate migrates over the focus of upwelling, a submarine ridge and/or seamounts may be constructed. Finally, interaction of plumes and thick continental lithosphere may, under certain conditions, allow emplacement of a continental flood basalt.

Several plume models which induce a broad region of hot mantle beneath the lithosphere have been advanced to explain emplacement of LIPs (Figure 8). One concept [*Richards et al.*, 1989; *Griffiths and Campbell*, 1990], primarily based on results of laboratory experiments, invokes a secondary mode of mantle convection in addition to the dominant plate scale mode. The former initiates LIP-forming, upwelling, buoyant plumes which detach from the weak, heterogeneous thermal boundary layer D″ at the base of the mantle. Such plumes have a large, hot head over a narrow stem (Figure 8a). When the deep plume impinges on the mechanical boundary layer at the base of the lithosphere, conductive heating and thinning of the lithosphere induce large-scale melting. The model suggests "active" rifting, that is, stress and strain are transferred from the plume to lithospheric plates. Uplift precedes and accompanies volcanism, but rifting may postdate the main volcanic event [*Griffiths and Campbell*, 1991a; *Hill*, 1991; *Hill et al.*, 1992]. The "active" plume model also suggests occasional coupling between core, mantle, and crustal processes. Although not covered in this review, magnetic field reversals, presumably originating in the core, have been related to major plate tectonic changes [e.g., *Vogt*, 1975] as well as LIP emplacements [e.g., *Larson*, 1991a, b; *Larson and Olson*, 1991], reviving and refining the ideas of *Vogt* [1972].

Another plume model [*Courtney and White*, 1986; *White and McKenzie*, 1989], developed from volcanic margin crustal structure and petrologic modeling, proposes adiabatic upwelling and decompressional melting of hot asthenosphere as a result of extending lithosphere. The plume-derived thermal mushroom (Figure 8b) causes dynamic uplift which accelerates the rate of extension and thereby the amount of melting. Thus magmatism is not driven by the plume but is a response

to lithospheric extension; maximum melting occurs during crustal breakup. This model has both "active" and "passive" elements and has been used to explain continental flood basalts and volcanic margins. A similar effect may be achieved if hot asthenosphere is trapped by a relict lithospheric relief, or "thin spots," from previous tectonic episodes [*Thompson and Gibson*, 1991; *Skogseid et al.*, 1992a].

Another concept, rooted in seismic tomography, relies on a thermally, chemically, and isotopically heterogeneous asthenosphere [e.g., *Anderson et al.*, 1992b]. Initially weak cratonic lithosphere or lithosphere weakened by plate reorganizations may be invaded by material from hot regions of upper mantle, allowing melts to surface and produce LIPs.

These models reflect fundamentally different primary convection systems in the mantle (Figure 9). *Davies and Richards* [1992], for example, visualize "whole mantle" circulation and interpret the mantle transition zone at the base of the asthenosphere (670 km depth) as an isochemical phase change. On the other hand, the "layered mantle" circulation model assumes the mantle transition zone to be a thermal boundary layer. *Anderson et al.* [1992b] state that the major predictions of the deep plume model are untestable and that geophysical techniques cannot resolve the narrow, 10- to 200-km-diameter plume tails. Furthermore, they claim that the effects of the large flattened plume heads, which seismic tomography only images to 200- to 300-km depths, may be produced by a number of phenomena consistent with upper mantle convection. *Courtney and White*'s [1986] plume model is also restricted to the upper mantle.

Mantle plumes in various forms represent the most plausible mechanism for explaining the large amounts of thermal energy required by massive melting anomalies. Although mantle plumes provide a convenient source for the excess melting, observational data are required in order to examine whether focused mantle plumes are necessary to generate LIPs. In particular, the concept that all LIPs are connected to hotspots [*White and McKenzie*, 1989] must be tested. *Mutter et al.* [1988] claim that some volcanic margins cannot convincingly be associated with known mantle plumes. For example, the U.S. East Coast margin [*Holbrook and Kelemen*, 1993] and the Cuvier margin off northwest Australia [*Hopper et al.*, 1992] are candidates for nonhotspot LIPs. Furthermore, the North Atlantic volcanic margins extend beyond the maximum diameter of the thermal anomaly predicted from common plume models [*Eldholm and Grue*, 1993], an argument that may also apply to the greater Ontong Java LIP [*Coffin and Eldholm*, 1993]. *Vogt* [1991] has pointed out that most of the midplate igneous activity and all topographic swells in the western North Atlantic and eastern North American region are difficult to reconcile with simple hotspots or plumes. He suggests they are caused either by episodic midplate stress
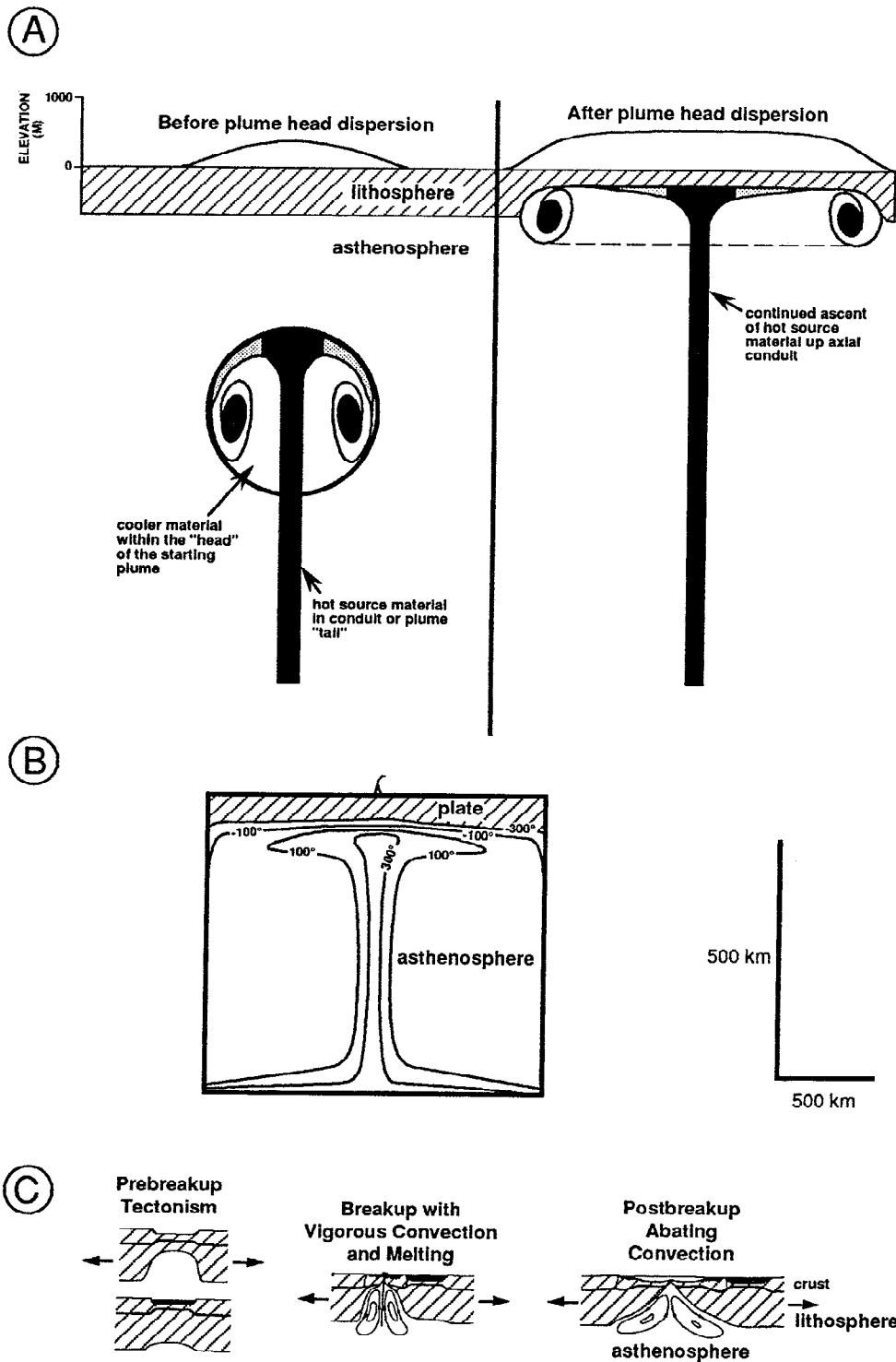
Figure 8. Three different models for LIP generation and emplacement: (*a*) Thermally induced "active" mantle plume entraining mantle material during its ascent to the base of the lithosphere where a large transient plume head develops (left), rifts the lithosphere (right), and causes excessive magmatism [*Campbell and Griffiths*, 1990], (*b*) steady state "passive" thermal plume over which rifting eventually causes excessive melting and volcanism [*White and McKenzie*, 1989], and (*c*) convective overturn caused by steep, cold lithospheric edges [*Mutter et al.*, 1988].

intensification or by shallow asthenospheric convection traveling with the plate and controlled by vertical thermal boundaries. Finally, observations of anomalous seafloor spreading crust [*Hinz et al.*, 1993; K. Hinz, personal communication, 1992] (Figure 1 and Table 2) may suggest that some smaller oceanic LIPs simply reflect periods of increased global mid-ocean ridge crustal production associated with divergent plate boundaries located over mantle regions characterized by high melt potentials.

A further feature of recent mantle plume models, as opposed to *Morgan*'s [1981] ideas, is that plumes and plate kinematics are unrelated phenomena. Some examples, such as the Hawaiian-Emperor seamount chain, strongly support this view, but some LIPs, such as the Ontong Java Plateau specifically and the Early Cretaceous Pacific volcanic events in general, are so large that they probably reflect primary modifications of mantle dynamics. Emplacement of the latter LIPs may be connected to changes in spreading rates in the
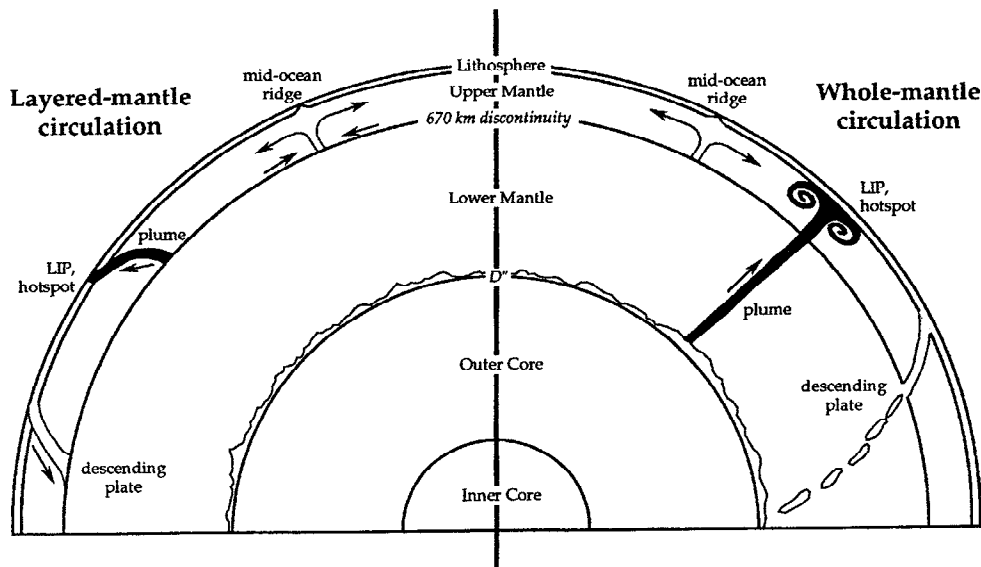
Figure 9. Cartoons of (left) layered [e.g., *Anderson et al.*, 1992a, b] and (right) whole [e.g., *Davies and Richards*, 1992] mantle circulation.

Early Cretaceous Pacific (W. J. Morgan, personal communication, 1990); one suggestion is that spreading rates increased during the Cretaceous magnetic quiet zone, because excessive plume melt lowered asthenospheric viscosity and hence resistance to plate motion (R. L. Larson, personal communication, 1993). Although some workers have proposed that plate readjustments may in places tap anomalously hot mantle and thus initiate a LIP [e.g., *Anderson et al.*, 1992a, b], such a model has yet to be rigorously tested. Other work has indicated that mantle plumes such as Hawaii must deflect in response to changes in direction of plate motion and that the shape of the bend in the surface track reflects this readjustment [*Griffiths and Richards*, 1989].

To compare dimensions of LIP sources in the mantle, we show minimum and maximum spherical thermal anomalies (Figure 10) based on volumes estimated by assuming 5–30% partial melting for basalts [e.g., *McKenzie and Bickle*, 1988; *Liu and Chase*, 1991; *McKenzie and O'Nions*, 1991; *Watson and McKenzie*, 1991; S. Eggins, personal communication, 1992]. The observed scale range and magnitude of these estimates imply that some LIPs can be explained by temperature and/or fluid anomalies restricted to a convecting upper mantle. Ontong Java and probably Kerguelen, however, appear to have been generated, at least in part, from the lower mantle (>670 km deep) (Figure 10). Furthermore, we note that degrees of partial melting are likely to be higher beneath thinner oceanic lithosphere than beneath thicker continental lithosphere, which suggests that thermal anomalies responsible for the North Atlantic volcanic province, Deccan traps, and Columbia River basalts may tend toward our maximum dimensions (Figure 10) and thus involve the lower mantle as well.

External causes for LIP magmatism, namely mete-

orites, comets, and asteroids, have been proposed [e.g., *Alt et al.*, 1988; *Rampino and Stothers*, 1988; *Rampino and Caldeira*, 1993]. A major impact would create a crater large enough to cause pressure relief melting in the asthenosphere, which would create the terrestrial equivalent of a lunar mare in the crust and persistent low-pressure cells in the mantle, that is, hotspots. Although some workers accept impacts as causes of lunar maria and flood basalts on Venus [e.g., *Solomon*, 1993], a strong case for impacts causing terrestrial LIPs has yet to be made, and devising experiments and models to test the hypothesis has lagged behind investigation of internal origins and sources.
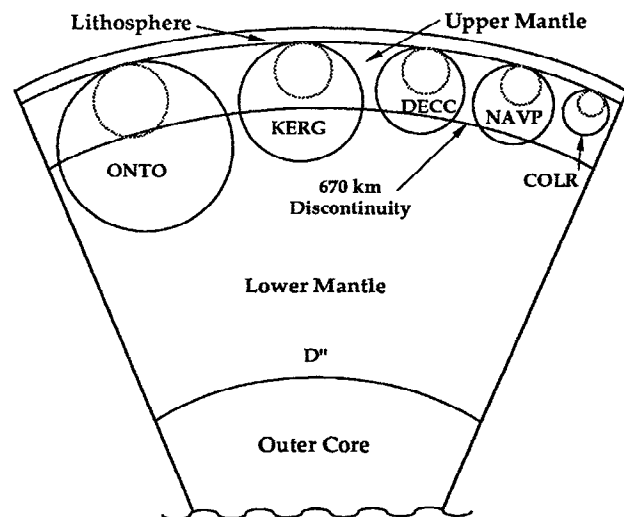
Figure 10. Minimum and maximum spherical diameters of five LIPs calculated assuming 5% and 30% partial melting (Table 1) plotted on a cross section of the Earth. For simplicity, spheres are placed below the base of the lithosphere, as experiments suggest [*Griffiths and Campbell*, 1991a, b]. Abbreviations as in Table 1.

Lithospheric Deformation: Uplift, Extension, Rifting, and Subsidence

Whether rifting precedes or accompanies emplacement of oceanic plateaus in intraplate settings is not possible to evaluate from available data. Where the mantle thermal anomaly is associated with continental breakup or located below thick continental crust, however, the relationship between thermal plumes and rifting has been much debated [e.g., *Duncan and Richards*, 1991; *Hill*, 1991; *Hill et al.*, 1992]. We have previously shown that for LIPs erupted both in intraplate and plate boundary settings, the tectonomagmatic history of LIP formation depends on interaction of the mantle asthenosphere anomaly, configuration and composition of mantle lithosphere and crust, and the amount of stress in the lithosphere. Nonetheless, one would expect volcanism and faulting to precede active rifting, while they would follow passive rifting [*Dixon et al.*, 1989].

The main tholeiitic phases of many continental flood basalts, such as Columbia River [*Hooper*, 1990], Deccan [*Hooper*, 1990], and Siberian [*Campbell et al.*, 1992], were erupted in what have been interpreted as largely nonextensional settings [*Hill*, 1991; *Hill et al.*, 1992]. Other investigators connect continental flood basalts to rifting: examples are Paraná [*Peate et al.*, 1990; *Kazmin*, 1991; *Harry and Sawyer*, 1992; *Renne et al.*, 1992], Karoo [*Kazmin*, 1991], and Deccan [*Kazmin*, 1991; *Biswas and Thomas*, 1992]. Because many continental flood basalts show little evidence of lithospheric extension, *Duncan and Richards* [1991] conclude that they result from melting events at the base of the lithosphere, whether or not such thermal events precipitate subsequent rifting. On the other hand, a nonextensional setting can probably only be expected if plume impingement at the base of the lithosphere and volcanism are nearly coeval events, whereas a more protracted deformation period, for example, the plume incubation model of *Kent et al.* [1992], would lead to regional uplift, extension, and intrusion of alkalic melts prior to volcanism. From field evidence in the Siberian, Karoo, Paraná, Rajmahal-Kerguelen, and Deccan-Seychelles-Mascarene Plateau continental flood basalt provinces, *Kent et al.* [1992] conclude that the latter emplacement history is most appropriate for continental flood basalts.

Various plume and upwelling models may all be applied to asthenospheric behavior during formation of volcanic margins. However, *Mutter et al.* [1988] have proposed an alternative model depending on development of short-lived, small-scale convection close to the trailing edges of cold, thick continental lithosphere (Figure 8c). Convection induces a pulse of excess magmatism during the onset of seafloor spreading. This entirely "passive" model, derived from data on the North Atlantic Vøring and northwest Australian margins [*Hopper et al.*, 1992] and driven by plate separation, does not require increased mantle temper-

atures, although they would facilitate large melt volumes. Implicitly, it requires little lithospheric extension prior to a relatively sudden breakup. However, recent work by *Skogseid et al.* [1992a] on the Vøring margin and its conjugate records a 16- to 18-m.y. rift phase prior to excess magmatism during and subsequent to early Tertiary continental breakup. Furthermore, they suggest that Late Cretaceous-Paleocene North Atlantic rifting affected a 300- to 350-km-wide crustal region. The rift zone is thus wider than the thickness of the lithosphere, and a wide region of asthenospheric upwelling has been predicted by *Eldholm and Grue* [1994], applying the concept of *Keen* [1987]. As the region of lithospheric extension narrowed and the deformation rate increased with time, melt production also increased, climaxing during breakup. Alkali basalts and gabbros, emplaced in continental crust, were produced during early rifting and for low stretching factors; melts became tholeiitic when the lithosphere was extended further [*McKenzie and Bickle*, 1988]. Final crustal extension, or the late rift stage leading to breakup, occurred rapidly, and the main excess melts may have surfaced within ~1 m.y. [*McKenzie*, 1984].

Each mantle plume model should predict characteristic lithospheric uplift and subsidence histories associated with LIP emplacement and evolution. Experimental results of "active" plume modeling predict surface uplift on a wavelength comparable to the diameter of the plume, uplift which commences before extension and ~25 m.y. prior to maximum uplift (Figure 11a) [e.g., *Griffiths and Campbell*, 1991a]. The uplift is controlled by the buoyancy of the hot plume material and lateral variation in lithospheric rheology and thickness. Maximum uplift can reach 1000 m in continental, and likely more in oceanic, lithosphere and precedes initiation of major basaltic volcanism by 3 to 30 m.y. [*Hill et al.*, 1992]. Results of numerical "active" plume modeling demonstrate that a plethora of uplift and subsidence histories are possible, given various lithospheric ages, plate speeds, and many more loosely constrained model parameters [*Monnereau et al.*, 1993].

Volcanism is a "passive" response to lithospheric thinning in *White and McKenzie*'s [1989] plume model; they predict that the bulk of LIP emplacement occurs contemporaneous with the main rifting. Uplift and subsidence are affected by lithospheric thinning, addition to the crust of igneous material produced by adiabatic decompression, dynamic support of the mantle plume, and reduction in density of residual mantle after removal of melt. Interaction of the various factors produces a variety of uplift and subsidence models (Figure 11b). Uplift and subsidence histories have not yet been predicted for the *Mutter et al.* [1988] and *Anderson et al.* [1992a, b] models.

Data with which to test the various models' uplift and subsidence histories are sparse. The geological
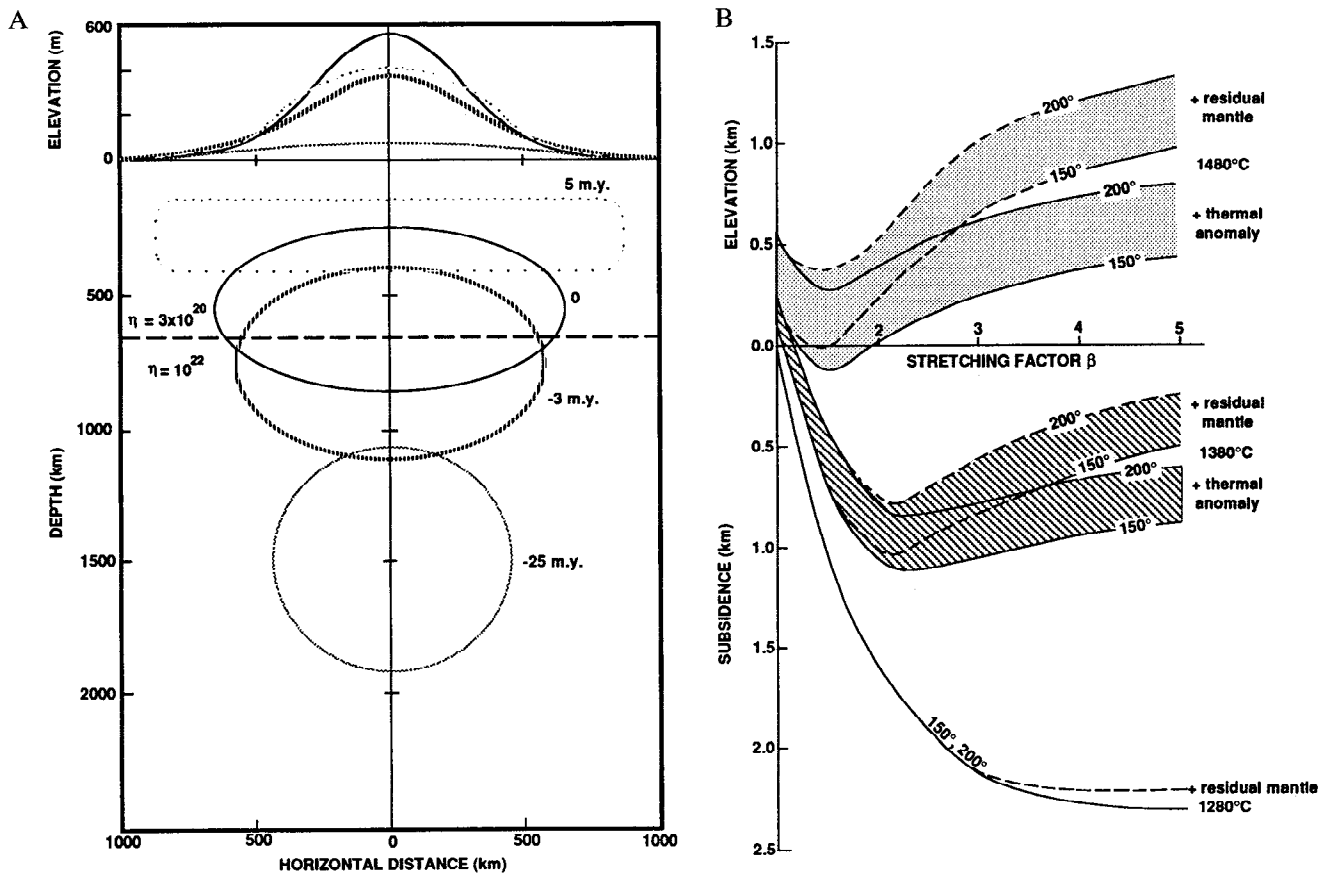
Figure 11. Vertical tectonic histories predicted by (a) "active" and (b) "passive" plume models. Figure 11a represents surface uplift as a function of plume head location in the mantle [after *Griffiths and Campbell*, 1991a]. Figure 11b represents uplift and subsidence as a function of lithospheric extension (beta) and residual mantle potential temperature [after *White and McKenzie*, 1989].

record of uplift is not commonly preserved, and subsidence studies have commonly focused on decay of a "normal" mid-ocean ridge thermal anomaly and on mechanical loading by sediment and water. Oceanic plateaus, in contrast to thermal/dynamic swells, offer arguably the best opportunity for testing and improving transient plume models in the simplest lithospheric setting; subsidence studies [*Coffin*, 1992; *Detrick et al.*, 1977] using DSDP and ODP material suggest that oceanic LIPs follow a subsidence curve dominated by thermal decay (Figure 12) which appears similar to that of "normal" oceanic lithosphere [*Parsons and Sclater*, 1977]. Preemplacement and synemplacement uplift of oceanic plateaus has not yet been investigated.

Although the question of how lithospheric uplift, extension, rifting, and subsidence relate to LIP emplacement and evolution is not fully resolved, a review of these LIP categories reveals a preponderance of extensional settings [*Coffin and Eldholm*, 1992]. The interpretation of purely nonextensional continental flood basalts may result from low stretching factors for provinces unrelated to plate boundary events, such as Siberian traps. For continental flood basalts adjacent

to volcanic margins, the tectonomagmatic history can only be appreciated if flood basalts are studied in the framework of the entire rift system, that is, in a conjugate margin setting [*Eldholm*, 1991]. Except for the North Atlantic, no such studies have been made. Nonetheless, many continental flood basalts arriving from the same mantle anomaly which causes continental breakup and volcanic margin formation actually rest on crust outside the rift zone or in the part of the rift zone characterized by low stretching factors. This effect is amplified if breakup occurs along a detachment plane near the flank of the rift zone where the crust is weak, for example, along a preexisting fault or thrust plane, possibly induced by small-scale convection near the edges of the wide upwelling zone [*Keen*, 1987]. Many rift models predict extension over a zone of upwelling. Other models show the locus of crustal failure displaced laterally because of a preexisting weak zone or a detachment [*Bassi and Bonnin*, 1988; *Braun and Beaumont*, 1989; *Dunbar and Sawyer*, 1989; *Lister et al.*, 1991; *Sawyer and Harry*, 1991; *Harry and Sawyer*, 1992]. This breakup mode does not depend on a low-angle fault through the entire lithosphere but rather on a fault passing through the crust
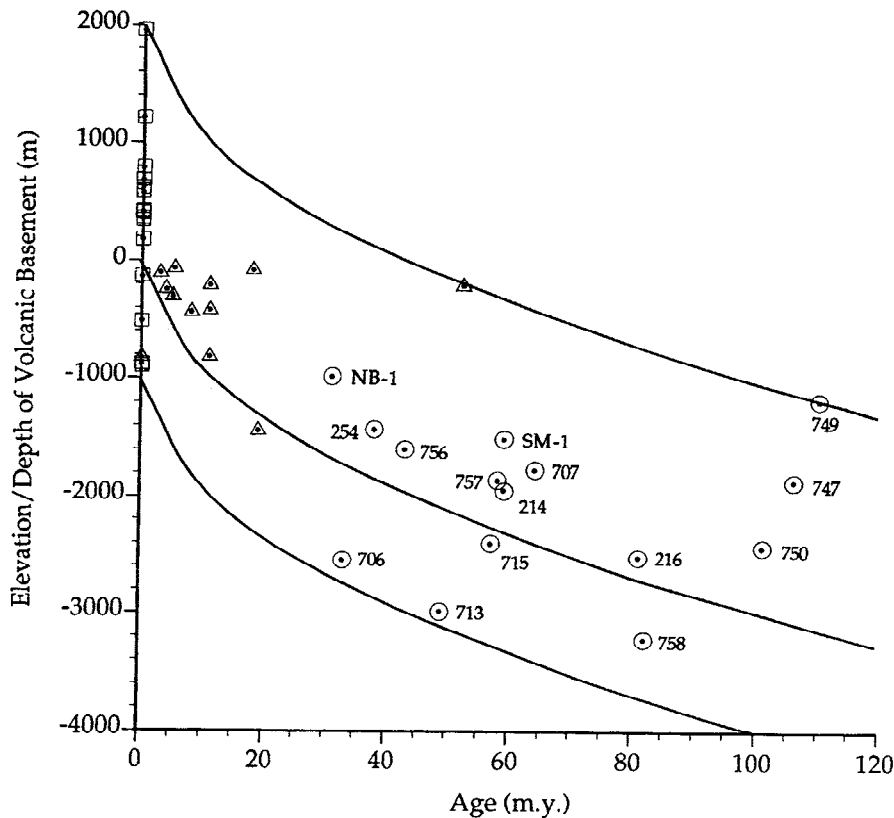
Figure 12. Age-depth curves for LIP volcanic basement at Indian Ocean drill sites [after *Coffin*, 1992]. Numbers identify Deep Sea Drilling Project and Ocean Drilling Program sites; letters and numbers indicate industry wells. Calculated basement depths for all sites are given for zero age (boxes) and the present (circles); where possible, calculated basement depths at the end of shelf deposition are indicated (triangles). Theoretical subsidence curves according to the equation of *Hayes* [1988] for crust emplaced 2000 m above sea level and at a depth of 1000 m (C = 3000) enclose at all points.

[*Dixon et al.*, 1989]. However, since simple shear extension is always less effective than pure shear in inducing partial melting [*Buck et al.*, 1988], a relatively steep detachment is required. A similar asymmetry may originate if the pre-LIP lithosphere contains a thin spot which laterally offsets the maximum melt zone and surface location of the LIP [*Thompson and Gibson*, 1991; *Skogseid et al.*, 1992a].

Discussion

Various LIPs are associated with continental and oceanic breakup, but causal mechanisms have yet to be well documented. The active plume head, steady state plume, or hot mantle models could account for all LIPs; the secondary convection model could apply exclusively to some volcanic passive margins. Mantle tomography suggests, however, that subduction, not plume upwelling, is the dominant upper mantle process today [e.g., *Dziewonski and Woodhouse*, 1987; *Anderson et al.*, 1992a], which suggests that at least some plumes originate in the lower mantle. A lower mantle origin is also supported by the huge calculated melt volumes for the largest igneous provinces; however, most LIPs might originate equally well in the upper mantle (Figure 10). Furthermore, longevity of hotspot sources is hard to reconcile with an exclusive plume model in a convecting mantle: Why, for example, should the thin plume stem be maintained over 100 m.y. or more [e.g., *Richards and Griffiths*, 1988]? And how do the different scales of convection interact?

Also problematic in a convecting upper mantle is the apparent fixity of hotspots within two distinct groups, an Atlantic/Indian and a Pacific group, over the past 90 m.y. [*Müller et al.*, 1993].

One might envision a convecting upper mantle occasionally being penetrated by a plume originating deep in the mantle (Figure 13). This would be consistent with a heterogeneous upper asthenosphere, as documented by the various mantle end-members [e.g., *Hart et al.*, 1992], in which temperature, composition, and fluid content vary regionally. These factors, together with the history and rate of lithospheric extension prior to continental separation, would then determine the amount, timing, and position of igneous rocks emplaced during breakup. A heterogeneous asthenosphere would produce the observed excess magma at and in the vicinity of the crustal focus of deep mantle plumes. On the other hand, it would also allow formation of volcanic passive margins by increased melt production away from a plume if rifting occurs within a region of "abnormal" asthenosphere. On the basis of numerical models, *Tackley et al.* [1993] have suggested heterogeneous mantle circulation in which cold downwelling plumes and hot upwelling plumes, but not linear forms (sheets), penetrate the 670-km discontinuity, accounting for long-wavelength lateral heterogeneity in the mantle. Nevertheless, additional observational data and modeling are clearly required to adequately address the origin and evolution of LIPs
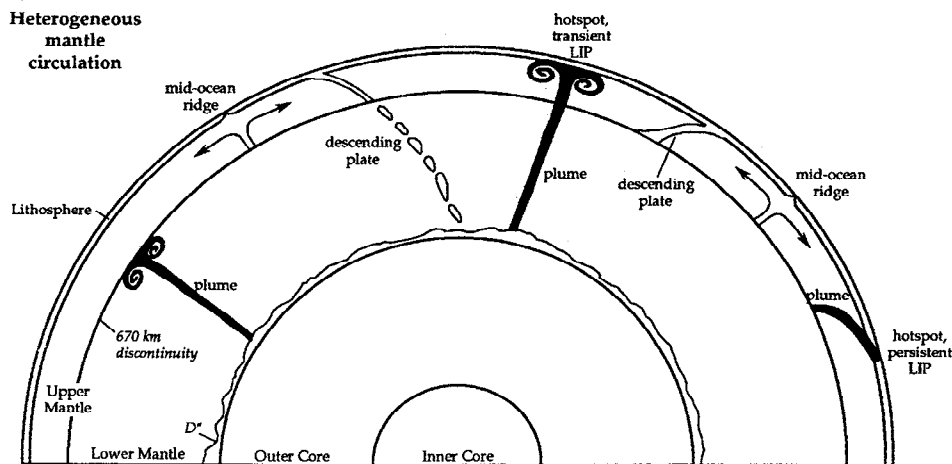
Figure 13. Cartoon of heterogeneous mantle circulation proposed in this paper.

and their relationship to continental and oceanic breakup.

Although available data may be interpreted both in terms of "active" and "passive" asthenospheric upwelling, melt volumes and geological settings of many LIPs suggest a combination of the two geodynamic models by an active thermal source impinging on lithosphere under extension. Therefore different models for generating LIPs are not necessary mutually exclusive, and components of them may contribute to the observed tectonomagmatic emplacement history.

5. ENVIRONMENTAL IMPLICATIONS

Construction of transient LIPs has the potential to induce environmental change and affect biotic evolution and extinction primarily because of (1) geometrical changes of the Earth's surface, in particular, changing of the ocean basin geometry, (2) chemical and physical changes of the hydrosphere caused by interaction of lava and seawater, and (3) increased transfer of gases and particulate matter to the atmosphere during eruptions (Figure 14). Consequences of LIP emplacement on basin geometry and on the biosphere may occur on local, regional, and global scales

[Lockley, 1990]. Moreover, the fact that many oceanic plateaus are isolated from continental detritus and usually lie above the surrounding seafloor and well above the calcium carbonate compensation depth makes them repositories of sediment deposited at varying paleodepths, commonly without major hiatuses. Thus LIPs which are "active" in forming or changing the environment also provide an opportunity to study environmental change by serving as "passive" sediment repositories [Coffin and Eldholm, 1991].

The intense debate on extraterrestrial impact versus endogenous mechanisms for changes at the Cretaceous-Tertiary (K-T) boundary [e.g., Lockley and Rice, 1990; Sharpton and Ward, 1990] has to some extent overshadowed the important message: the temporal proximity of LIP emplacement and environmental change (e.g., Figures 2 and 15). Durations of these mechanisms are distinctly different, and also, it is difficult to clearly determine the cause and effect relationships of tectonomagmatic events and global environmental events. LIP formation, in particular, is associated with many individual factors of different duration and polarity that alone or combined, impact on the environment; the geologic record reflects the sum of these factors and provides few clues for differ-
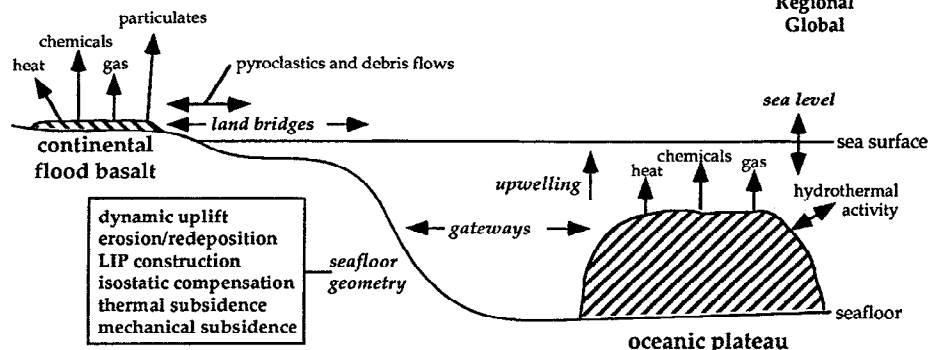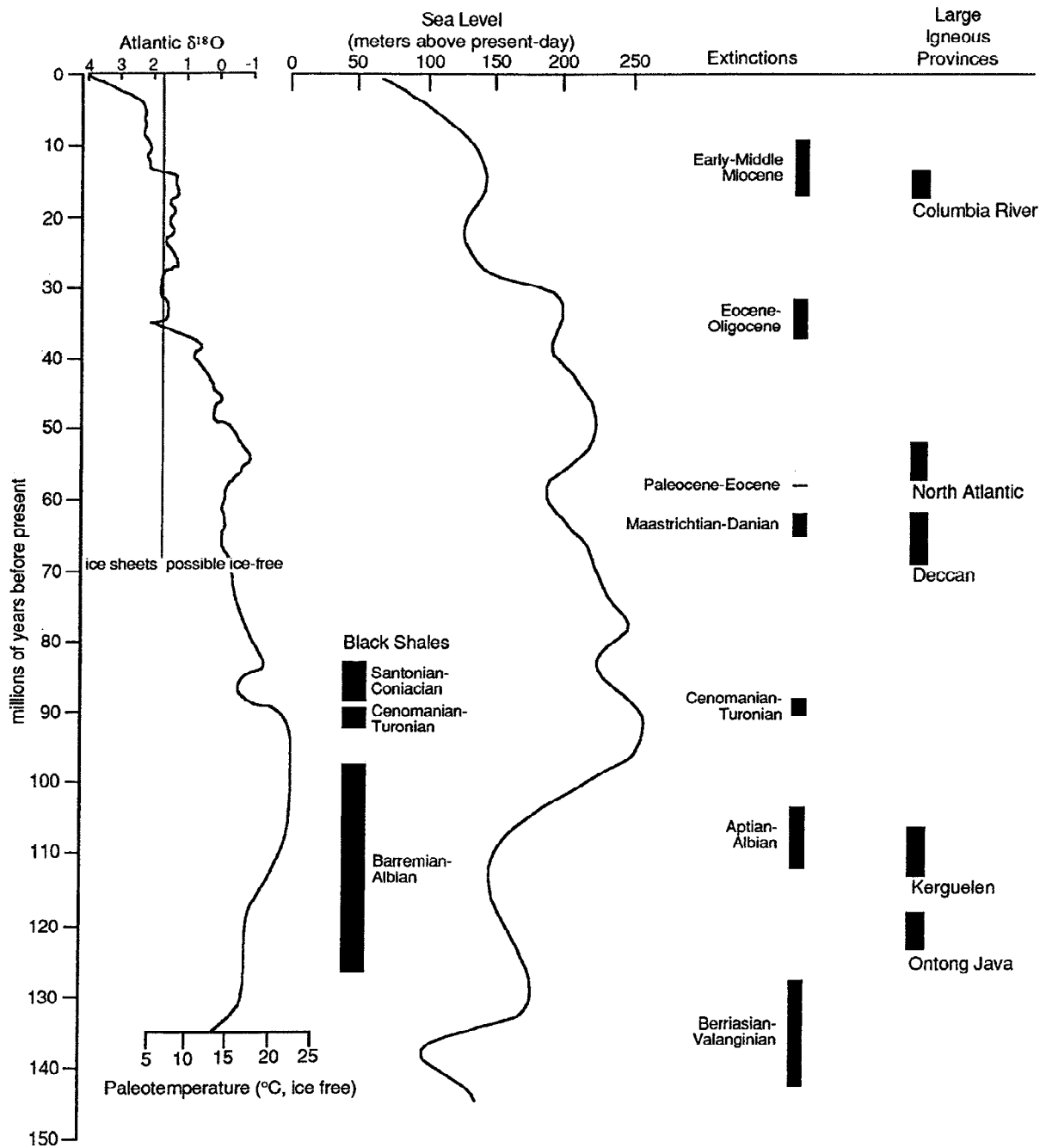

Figure 14. Cartoon of environmental parameters illustrated by emplacement of transient LIP end-members continental flood basalts and oceanic plateaus. All parameters are relevant for volcanic margins; absolute and relative impact depends on tectonomagmatic setting and magmatic intensity.

Figure 15. Cenozoic and Cretaceous climate, black shales, sea level, mass extinctions, and five LIP emplacements plotted on the timescale of *Harland et al.* [1990]. Atlantic $\delta^{18}O$ is after *Raymo and Ruddiman* [1992], paleotemperature is after *Arthur et al.* [1985], black shales are after *Jenkyns* [1980], sea level is after *Haq et al.* [1987], extinctions are after *Rampino and Stothers* [1988] and *Thomas* [1992], LIP emplacements are from *Coffin and Eldholm* [1993] and this paper.

entiation. Therefore we focus on temporal and spatial similarities without excluding other causes either separated in time or coeval with LIP formation and draw on examples from the three major LIP categories.

Temporal Correlations

Environmental impact of LIP emplacement is supported by temporal correlations of continental flood basalt volcanism and periods of major biotic changes or extinctions of terrestrial and marine organisms [e.g., *Rich et al.*, 1986; *Officer et al.*, 1987; *Rampino* and *Stothers*, 1988; *White*, 1989a; *Rampino and Caldeira*, 1993] (Figure 2). Although most such correlations are based on few and relatively uncertain age determinations, a general relationship is apparent. Three spectacular events are the synchronism of the greatest known extinction of biota on Earth at the Permian-Triassic boundary (~248 Ma) and the Siberian traps [*Campbell et al.*, 1992] and the biotic changes at the K-T [e.g., *Officer et al.*, 1987] and Paleocene-Eocene [e.g., *Rea et al.*, 1990; *Thomas*, 1990] transitions during eruptions of the Deccan flood

basalts and the North Atlantic volcanic province, respectively.

The geological record shows changes in populations of marine and terrestrial biota at or near the K-T transition [e.g., *Raup and Sepkoski*, 1986]. Extinctions were particularly severe among calcareous planktonic organisms and among many species of reptiles. In contrast, some biotic groups show no significant changes, for example, benthic organisms. However, ~10 m.y. later, within chron 24r, many benthic foraminifera experienced global mass extinction [*Kennett and Stott*, 1991; *Thomas*, 1992]. In fact, there is a compelling temporal correspondence between emplacement of the North Atlantic volcanic province and major paleoceanographic changes comprising this "Paleocene-Eocene boundary event." Among the most significant events are massive extinctions of deep-sea organisms and land mammals, a large shift in oceanic carbon isotopes, greatly increased hydrothermal activity, and reduced atmospheric circulation [*Rea et al.*, 1990; *Thomas*, 1992]. Increased ocean temperature corresponds with Paleocene-Eocene warming (Figure 15) and with the early Eocene climate maximum which marks the warmest period on Earth during the last 70 m.y. [*Owen and Rea*, 1985]. As another example of temporal correspondence, the peak eruptive phase of the Columbia River basalts correlates with early and middle Miocene extinctions [*Rampino and Stothers*, 1988].

No similar temporal correlations have been well documented for emplacement periods of the giant oceanic plateaus. *Tarduno et al.* [1991], in fact, note that the differences between biotic response to interpreted submarine Ontong Java Plateau volcanism and subaerial volcanism at the K-T transition are much greater than the similarities, suggesting that emplacement of the largest igneous provinces may occur without a major negative global impact on the marine ecosystem. Diversification of nannoplankton after emplacement of the Ontong Java Plateau may represent, in fact, a positive impact on the marine ecosystem [*Erba and Larson*, 1991]. The igneous event would cause atmospheric CO_2 levels to rise, increasing the global mean temperature. Sea level would also rise, mobilizing nutrients from the shelf. Both of these factors would promote biologic evolution. In contrast, *Raup and Sepkoski* [1986] report an extinction event in the Aptian which correlates with emplacement of the Kerguelen Plateau and Broken Ridge.

LIP Properties

Great variability in LIP magnitude and emplacement rates (Figures 2, 6, and 10 and Table 3) does not have to translate directly to a corresponding scale of environmental impact; such a direct correlation with size may, in fact, be misleading. Admittedly, the largest igneous province represents greatly increased crustal production (Figure 2) and associated uplift,

factors that influence basin geometry and eustatic sea level [*Pitman*, 1978]. However, other significant factors or combinations of factors may not be directly related to LIP size [*Self and Rampino*, 1988]. For example, geological setting governs interaction between lavas and the biosphere, as well as the impact of volcanism on atmospheric and oceanographic circulation patterns. For the latter, we consider continental flood basalts and intraplate oceanic plateaus to be end-members, while the impact of the intermediate volcanic margin members depends on the magnitude of the mantle thermal anomaly, mode of rifting, plate tectonic setting, and explosivity of the volcanism. In particular, subaerial flood basalt activity may have had a larger effect than submarine eruptions because of direct input of climatically important volatiles into the atmosphere instead of into the oceans.

Physical and chemical properties of the lava comprise another key factor. Although the major proportion of extrusive lava at LIPs has a basaltic composition, volumetrically smaller acidic and intermediate varieties contribute greater amounts of gases and particles that affect biota, atmospheric chemistry, and insolation. On the other hand, lava composition also governs flow properties, and basalts in particular are able to flow for great distances. For example, flow lengths of >750 km have been reported from the Columbia River province [*Tolan et al.*, 1989], whereas acidic lavas are more viscous and remain closer to feeders.

Volume and rate of emplacement of extrusive cover may be equally as important in affecting the environment as crustal volume and total crustal production rate. *Eldholm and Grue* [1993] have estimated extrusive parameters for the North Atlantic volcanic province and noted that volumes and geometries of extrusive cover on the two giant oceanic plateaus, Ontong Java and Kerguelen, remain virtually unknown. Nonetheless, by recognizing similarities in crustal velocity structure among volcanic margins and oceanic plateaus, they inferred that the upper crustal layer on these plateaus, with an average velocity of 5.5 km s^{-1} [*Hussong et al.*, 1979; *Recq et al.*, 1990], roughly corresponds to flood basalts. Assuming average thicknesses of 4.5 and 5.0 km for Ontong Java and Kerguelen, respectively, they estimated extrusive volumes and eruption rates (Table 3). Although these values are still considered uncertain, eruption rates are of the same order of magnitude for the three types of LIPs studied and the giant oceanic plateaus are characterized by maximum average rates. We emphasize that the rates in Table 3 are averaged over large areas and time periods, whereas the actual eruption history probably consisted of many intense, voluminous, discrete events separated by quiet intervals [*Eldholm et al.*, 1989; *White*, 1989b]. Regardless, these calculations and the variable biotic response to LIP emplace-

ment show that LIPs smaller than the giant oceanic plateaus must be considered.

Although possible effects on global environment may directly result from LIP "forcing," it is likely that the impact of individual LIP events is a function of the state of the global environment during emplacement. In periods of environmental stress close to ill-defined threshold levels, LIP genesis might trigger nonlinear responses and rapid global change.

Basin Geometry

Rapid emplacement of offshore LIPs changes ocean basin geometry by the morphology and location of the new constructional features, which in turn displaces seawater and thus modifies sea level. These changes can directly influence water mass formation and circulation, and erosion and sedimentation, influences that might reach far beyond the LIP proper.

Initial thermal and dynamic uplift and subsequent rapid build-up of the LIP above normal oceanic crustal depths result in a eustatic sea level rise. Such rises might be detected on continental margins. Rate and magnitude of the rise depends on the volume and rate of LIP emplacement and on isostatic and flexural responses of the lithosphere. The volume of water-displacing igneous rock on the Ontong Java Plateau, for example, is $\sim 4 \times 10^6$ km^3 [Schubert and Sandwell, 1989], which would elevate sea level by ~ 10 m. If the entire Ontong Java LIP is included, this rise is larger. The magnitudes, however, are much smaller than those calculated by Haq et al. [1987] (lower Aptian black shale deposition, for example, corresponds to a short-term eustatic rise of ~ 75 m), but they are comparable to those of Pitman and Golovchenko [1991]. Because fluctuations in global crustal accretion (seafloor spreading) rates are more gradual (Figure 2) than LIP emplacement (Table 3), we consider the latter more likely to produce "events" in the stratigraphical record on continental margins. When the transient igneous episode ceases, LIP subsidence contributes to eustatic sea level fall. Thus the net effect of a LIP emplacement would be a gentle eustatic sea level rise (lithospheric uplift) culminating with a rapid flooding event (LIP emplacement) and a subsequent gentle eustatic sea level fall (LIP subsidence).

Oceanic plateaus and volcanic margins affect water mass circulation by creating physical barriers. The barriers sometimes function as gatekeepers between major ocean basins which over short time intervals might inhibit existing or create new water mass routes. The significance of such gatekeepers, which significantly affect the global transport system of heat and moisture, depends on their location, areal extent, and depth profile. The marine environment responds rapidly to physical oceanographic events of this kind, and establishment of gateways is well documented in the sedimentary record [Hay, 1988]. Previously, ocean basin evolution and fragmentation, and thereby gate-

way development, have been related to plate tectonic events and the subsidence history of persistent LIPs, or hotspot trails (e.g., the Greenland-Scotland ridge, Walvis Ridge-Rio Grande Rise). Formation of the North Atlantic volcanic province (Figure 7), however, introduced an additional component of basin fragmentation associated with synrift uplift and buildup of extrusive complexes which significantly influenced regional circulation and sedimentation and may have delayed opening of gateways as well as contributed to biotic endemism [Eldholm, 1990].

Compared with a "standard" plate tectonic model of ocean basin development characterized by initial subsidence and basin widening and deepening with time, oceanic LIP formation, either at plate boundaries or intraplate, represents an additional event of basin deformation which might change water mass circulation patterns both on regional and global scales. Although LIP emplacement normally will affect both surface and deep waters, the impact on surface water is most significant during the constructional phase, while the impact on bottom-water circulation might last for longer periods. In addition, the surface waters above some LIPs may be affected by topographic upwelling which may enhance productivity.

Impact on the Biosphere

Submarine volcanism might contribute to biotic change, including mass extinction events, by addition of trace metals such as As, Sb, and Se, which are poisonous to marine life. Through heat output, undersea volcanism also destabilizes the water column, inducing large-scale overturn affecting surface-dwelling organisms. Furthermore, release of volatiles such as CO_2 and their redistribution through the ocean might lead to changes in the alkalinity of the ocean affecting climate and marine life. Increased CO_2, in particular, may cause dissolution of carbonate. This would augment high-productivity episodes recorded as black shales during oceanic anoxic events [Schlanger and Jenkyns, 1976]. Pacific Ocean sediment records intervals of major carbon burial at 91, 112, and 117.5 Ma, which correspond to patterns of biotic evolution [Sliter, 1992]. The two oldest events are approximately contemporaneous with formation of the Ontong Java and Kerguelen LIPs, respectively (Figures 2 and 15).

CO_2 content in the oceans and atmosphere may change in response to seafloor hydrothermal activity [Berner et al., 1983] and volcanism [Arthur et al., 1985]. Noting that hydrothermal activity is most intense during continental rifting and oceanic plate boundary rearrangement, Owen and Rea [1985] and Rea et al. [1990] postulated that the geochemical impact of these tectonic events could affect the biosphere. Hydrothermal activity along the present midocean ridge system accounts for $\sim 20\%$ of the preindustrial atmospheric CO_2 level, and enhanced hydrothermal activity might significantly raise global

CO_2 levels, potentially inducing global warming [Owen and Rea, 1985]. On the other hand, Varekamp et al. [1992] infer that the combined effects of mid-ocean ridge volcanism, subduction magmatism, and slab alteration have little impact on atmospheric CO_2 content and climate, whereas intense flood basalt volcanism may influence climate on timescales in excess of 10^4 years. Thus transient volcanism may force rapid environmental change.

Continental flood basalts and the subaerial part of oceanic LIPs transfer CO_2 directly to the atmosphere [McLean, 1985; Rampino, 1991]. For periods >1 m.y., atmospheric CO_2 levels are governed mainly by magmatic sources and chemical weathering [Raymo and Ruddiman, 1992]. The Deccan traps, for example, could have transferred as much as 2×10^{17} mol of CO_2 over a period of several hundred thousand years, a volume which has been estimated to create a global greenhouse effect of $\sim 2°C$ [Caldeira and Rampino, 1990]. Coeval volcanism along the conjugate passive margins of western India and the Seychelles-Mascarene Plateau might have greatly enhanced this effect. Furthermore, Eldholm and Thomas [1993] have suggested that CO_2 output of North Atlantic volcanic province basalts could have triggered rapid environmental change through increased high-latitude surface temperatures, in turn changing deep-sea circulation and productivity. The Early Cretaceous superplume [Larson, 1991b] provides another source of potential large-scale atmospheric impact. It may have led to CO_2 levels from 3.7 to 14.7 times the modern preindustrial value, resulting in a global warming of 2.8°–7.7°C with respect to the present global mean temperature. Including changes in paleogeography and higher sea level, the temperature range increases to 7.6°–12.5°C [Caldeira and Rampino, 1991]. It is probably not coincidental that the Cretaceous greenhouse culminated from late Aptian through late Albian time [Arthur et al., 1991] following Ontong Java LIP formation and that the early Eocene greenhouse followed the North Atlantic volcanic margin LIP emplacement.

Volcanic eruptions also transfer other volcanic gases, aerosols, ash, and heat to the atmosphere [Sigurdsson, 1990]. Aerosols are commonly generated from violent, explosive silicic eruptions which transport sulfuric volatiles to the stratosphere where they are converted to sulfuric acid aerosols [Rampino and Self, 1984; Self and Rampino, 1988]. Furthermore, widely dispersed early Tertiary tholeiitic tephras in the North Atlantic realm [Knox and Morton, 1988] imply that the North Atlantic volcanic province had at least a regionally significant environmental impact [Eldholm and Thomas, 1993]. In addition, basalts drilled at ODP site 642 reflect phreatomagmatic eruptions [Viereck et al., 1988] where local magma-water interaction amplified the explosive nature. Sulfur contribution per unit volume from basaltic eruptions like those at LIPs is about 1 order of magnitude greater than that from

silicic magma [Devine et al., 1984]. Thus the gases and particles from both central vent and fissure eruptions might, if carried to the stratosphere, cause dramatic short-term effects such as acid rain, darkening, and climatic cooling spells [Officer et al., 1987]. Stothers et al. [1986] proposed that convective plumes over vents with very high eruption rates and associated large fire fountains provide the injection mechanism and calculated that the larger Columbia River flow units were able to trigger environmental effects of this kind.

Synrift uplift at some volcanic margins and continental flood basalts is another tectonic mechanism that might influence environmental parameters by modifying atmospheric circulation [Ruddiman and Kutzbach, 1989] and by increasing chemical weathering, resulting in withdrawal of CO_2 and global cooling [Raymo, 1991].

Resources

Relations between LIP emplacement and regional and global environmental events may have implications for resource generation and accumulation [Keith, 1982]. Larson [1991a] pointed out the coincidence of Early Cretaceous superplume formation, global temperature increase, black shale deposition, eustatically transgressive sea level, and oil generation. The resulting flooded continental platforms would promote marine deposition of organic carbon (phytoplankton) that would later mature and form oil. He also explained the large Pennsylvanian-Permian coal deposits by a similar superplume coinciding with sea level fall.

On volcanic margins, several factors may relate to hydrocarbon potential [Eldholm, 1991; Skogseid et al., 1992b]. First, LIP emplacement affects prerift sedimentary basins on the incipient margin by regional uplift, erosion, tectonism, and emplacement of intrusive and extrusive rock complexes. Second, the thermal anomaly responsible for LIP formation might affect maturation. Third, regional synrift uplift provides an additional source area for synrift and postrift sediment. If basin fragmentation during rifting and initial continental margin development forms restricted and poorly ventilated basins during hot and humid climates thereby enhancing biological productivity, local potential lacustrine or marine source rocks may develop. This setting existed during breakup of the North Atlantic, and a similar setting should be considered for anoxic sequences along other passive continental margins, such as the Early Cretaceous South Atlantic.

6. CONCLUSIONS

Evidence that LIPs both manifest a fundamental mode of mantle circulation commonly distinct from that which characterizes plate tectonics and contribute episodically, sometimes catastrophically, to global environmental change is accumulating rapidly. Dating of

many continental flood basalts, from the Deccan traps in particular, have convincingly shown the geological suddenness of LIP emplacement. Geophysical and drilling data have demonstrated in one case, the North Atlantic volcanic province, that a large volumetric percentage of those continental flood basalts associated with continental breakup lie offshore, that the uppermost extrusives were erupted subaerially, and that the intrusive component of LIPs is at least as voluminous as the extrusive component. Geophysical and drilling data from two oceanic plateaus, Ontong Java and Kerguelen-Broken Ridge, have illustrated that volumetrically these LIPs dwarf all others known, that much of Kerguelen's uppermost crust was erupted subaerially, and that extrusives of the Ontong Java LIP affected ~1% of the Earth's surface. However, this study clearly documents that the database, particularly with respect to deep crustal structure and drillholes providing reliable age, compositional, dimensional, and environmental data with which to formulate and constrain geological models, is significantly smaller than for most other comparable onshore and offshore features. It is important to recognize that to date, we have literally only scratched the surface of offshore LIPs. Nevertheless, we believe that the relative scale of LIPs and the large dimensions of some LIPs are real, whereas absolute values will undoubtedly change with new data.

Our dimensional analysis, combined with seismological, geochronologic, geochemical, and isotopic data and results, leads us to favor a complex model of mantle circulation. Plumes responsible for the largest igneous provinces likely originate from the D″ layer at the base of the mantle. Smaller plumes may well originate in the transition zone between the lower and upper mantle. How plumes at any scale interact with primary plate tectonic mantle convection and why some relatively weak plumes should persist for 100 m.y. or more remain very much open questions. The heterogeneous thermal character of the mantle, the presence of at least four distinct mantle reservoirs, and the two fundamental modes of mantle circulation suggest a complexity beneath our feet which will occupy geoscientists' attention well into the future.

Tectonic contributions to global change are increasingly being recognized. Although LIPs, except in the case of the Deccan traps, have largely escaped attention until recently, we have attempted to explain various ways in which LIPs would influence the environment. Episodic geometrical, chemical, and physical changes accompanying LIP emplacement would potentially affect the hydrosphere and atmosphere; LIPs which form when the global environment is in or near a "threshold" condition, for example, the North Atlantic volcanic province near the Paleocene-Eocene boundary, can potentially push the planet over the threshold. In this context we view the mid-ocean ridge system as more a "regulator" and LIPs as more "instigators" of global environment change.

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