INTRODUCTION

Volcanic rifted margins (Fig. 1) are produced where continental breakup is associated with the eruption of flood volcanism during prerift and/or synrift stages of continental separation (Fig. 2) (Mutter et al., 1982; White et al., 1987; Holbrook and Kelemen, 1993; Eldholm and Grue, 1994; Courtillot et al., 1999). These margins are easily distinguished from nonvolcanic margins, like the Iberian margin, that do not contain such a large amount of extrusive and/or intrusive igneous rock and that may exhibit unusual features, such as unroofed mantle peridotites (e.g., Pickup et al., 1996; Louden and Chian, 1999). Mapping of flood basalt provinces and subsurface seismic volcanic-stratigraphic analyses show that volcanic rifted margins border the northern, central, and southern Atlantic Ocean, the southern Red Sea, the east coast of Africa, circum-Madagascar, the east and west coasts of India, the western and eastern coasts of Australia, and possibly parts of Antarctica (Coffin and Eldholm, 1992, 1994; Mahoney and Coffin, 1997; Planke et al., 2000) (Fig. 1). The initiation of a flood basalt province (or of a large igneous province [LIP]) (Fig. 2) is commonly a prerift phenomenon and takes the form of a subaerial and/or submarine seaward-dipping reflector series and a significant thickness (to 15 km) of juvenile, high-velocity lower crust seaboard of the continental rifted margin. Herein we outline the similarities and differences between volcanic rifted margins worldwide and list some of their diagnostic features.

Characteristics of volcanic rifted margins

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ABSTRACT

Volcanic rifted margins evolve by a combination of extrusive flood volcanism, intrusive magmatism, extension, uplift, and erosion. The temporal and spatial relationships between these processes are influenced by the plate tectonic regime; the preexisting lithosphere (thickness, composition, geothermal gradient); the upper mantle (temperature and character); the magma production rate; and the prevailing climatic system. Of the Atlantic rifted margins, 75% are believed to be volcanic, the cumulative expression of thermotectonic processes over 200 m.y. Volcanic rifted margins also characterize Ethiopia-Yemen, India-Australia, and Africa-Madagascar. The transition from continental flood volcanism (or formation of a large igneous province) to ocean ridge processes (mid-ocean ridge basalt) is marked by a prerift to synrift transition with formation of a subaerial and/or submarine seaward-dipping reflector series and a significant thickness (to 15 km) of juvenile, high-velocity lower crust seaboard of the continental rifted margin. Herein we outline the similarities and differences between volcanic rifted margins worldwide and list some of their diagnostic features.

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probably variable amounts of sedimentary detritus shed from the volcanic rifted margin during uplift and tectonic denudation of the kilometer-scale rift mountains. The formation of SDRS is associated with the establishment of thicker than normal oceanic crust, seaward of the rifted margin at the continent to ocean transition (Fig. 2). Eventually stretching and heating lead to effective rupture of magmatically modified continental lithosphere, and seafloor spreading commences. This early oceanic crust may be thicker than normal owing to hotter asthenosphere associated with the plume and/or steep gradients at the lithosphere-asthenosphere boundary (e.g., Boutilier and Keen, 1999) (Fig. 2). The interval between the first expression of volcanic rifted margin formation on the prerift continental margin and the formation of true ocean floor can be tens of millions of years (Fig. 3).

In this paper we focus on evidence from a few of the better-known volcanic rifted margins, Ethiopia-Yemen, the Atlantic margins, and the Australia-India conjugate margins. Miocene to recent volcanic rifted margins (<30 Ma) exist in northeastern Africa (Hoffmann et al., 1997; George et al., 1998; Ebinger and Casey, 2001) and southwestern Arabia (e.g., Baker et al., 1996a; Menzies et al., 1997a, 1997b), where the conjugate margins are separated by ocean spreading centers in the southern Red Sea. The southern Red Sea and eastern Gulf of Aden are the youngest, hottest, and most active volcanic rifted margins. They are characterized by active volcanism, high heat flow, and shallow earthquakes (e.g., Davison et al., 1994; Manighetti et al., 1998; Ebinger and Casey, 2001). On the volcanic rifted margins of the southern Red Sea one can observe the temporal transition from prerift flood volcanism to synrift domino fault-block terranes (Yemen) and subaerial seaward-dipping reflector series (Ethiopia) (Fig. 1). Cretaceous-Tertiary volcanic rifted margins occur in the North Atlantic (i.e., western Greenland, Norway, and the United Kingdom) (e.g., Larsen and Jakobsdottir, 1988; Larsen and Saunders, 1998; Saunders et al., 1997; Klausen and Larsen, this volume), around peninsular India, and western Australia (e.g., Kent et al., 1997) (Fig. 1). In Brazil and Namibia, volcanic rifted margins are related to the opening of the South Atlantic (Peate, 1997), beginning with Parana and Etendeka flood volcanism (Hawkesworth et al., 1992; Renne et al., 1992, 1996b; Peate, 1997; Mohriak et al., this volume; Corner et al., this volume). One of the spatially most extensive volcanic rifted
margins, the Central Atlantic magmatic province, formed ca. 200 Ma (Holbrook and Kelemen, 1993; McHone, 1996; Hames et al., 2000) and records the initial breakup of Pangea. This volcanic rifted margin has been reduced to erosional remnants of intrusive and/or extrusive complexes (McHone, 1996) and offshore SDRS (Benson, 2002) spread over 10^6 km^2 (Fig. 1). Older intraplate flood basalt provinces erupted in the Permian-Triassic occur in Siberia, Russia, and Emeishan, China. However, their relationship to rifted margins is unknown, and they are not discussed herein. In addition, we do not consider relics of Precambrian flood basalt provinces that are apparent as dike swarms and unroofed plutonic complexes, because links to continental breakup are even more elusive (Mahoney and Coffin, 1997, and references therein).

We use these better known volcanic rifted margins as we discuss some of the controversies surrounding the origin of volcanic rifted margins (whether associated with active or passive rifting) and describe some of their diagnostic features, i.e., common association of silicic volcanism with the dominant flood basalts; crustal architecture of HVLC and SDRS at the continental-ocean transition; temporal relation between extension and magmatism; and rift-margin uplift and mountain building.

**Passive continental rifted margins: Plate-driven and plume-driven processes**

Traditionally, passive (plate driven) and active (plume driven) rifting models were invoked as an explanation for the formation of nonvolcanic and volcanic passive margins, respectively. Passive or plate-driven rifting models required that continental breakup was initiated by extensional forces, followed by surface uplift and magmatism related to the passive upwelling of normal asthenospheric material. In this case melts would be generated by shallow decompression melting processes. In contrast, active or plume-driven rifting models required deeper melt generation and subsequent interaction with the continental lithosphere. In these models the expectation is that volcanic rifted margins formed by kilometer-scale surface uplift prior to LIP formation and extension. However, recent research on volcanic rifted margins indicates that such simple chronologies do not apply to many rifted margins, suggesting that their formation is also not simple. The timing of uplift and extension relative to LIP formation is complex, requiring more detailed observations in individual provinces (Table 1).

Lithospheric thinning is an uncontested requirement for volcanic rifted margin formation. Extensional forces large enough to initiate rifting are generated by the presence of hot, low-density asthenosphere and subsequent heating of mantle lithosphere (Crough, 1978). More controversial is the mechanism responsible for the production of large volumes of basaltic volcanism at the Earth’s surface, which in the majority of cases is spatially and temporally related to continental breakup (Watkeys, this volume). Adiabatic decompression melting due to active upwelling of normal asthenosphere is triggered by lithospheric thinning, as observed at ocean ridges. This occurs without a thermal anomaly and/or plume and may have led to considerable melt production on rifted margins (e.g., Holbrook and Kelemen, 1993; Boutilier and Keen, 1999; Korenaga et al.,...
2000). Alternatively, thermal anomalies or plume processes are believed to be a vital prerequisite for the generation of large volumes of melt (e.g., White and McKenzie, 1989; Sleep, 1996; Ernst and Buchan, 1997). In such cases the mantle potential temperature is elevated above that of the normal asthenosphere (~1300°C). However, there is considerable controversy over whether plumes initiate rifting, or rifting focuses plume activity (e.g., King and Anderson, 1998; White and McKenzie, 1989; Ebinger and Sleep, 1998; Nyblade, this volume). Controversy also continues with regard to the geometry of plumes; their temperature; their depth of origin; and their chemical identity (e.g., Turcotte and Emerman, 1983; Richards et al., 1989).

Geophysical and geochemical evidence, claimed as proof for the origin of plumes in the deep or shallow mantle, is equivocal. Mantle tomography models show distinct low-velocity zones at the core-mantle boundary, but their continuity with upper mantle low-velocity zones may be ambiguous. Ray dispersion complicates the simultaneous resolution of the width and depth of particular features in the upper mantle (Shen et al., 1998). However, as more and more receiver-function studies are undertaken in plume provinces, an exciting new plume detection method has evolved that allows direct measurement of the 410 km and 670 km discontinuities. As a result, the locations of plume stems can be mapped (Wolfe et al., 1997; Shen et al., 1998). For petrologists and geochemists the controversy surrounds the identification of volcanic rocks with unequivocal primary mantle signatures. Low magnesian volcanic rocks that have undergone low-pressure fractionalation are obviously inappropriate probes of high-pressure mantle processes. Even highly magnesian unfractiated volcanic rocks can inherit the chemical signature of the lithosphere (crust and mantle) because of their higher temperature. However, advances have been made toward the identification of geochemical criteria that may help resolve mantle and/or crustal characteristics (e.g., Baker et al., 2000; Breddam et al., 2000; Melluso et al., this volume; Baker et al., this volume).

What may be stated, with some certainty, is that large-scale thermal anomalies, plumes, and/or hotspots exist in the Earth, and their distribution, temporal evolution, and spatial extent are highly variable. Variations may arise because of preplume lithospheric structure and differences in the velocity of plates over plumes. Slow moving plates may show larger volumes of melting than fast moving plates. In addition, their longevity and genesis may manifest as uplift, subsidence, and extension of the lithosphere and associated magmatism, occurring to different degrees and in various sequences. In volcanic rifted margins melts are produced by variations in pressure and temperature, and these can be achieved, respectively, by lithospheric thinning and thermal anomalies.

Figure 3. Formation of flood basalt provinces (large igneous provinces [LIPs]) on volcanic rifted margins and timing of formation of oceanic crust. Subaerial or submarine seaward-dipping reflector series (SDRS) characterize prerift to synrift (continent to ocean) transition and predate oldest oceanic crust on that volcanic rifted margin. Note that in most instances flood volcanism or LIP formation precedes breakup and formation of ocean crust. High-velocity lower crust and SDRS tend to form after the main flood basalt episode and before the youngest oceanic crust. See text and Table 1 for references.
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<th>Table 1: Characteristics of Volcanic Rifted Margins</th>
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<td><strong>LIP:</strong> Present-day thickness of sub-aerial volcanic rocks</td>
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Note: **LIP** large igneous province; **SDRS** seaward dipping reflector series; **HVL:C** high velocity lower crust.

Examples of source references for specific volcanic rifted margins: Ethiopia and Yemen: Berckhemer et al. (1975); Davison et al. (1994); Baker et al. (1996 a,b); Menzies et al. (1997 a,b); Egloff et al. (1997); Al’Subbary et al. (1998); George et al. (1998); Hoffmann et al. (1997); Baker et al. (2000); Ebinger and Casey (2001); Ukstins et al. (2002); Baker et al. (this volume). Greenland and UK: Roberts et al. (1979); Mutter et al. (1982); White et al. (1982); White and MacKenzie (1989); Brodie and White (1994); Saunders et al. (1997); Larsen and Saunders (1998); Jolley (1997); Korenaga et al. (2000); Planke et al. (2000); Klausen et al. (this volume). India and Australia: Von Stackeberg et al. (1980); Von Rad and Thurov (1992); Storey et al. (1992); Colewell et al. (1994); Exxon and Colewell (1994); Milner et al. (1995); Frey et al. (1996); Kent et al. (1997). Parana and Etendeka: Hawkesworth et al. (1992); Renne et al. (1992); Gallagher et al. (1994); Turner et al. (1994); Renne et al. (1996a, 1996b, 1998); Gladczuk et al. (1997); Peats (1997); Gerson et al. (1999); Davison (1999); Jeram et al. (1998); Stewart et al. (1999); Hinz et al. (2000) and refs therein; Brauer et al. (2000); Corner et al. (this volume); Moiriasik et al. (this volume); Trumbull et al. (this volume); Watkeys et al. (this volume); Central Atlantic Magmatic Province: McBride (1991); Holbrook and Kelemen (1993); McHone (1996); Lizaralde and Holbrook (1997); Withjack et al. (1998); Hanes et al. (2000); Benson (2001); McHone and Puffer (2001); Schlichter et al. (2001).
LIP continental basaltic and silicic flood volcanism: Shallow and deep sources

The birth of volcanic rifted margins (Table 1) is associated with the subaerial eruption of basaltic rocks and the minor eruption of submarine pillow lavas (e.g., Jolley, 1997; Planke et al., 2000) (Fig. 2). Whereas basaltic volcanism normally dominated the evolution of the LIP, silicic volcanism may have contributed significantly to the total volume of the volcanic pile (e.g., Peate, 1997; Bryan et al., this volume; Jerram, this volume). LIPS, which characterize all volcanic rifted margins, are rarely thicker than 2 km (Table 1) because they represent the erosional remnants of earlier sequences estimated to have been as much as 2–3 times as thick at the time of eruption (Table 1) (Cox, 1980; Mahoney and Coffin, 1997). These estimates of the original erupted thickness on the continental margin take into account the amount of subaerial volcanism that has been eroded by synrift or postrift processes. However, the erupted thickness differs from the actual melt thickness produced during volcanic rifted margin formation, which must include igneous intrusives added to the continental crust as dike-sill complexes and plutonic centers. Today these may be evident as unroofed magma chambers extending the length of rifted margins (e.g., Namibia, Scotland), exposed dike swarms (e.g., Saudi Arabia), or overthickened HVLC. HVLC is never exposed at the surface, but is frequently reported from seismic data across volcanic rifted margins.

Many geochronological methods have been applied to volcanic rifted margins (e.g., Rb-Sr, K-Ar, Ar-Ar), but major advances in argon-argon dating using K-rich phenocryst phases (e.g., sanidine, amphiboles) and lasers have led to an improved understanding of the genesis of silicic and basaltic volcanic rocks in volcanic rifted margins (Renne et al., 1992, 1996a, 1996b; Turner et al., 1994; Hames et al., 2000; Ukstins et al., 2002; Miggins et al., this volume). There is considerable debate, however, about the age of individual provinces (see Peate, 1997, for review). In the majority of volcanic rifted margins, dating indicates that the main pulse (i.e., 70%–80%) of subaerial continental margin volcanism, both basaltic and silicic, occurred over a relatively short period of time ranging from 1 to 4 m.y. (Table 1).

In some volcanic rifted margins, basaltic volcanic rocks are dominant (e.g., Central Atlantic magmatic province, Greenland), while in others silicic volcanic rocks can constitute a significant part of the volcanic stratigraphy (e.g., northeastern Africa, South America, Africa) (Table 1). Silicic volcanism can occur early during the main basaltic episode or after the main basaltic eruptions (e.g., Ethiopia and Parana). Extrusive silicic rocks do not exist in all volcanic rifted margins, but may occur as silicic intrusives (e.g., Greenland; Table 1). The coexistence of basaltic and silicic volcanic rocks or the eventual switch from basaltic to silicic volcanism reveals the complexity of magmatic processes within volcanic rifted margins. Overall the complex relationships vary from basalt-dominated volcanic rifted margins, bimodal basalt-rhyolite volcanic rifted margins, to intermixed basalt-rhyolite volcanic rifted margins (Table 1). Crustal magma chambers play a pivotal role in the formation of silicic magmas, as does melting of the lower crust, perhaps fueled by basaltic underplating (Cox, 1980, 1988). In Yemen, the silicic volcanism that postdated basaltic volcanism and lasted 3 m.y. (Baker et al., 1996a) is believed to have originated by processes of assimilation and fractional crystallization of mantle-derived melts. Individual silicic volcanic units can be geochemically linked to nearby intrusive centers, often unroofed as granite-syenite-gabbro complexes (e.g., UK Atlantic margin, Yemen). These are presumed to have acted as source regions for the silicic volcanic rocks. In contrast, the origin of silicic volcanic rocks from Etendeka-Paraná erupted during the lifetime of the flood basalt province (Peate, 1997) may relate more to the formation of large-scale crustal melts. While several igneous complexes in Namibia have been identified as sources for the volcanic rocks on the basis of similar ages, the extensive synrift lava cover in many other examples may hide the identity of associated plutonic complexes. In other volcanic rifted margins (e.g., Greenland, Paraná, Yemen) the presence of a monotonous basalt stratigraphy on the rifted margin, and a paucity of plutonic rocks, may indicate that the plutonic rocks are offshore (e.g., Deccan), or are preserved on the conjugate margin (e.g., Yemen). It is possible that the basalt stratigraphy that dominates the volcanic rifted margins in Brazil, Ethiopia, and Greenland was inextricably linked to igneous centers now preserved in their conjugate margins, Namibia, Yemen, and Scotland, respectively.

Along the youthful northeastern African margins, silicic volcanic rocks were explosively erupted, typically venting $10^2$–$10^3$ km$^3$ of magma (Ukstins et al., 2002). In the Deccan Traps and the North Atlantic Tertiary volcanic province, the presence of ash layers in the volcanic stratigraphy may indicate silicic volcanism (Deccan) or alkaline volcanism (Greenland) between periods of basaltic volcanism (e.g., Heister et al., 2001). In other volcanic rifted margins (e.g., Etendeka) (Peate, 1997), individual silicic eruptive units have thicknesses of ca. 100 m, aerial extents $>$8000 km$^2$, and volumes of 3000 km$^3$. These silicic units are comparable in volume to individual maﬁc lava units from LIPS like the Columbia River. Plinian eruption columns associated with the emplacement of voluminous ignimbrites in these volcanic rifted margins could have injected large amounts of aerosols into the atmosphere, and so affected global climate more than basaltic eruptions of similar volume.

Eruption rates in volcanic rifted margins have not been adequately deﬁned by volume-time studies of individual eruptive units, but as a ﬁrst approximation, thickness-time relationships reveal a marked decline in eruption rate from the maﬁc to the silicic eruptive stages of volcanic rifted margins (e.g., Hawkesworth et al., 1992; Baker et al., 1996a). This is consistent with the requirement for longer time periods to allow basaltic magmas to pond in shallow magma chambers and to evolve toward silicic derivatives by a combination of fractionation processes and assimilation of surrounding basement and/or roof rocks.
**Characteristics of volcanic rifted margins**

Voluminous subaerial flood volcanism on a continental margin lasting for millions of years requires a well-established magma transfer system within the crust and shallow mantle (Fig. 2). Cox (1980) first alluded to the potentially important contribution of sill-like complexes to crustal growth during flood volcanism. Shallow (i.e., caldera structures) and deeper crustal magma chambers are a requirement of many models where the mineralogy and chemistry of mafic magmas indicate fractionalization at lower crustal pressures and temperatures. The presence of plagioclase, clinopyroxene, and olivine phenocrysts in basaltic rocks alludes to fractional crystallization processes in lower crustal magma chambers, and, in many instances the geochemistry of these rocks reveals crustal contamination probably occurring concomitantly with evolution of the magmas in shallow or deep crustal chambers (e.g., Cox, 1980; Hooper, 1988). Even more extreme fractionation processes are apparent in the rhyolites found within volcanic rifted margins. Such rocks contain quartz, mica, and amphibole phenocrysts indicative of high-level processes. While some authors argue for an inextricable link underplating and basin inversion (Brodie and White, 1994), there are few reports of kilometer-scale, underplated, high-velocity layers spatially limited by several basins that could be analogues of the well-documented HVLC at volcanic rifted margins (Lizarralde and Holbrook, 1997; Korenaga et al., 2000).

Characteristic features of volcanic rifted margins are zones of HVLC (Fig. 2) between stretched continental crust and normal thickness oceanic crust (e.g., Kelemen and Holbrook, 1995; Boutilier and Keen, 1999; Korenaga et al., 2000; Benson, 2001; Trumbull et al., this volume). Most likely the HVLC was emplaced during the breakup stage or, if it was a synrift feature, was associated with mantle upwelling (e.g., Kelemen and Holbrook, 1995; Boutilier and Keen, 1999). In southeast Greenland crustal thicknesses, at equivalent positions on the continental margin, vary from 30–40 km thick close to the thermal anomaly (i.e., track of Iceland hotspot) to 18 km 500–1000 km from the anomaly (Korenaga et al., 2000) (Fig. 2). In some volcanic rifted margins, the continent-ocean transition can be abrupt (e.g., Namibia) with entirely new HVLC formed seaward of almost unchanged, perhaps slightly thinned, continental crust. In this case the generation of additional igneous material may have more to do with extension and decompression melting than plumes and/or hotspots.

Current models for volcanic rifted margins are largely based on the results of geophysical surveying and scientific drilling in the northeastern Atlantic although few deep wells are available to calibrate interpretations (e.g., Korenaga et al., 2000). Scientific drilling in the northeastern Atlantic and industry drilling off Namibia (Kudu Field) show that lavas were erupted subaerially (e.g., Mutter et al., 1982; Clemson et al., 1999). SDRS, first recognized along the North Atlantic margin, mark the synrift stage in continental breakup and as such are characteristic of volcanic rifted margins (Roberts et al., 1979; Mutter et al., 1982; White et al., 1987; Larsen and Jakobsdottir, 1988; Korenaga et al., 2000; Benson, 2001). Volcanic rifted margins have thick sequences of seaward-dipping volcanic-sedimentary strata above, or seaward of, the region of HVLC, and extending landward to the ocean-continent transition zone (e.g., Mutter et al., 1982; Clemson et al., 1999). Reflectors packages within these SDRS diverge downward and dip oceanward 20° or more (Fig. 2). Planke et al. (2000) divided these SDRS into “inner” and “outer” packages (Fig. 2) on the basis of studies of the North Atlantic margins (Fig. 1). The inner SDRS were subaerially emplaced flows, the geometry of which was affected by basin architecture. They proposed that this phase of volcanism occurred during subaerial seafloor spreading or syntectonic infilling of rift basins. The outer SDRS are believed to represent sheet flows in marine basins, and have similarities to subaerial flows. Submarine eruptions (i.e., pillow flows and hyaloclastites) characterize this developmental stage. SDRS are synrift phenomena and are distinct from flood basalts; they straddle the continent-ocean boundary and can include subaerial and submarine volcanic and sedimentary rock types.

On the Namibian margin, modeling of magnetic data from seismic profiles suggests that the SDRS is a mixture of volcanic and sedimentary rocks. Presumably some portion of the SDRS must comprise sedimentary rocks, given that the volcanic stratigraphy of the uplifted margin can be reduced in thickness during synrift erosional processes (Gallagher et al., 1994). However, whether these sediments are argillaceous or arenaceous depends on the nature of the material removed from the margin (e.g., metamorphic, sedimentary, or igneous rocks). On the Norwegian volcanic rifted margin, seismic sections have been interpreted as representing a transition from subaerial to submarine volcanic deposits that comprise lavas and volcaniclastic sedimentary rocks (Planke et al., 2000). If we take the Yemen margin as an indication of what might constitute seaward-dipping reflector series, it is clear that a significant proportion (at least 50%) of these features must be sedimentary in nature. A sediment-budget analysis of the Red Sea margin (Davison et al., 1994) in Yemen indicated that several kilometers of basaltic and/ or silicic volcanic rock were removed from the volcanic rifted margin during classic synrift extension. This erosional period would have contributed to the SDRS constructed on the stretched continental crust and embryonic oceanic crust.

Several volcanic rifted margins show an abrupt termination of the SDRS against a high-velocity structural high, which may be a late synrift intrusion (e.g., Planke et al., 2000), a fault, or an abandoned spreading ridge marking the ocean-continent boundary (e.g., Korenaga et al., 2000). Ebinger and Casey (2001) provided a mechanism for synrift emplacement of some SDRS via the development of high-strain neovolcanic zones and the abandonment of crustal detachments. Formation of SDRS on volcanic rifted margins is synchronous with the prerift to synrift transition on the continental margin. SDRS typically postdate
flood volcanism on the rifted margin, and their formation may be synchronous with a hiatus in magmatism, a change in magmatic source area, and a peak in denudation. Because this is a situation that would not be associated with the generation of melt, it is likely that strain localization and focused extension accelerated melt generation. Although SDRS may predate ocean crust formation at a mid-ocean ridge, they are transitional between rifted continental margin processes and ocean ridge processes (Fig. 2). The continent-ocean transition is difficult to determine, and therefore considerable controversy surrounds the nature of the crust beneath many SDRS. The petrology and geochemistry of both SDRS and the HVLC hold a vital clue to a major change in the source of magmas, from one that fed a LIP to one that produced oceanic crust. SDRS and HVLC are two principal diagnostics of volcanic rifted margins (Fig. 2).

**Breakup extension: Pre-LIP, syn-LIP, or post-LIP?**

The relationships between the timing of LIP formation and rifting leading to ocean-floor formation are complex. This may in part be explained by the fact that some volcanic rifted margins are proximal, others distal, to plume heads and/or stems, so it is unlikely that volcanic rifted margins will show the same relationships. It is also complicated by the possibility that magma sources for volcanic rifted margins may reside either in the deep mantle (i.e., plumes) or the shallow mantle (i.e., asthenospheric small-scale convection). The temporal relationship between magmatism and extension may differ greatly if, as we believe, in deep-sourced plumes enhanced temperatures triggered melt production, whereas asthenospheric melts are decompression melts triggered by lithospheric thinning. We envisage plume-derived magmatism occurring at any stage in the development of a rifted continent (perrift, synrift, or postrift), whereas magmatism derived from the shallow mantle would largely be synrift or postrift.

Another problematic aspect of understanding the relationship between extension and magmatism is defining the timing of rifting and/or extension. Extension may be fault controlled or via dike injection (Klausen and Larsen, this volume), and may be identified as the appearance of the first fault, the first volcanic rock, or the first depocenter. Is the onset of extension the timing of the initiation of continental extension, or is breakup marked by the formation of seafloor sensu stricto? Tens of millions of years can pass between the initiation of LIP formation (perrift) and the generation of seafloor, so it is important to understand absolute and relative timing of the geological processes leading to the formation of new seafloor. Any generalization about the apparent synchronicity of magmatism, extension, and uplift ignores the reality that, with the technology available, we can resolve the relative and absolute timing of these processes and so better understand rift processes.

In Figure 3 the relationship between flood volcanism (i.e., LIP formation) and the formation of oceanic crust is summarized for many of the volcanic rifted margins that formed in the past 200 m.y. (see also Courtillot et al., 1999). The age of the oldest oceanic crust adjacent to the volcanic rifted margin in question can be used as a minimum age of seafloor spreading because it is conceivable that this is not the oldest ocean floor, but merely the oldest seafloor for which samples exist (Fig. 3). The age of oceanic crust can be compared with the age of flood volcanism on the volcanic rifted margin to better understand the relationship between extension and magmatism.

In Ethiopia-Yemen, magmatism is dated by Ar-Ar methods as 31–26 Ma (Baker et al., 1996a; Hoffman et al., 1997; Ukstins et al., 2002). Extension (leading to the formation of domino fault-block terranes) is defined by Ar-Ar and fission-track dating of hanging-wall and footwall lithologies (Menzies et al., 2001). Extension in Yemen (i.e., southern Red Sea margin) began in the late Oligocene (ca. 26 Ma), coincident with a marked hiatus in extrusive activity and significant tectonic erosion and/or crustal cooling dated by fission-track methods and validated by Ar-Ar dating of unconformities as 19–25 Ma (Baker et al., 1996a; Menzies et al., 1997a). On the conjugate margin in Ethiopia, extension occurred along the length of the western escarpment ≥25 Ma, indicating that rifting occurred after the onset of flood basaltic volcanism ca. 31 Ma (Ukstins et al., 2002). Volcanic rocks were erupted from isolated centers located along the western escarpment in Ethiopia (Kenea et al., 2001; Ukstins et al., 2002). We conclude that much of the Ethiopian-Yemeni flood volcanism was perrift in character. While the timing will not be the same for all volcanic rifted margins, the southern Red Sea is an illustration of how breakup and the continent-ocean transition can be protracted.

In the case of the North Atlantic (Greenland-UK) (Fig. 1), LIP formation lasted from 61 to 53 Ma (e.g., Eldholm and Grue, 1994; Saunders et al., 1997) and the oldest oceanic crust indicates that extension must have taken place before 52 Ma (Fig. 3). From this it appears that volcanism in the North Atlantic straddled breakup with a perrift (LIP) and a synrift stage (SDRS) (Larsen and Saunders, 1998). Such a protracted period of volcanism may explain the attenuated, heavily intruded nature of the broad continent-ocean transition.

The details of the timing are less well known for Australia-India (Fig. 3). Volcanism on the Indian and Australian margins occurred between 100 and 130 Ma (e.g., Kent et al., 1997), and breakup between Australia, India, and Antarctica was 125–133 Ma (Fig. 3). It appears that volcanism on the rifted margin was synchronous with continental breakup, but that volcanism continued (sporadically?) during formation of oceanic crust.

In the Paraná-Etendeka volcanic rifted margins (Fig. 1), oceanic crust located off Africa is slightly older than that known off South America. The age of the oceanic crust indicates that extension occurred ca. 135 Ma, overlapping with the Paraná-Etendeka LIP (Peate, 1997). Because the main pulse of basaltic magmatism occurred ca. 130–133 Ma, it can be inferred that the LIP was largely perrift to synrift. A synrift stage is supported by the fact that the main volcanic units can be traced, and the vol-
canic stratigraphies matched, from the Etendeka across the Atlantic Ocean to the Paraná of Brazil (e.g., Milner et al., 1995; Mohriak et al., this volume). Synrift magmatism is supported by offshore valley systems that appear to be filled with extrusive lavas with later deformation and faulting-controlled emplacement of the volcanic units (cf. Clemson et al., 1999). Alternatively, both these observations could be explained by a synrift model for the magmatic activity. Initial pulses of magmatism would fill topographic lows, as described by Clemson et al. (1999), and further synrift activity would mantle the filled topography such that units were traceable from South America to Africa, as reported by Milner et al. (1995).

The relationships for the Central Atlantic magmatic province (Fig. 3) appear more complex, probably because of the size of the province and the extent to which it has been eroded. Continental magmatism has been dated at 198–201 Ma (Hames et al., 2000). However, along the eastern margin of North America the relationship between magmatism and tectonics is variable (J. McHone, 2001, personal commun.). In southeastern North America, volcanic rocks of the Central Atlantic magmatic province appear to postdate both the cessation of rifting by ca. 10 Ma, and uplift and/or erosion. This should be contrasted with Central Atlantic magmatic province magmatism in northeastern North America and northwestern Africa, where magmatism is synrift and rifting continued for ~25 m.y. after magmatism followed by Middle to Late Jurassic uplift (J. McHone 2001, personal commun.). SDRS from offshore northeastern United States are thought to have been emplaced ca. 175 Ma (Withjack et al., 1998; Benson, 2002; Schlische et al., 2002), and flood volcanism appears to be synrift or postrift. This contrasts with the North Atlantic margins (Greenland, UK) where a significant prerift flood volcanic stage is evident. However, there may be a bias in the rock record. In the Central Atlantic magmatic province, onshore intrusive rocks are used to define the timing of magmatism on the rifted margin. However, in deeply eroded volcanic rifted margins, like the Central Atlantic magmatic province, these hypabyssal and/or plutonic rocks may bias the dating toward the synrift stage. We use the Yemen volcanic rifted margin as an illustration of how hypabyssal and/or plutonic rocks may be largely synrift in age, despite a prerift history of 4–5 m.y. of flood basalt volcanism unrepresented in these exposed hypabyssal and/or plutonic rocks. In Yemen the original subaerial volcanic stratigraphy has an age of 31–26 Ma, and is known to be prerift (Baker et al., 1996a; Menzies et al., 1997a, 1997b). Hypabyssal and plutonic rocks underlyng or intruding the volcanic rifted margin have ages that are primarily younger than 25 Ma (Chazot et al., 1998, and references therein) and so intrusive activity, as exposed, is largely synrift. It appears that peak extension (and erosion 19–26 Ma) was associated with a possible extrusive hiatus, but with significant intrusive activity exemplified by the dike swarms and granite-gabbro-syenite laccoliths. This synrift intrusive stage is confirmed by <25 Ma dikes that are parallel to the Red Sea margin and that occur in Saudi Arabia and Egypt (Chazot et al., 1998, and references therein). If these hypabyssal rocks were all that remained of the Yemen volcanic rifted margin, their ages (21–25 Ma) would be biased toward dating the peak of extension (<25 Ma), possibly the period of formation of the SDRS (<25 Ma), but definitely not the onset of flood volcanism (31 Ma) or the true age of the magmatic period (31–19 Ma).

The relationship between magmatism and faulting is complex, and synchronicity between the two processes, a model driven expectation, appears to be unsupported in many volcanic rifted margins. Magmatism may have predated rifting by several million years as in the Ethiopia-Yemen volcanic rifted margins, postdated rifting as in some of the Central Atlantic volcanic rifted margins, or straddled the prerift to synrift transition as in the North Atlantic volcanic rifted margins of Greenland-UK. To resolve this issue, as in all volcanic rifted margins, accurate dating of extrusive and intrusive volcanic rocks on the continental margin and the ocean floor and the timing of extension are needed. The volcanic part of volcanic rifted margins is complex, with the possibility, as seen in Yemen, of prerift (i.e., flood basalts), synrift (i.e., hypabyssal and/or plutonic rocks), and posterosional (i.e., alkaline volcanic rocks) stages (Chazot et al., 1998). Selection of any of those rock types in trying to understand the relationship between magmatism and tectonics may predetermine the outcome. The magmatic source for volcanic rifted margins may be deep (i.e., plumes) or shallow (i.e., asthenosphere). Shallow melt production from decompression melting should be synrift or postrift (e.g., Central Atlantic magmatic province). Deep melt production can exploit already established rift systems (e.g., India-Australia) or help generate new rift systems (e.g., Yemen-Ethiopia) and so may be prerift, synrift, or postrift.

**Rift margin mountains: Prevolcanic or synvolcanic uplift and erosion**

Many rifted continental margins are bordered by eroded mountain ranges, evident as topographic highs proximal to the rifted margin. This can be seen in the Paraná-Etendeka, north-eastern African, Greenland-Scotland, and the Deccan Traps (Fig. 2). The highest points in Arabia and the UK are atop relic mountain ranges that border the volcanic rifted margin of the southern Red Sea and the North Atlantic, respectively. The continental margins of eastern Brazil and western India are bordered by steep scarps facing the rift valley, whereas the rift shoulder is characterized by less dramatic slope development (Gallagher et al., 1994). In most of these volcanic rifted margins no kilometre-scale mountain range existed prior to rifting and magmatism, so an important part of the evolution of volcanic rifted margins is mountain building and erosion. The juvenile nature of the mountains and/or landscape is evident from the drainage patterns, which were influenced by mountain building and margin uplift. Cox (1989) drew attention to the drainage patterns on volcanic rifted margins and the fact that, in many cases, major rivers flowed away from the present-day coastline and followed a
lengthy inland course because of the uplifted rift margins. Some of the best examples are the Rio de la Plata–Parana (Brazil), which has its headwaters <250 km from the Atlantic margin; the river follows a course of several thousand kilometers to the west and south away from the present coastline, reaching the Atlantic Ocean at Buenos Aires. Similarly the Blue Nile has headwaters in the rift mountains of the Ethiopian highlands within 500 km of the Red Sea. However, these waters flow westward and northward for several thousand kilometers to reach the Mediterranean Sea at Alexandria.

To help bracket the period of uplift and mountain building, limitations have to be placed on the prevolcanic rifted margin paleoenvironment and the onset of erosion. Paleoenvironmental clues have to be present in the prerift sedimentary rocks that underlie the earliest volcanic rocks of volcanic rifted margins. However, caution has to be exercised in interpreting the prerift sedimentary rocks, because in many cases they are very difficult to date and therefore cannot be definitively shown to relate in space and time to the volcanic rifted margin sensu strictu. Whereas paleoenvironmental analysis helps determine the approximate time when mountain building began, the timing of crustal cooling or denudation may be used to define the minimum time when topography existed on the margin. On most volcanic rifted margins (Table 1) the magnitude of uplift prior to volcanism is on a scale of hundreds of meters, typically measured by locating marine horizons within the volcanic stratigraphy. Around the Red Sea, as in many other volcanic rifted margins, marine sediments are several kilometers above present sea level (Davison et al., 1994), revealing significant rift shoulder uplift.

In Yemen a paleoshoreline existed (>31 Ma) close to the present location of a 4-km-high mountain range. Paleocurrent information and the maturity of the prevolcanic sediments in Yemen require a hinterland on what is now the opposite side of the rift in the Danakil horst, Eritrea (Al’Subbary et al., 1998). The prevolcanic sedimentary rocks imply that the continental masses were close to sea level in the southern Red Sea and, by inference, Eritrea (Al’Subbary et al., 1998). The predominance of subaerial volcanic rock units (rather than submarine flows or hyaloclastites) also indicates a subaerial continental environment at 31 Ma; some rift-related uplift must have occurred before that time. Possibly the initiation of uplift is recorded in changes to the orientation of the paleoshoreline, along with a shallow marine to continental transition that occurred prior to volcanism. These changes indicate that uplift of that surface (ca. 31 Ma) was tens to hundreds of meters. However, the exact age of these sediments relative to the period of formation of the volcanic rifted margin is unknown. In the unlikely event that these sedimentary rocks are considerably older than the volcanic rifted margin, the paleoenvironmental changes would not relate to the evolution of the volcanic rifted margin.

Fission-track ages date crustal cooling and hence rapid tectonic denudation as having occurred between 19 and 26 Ma (Menzies et al., 1997a) on the Red Sea margin. We presume that Oligocene-Miocene denudation required greater topography than the paleoshoreline inferred to exist at 31 Ma, hence such topography was generated, and the period of uplift and exhumation is bracketed, in the late Oligocene (26–31 Ma). Independent verification of this period of denudation is to be found in unconformities in the volcanic stratigraphy that formed between 19 and 26 Ma (Baker et al., 1996a). In contrast to the largely synvolcanic uplift in Yemen, uplift on a scale of hundreds of meters is believed to have preceded volcanism in the North Atlantic province (e.g., Larsen and Saunders, 1998). In western Greenland major unconformities beneath the volcanic rocks are associated with fluvial peneplanation and valley incision, indicating a period of prevolcanic uplift and erosion. However, in eastern Greenland during the same time the landscape was close to sea level and, in northwest Scotland- Faeroes, the prevolcanic landscape was a low-relief, vegetated land surface (Jolley, 1997). Furthermore, in northwest Scotland subaerially weathered marine sediments (chalk) underlie the low-estmost flood volcanic rocks (G. Fitton, 2001, personal commun.). The prerift North Atlantic volcanic rifted margin (Brodie and White, 1994) could be classified as a low-relief land surface with some incised river systems. The proximity of that land surface to sea level is revealed by studies in Norway (Planke et al., 2000), where the margin is believed to have developed in the continental to oceanic transition.

In Namibia and Brazil (Etendeka-Parana) the basal volcanic rocks overlie and are interbedded with continental eolian sandstones, which occasionally overlie fluvial deposits. A large eolian erg system is reported intercalated with the lowermost flood basalts (Jerram et al., 1999), but how far above sea level it was formed is not known. On this volcanic rifted margin uplift and doming prior to rifting could be argued for due to the lack of Upper Karoo sediments (Clemson et al., 1999). This may be consistent with conclusions based on fission-track data that argue for a pre rift elevation of ~500 m in southeastern Brazil (Gallagher et al., 1994).

The degree of preservation of volcanic rifted margins is inextricably linked to climate, elevation, and the amount and/or rate of erosion. The youngest volcanic rifted margins survive as 3–4-km-high mountain ranges in the desert climate of northeastern Africa, and Cretaceous-Tertiary volcanic rifted margins are characterized by ~5–7-km-thick volcanic sections in the mountain ranges of subpolar Greenland, deeply eroded rift mountains in the west maritime climate of the UK, and major scarp retreat in tropical India and Brazil. The western Ghāt escarpment (Deccan) is believed to have an erosional, rather than a tectonic origin, and scarp retreat is believed to be the major determinant of landscape with the original continental margin ~75 km west of its present location. Erosion over 200 m.y. has reduced the subaerial portion of the Central Atlantic magmatic province in the eastern United States and western Africa to a dike swarm. However, submarine equivalents to these margins have survived offshore as SDRS. Just as in the relationship between magmatism and rifting, we have a complex history of up-
Volcanic rifted margins: Summary

Volcanic rifted margins evolved in response to local thermotectonic conditions, and consequently marked differences can be found in the temporal and spatial relationships between tectonics, magmatism, uplift, and erosion. Of all passive continental margins around the world, 90% are volcanic rifted margins to varying degree, the exceptions being continental margins in eastern China, Iberia, the northern Red Sea, South Australia, the Newfoundland Basin–Labrador Sea, and possibly the Gulf of California. Although the Arctic and Antarctic margins have largely unknown status, parts of the Antarctic margin are believed to be volcanic (Fig. 1).

A considerable variation exists in volcanic rifted margins. The prevolcanic environment can vary from shallow marine-continental (e.g., North Atlantic, fluvial-continental (e.g., Yemen–Ethiopia) to eolian continental (e.g., Etendeka). Flood volcanism can be thick (e.g., 7 km, Greenland) or relatively thin (e.g., 1.5 km, Deccan). Volcanism can be represented by predominantly basaltic volcanic rocks at the base and by mainly silicic volcanic rocks at the top (e.g., Yemen), or silicic volcanic rocks may be found throughout the volcanic stratigraphy (e.g., Deccan). Processes related to volcanic rifted margin can vary in time and space. Magmatism can predate breakup extension by several million years (e.g., Yemen–Ethiopia), magmatism and breakup can be synchronous (e.g., Greenland–North Atlantic Tertiary volcanic province), or magmatism can postdate breakup by several million years (e.g., Australia–India). Magmatism in volcanic rifted margins may have originated by decompression melting associated with lithospheric thinning and/or upwelling of thermal anomalies modified by melting of lithospheric rocks. In most volcanic rifted margins prevolcanic uplift can vary from tens of meters (Yemen) to hundreds of meters (North and South Atlantic), but it appears that kilometer-scale prevolcanic uplift is not as widespread as some models have predicted.

The magmatic and structural evolution of individual volcanic rifted margins is complex and may not fit simple models. This may be due to the geology, age, and thickness of the pre-rift lithosphere and proximity to plume heads, which are potentially variable in temperature, longevity, and dimensions. There appears to be a continuous gradation from volcanic rift margins to nonvolcanic rifted margins; a possible continuum is evident in the southern Red Sea. Much remains to be learned about the extent to which plumes drive, or are focused by, lithospheric extension and the exact geophysical and geochemical nature of plumes. Perhaps we can better understand the process of formation of volcanic rifted margins by comparing and contrasting their geological characteristics with nonvolcanic rifted margins where continental rifting occurs without thermally enhanced mantle (e.g., Newfoundland, Iberia), or the formation of intraplate large igneous provinces in ocean basins (e.g., Ontong-Java oceanic plateau) and continents (e.g., Siberian flood basalts) where widespread rifting is absent. Clearly a single active rifting model cannot explain the formation of all volcanic rifted margins around the world.

Volcanic rifted margins: Characteristics

The following characteristics are sufficiently common in volcanic rifted margins to be diagnostic.

1. Flood volcanism may have reached 4–7 km thickness prior to erosion, which has reduced several margins to thicknesses of 1–2 km.

2. Basaltic and silicic volcanic rocks are erupted subaerially. Of the exposed subaerial basaltic rocks, 70%–80% occurred in <3 m.y. The eruption of silicic volcanic rocks can occur during or after eruption of the basaltic rocks and can last for as many as 5 m.y.

3. Magmatism and rifting are not necessarily synchronous. Magmatism can occur before, during, or after rifting. In some volcanic rifted margins a magmatic hiatus coincided with the peak of extension.

4. Rift mountains are uplifted and rapidly eroded by synrift (or postrift) processes.

5. Seaward-dipping reflectors comprise a mixture of volcanic flows, volcanioclastic deposits (e.g., hyaloclastites), and nonvolcanic sediments. Formation of the SDRS postdates flood volcanism.

6. HVLC (~7.4 km/s) forms in the continent-ocean transition by igneous processes and reaches considerable thicknesses (10–15 km). The exact relationship between the formation of the high-velocity lower crust and continental (flood) or oceanic (mid-ocean ridge basalt) volcanism is unknown.

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