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55. PETROLOGIC SYNTHESIS: LAU BASIN TRANSECT (LEG 135)[1]

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ABSTRACT

The Lau Basin–Tonga Ridge drilling transect, Ocean Drilling Program Leg 135, collected cores from six sites in the backarc basin and two on the Tonga Ridge. The Tonga Ridge cores sampled some of the oldest crust known in the region at 5200 m below sea level. These rocks are remnants of a late-middle Eocene rhyolitic volcanic feature that was subaerial, or in very shallow water, at the time of eruption. It represents an early phase of arc volcanism of the Vitiaz Arc. Mafic dikes or sills of arc-affinity basalt and basaltic andesite were found at the same site higher in the cored section. These are intrusive into distal facies of arc-derived volcaniclastic turbidites of late Miocene age. The turbidites and intrusions are attributed to the Lau Ridge, which was the active volcanic arc at that time.

Crustal extension that formed the Lau Basin began in late Miocene time (about 6 Ma) and overlapped with extension and volcanism on the Lau Ridge. The initial phases of crustal extension involved rifting to form basin-range type structures, and were accompanied by magmatism. Developing sub-basins were partially filled with both tholeiitic-basalt and arc-composition volcaniclastic sediments. The latter probably were derived from ephemeral intrabasin arc-composition volcanoes. It is not likely that the early extension was accomplished by seafloor spreading; essentially static long-lived magmatism (e.g., >3 Ma) prevailed in individual sub-basins. Drill-core data suggest that a wave of basaltic volcanism crossed the basin from west to east. Seafloor spreading probably began from a point on the Peggy Ridge at about 4–5.5 Ma and propagated southward, forming the Eastern Lau Spreading Center. Its present southern apex is at the Valu Fa Ridge. A second propagator began at about 1.5 Ma, forming the present Central Lau Spreading Center. Data from the sediment cores suggest that the intrabasin, arc-like volcanoes migrated across the basin, perhaps tracking eastward migration of the Tonga Ridge and Trench. Some intrabasin arc constructs were cut off and isolated on the west side of the propagator to form isolated rafts of arc material within MORB-like crust.

Mineralogical, chemical, and isotopic data support inferences that the basaltic crust of the Lau Basin formed by mixing between MORB-source melts and SSZ arc-like melts. Isotope data also show that early melts had a "Pacific" mantle signature; more recent melts carry an "Indian" mantle signature.

INTRODUCTION

The western Pacific Ocean basin is rimmed by an array of island-arc systems and their related trenches. Marginal seas, most of them relatively shallow, are situated between the emergent and submarine arc systems and represent regions of oceanic crust formed contemporary with adjoining volcanic island arcs in suprasubduction zone settings. Alfred Wegener (1929) called attention to these arc systems in his treatise on the *Origin of Continents and Oceans,* in which he proposed that "the island arcs, and particularly the eastern Asiatic ones, are marginal chains which were detached from continental masses, when the latter drifted westward and remained fast in the old sea floor, which was solidified to great depths. Between the arcs and the continental margins later, still-liquid areas of seafloor were exposed as windows." More than 40 yr later, a series of nearly contemporary papers by Karig (1970, 1971), Packham and Falvey (1971), Sleep and Toksöz (1971), and Moberly (1972) presented the first discussions of the geometry and tectonics of western Pacific backarc basin-arc-trench systems. These authors all drew attention to the anomalous situation of extensional geologic processes operating in zones of plate convergence and proposed some ideas as to the mechanisms involved. The general theme proposed in these papers was that new crust had been formed in a zone of extension between an inactive remnant arc and an active volcanic island arc; a kinematic relation between the subduction process and the backarc basin extension was inferred but neither the mechanism nor the evolution were well understood. More recent studies have suggested that the subduction process causes, or allows crustal extension above the inclined seismic zone (e.g., Furlong et al., 1982, proposed that an increase in subduction velocity may promote backarc spreading). Counterflow above the subducting lithosphere plate and concurrent mantle upwelling probably are the main driving forces behind backarc basin evolution (e.g., Hawkins et al., 1984). Ribe (1989) proposed that backarc spreading induces mantle flow above the subduction zone and helps account for the observed distribution of distinctive magma types. The upwelling and lithosphere extension are either a cause or the result of eastward "rollback" of the trench (Elsasser, 1971; Uyeda and Kanamori, 1979); the Tonga Trench may be migrating eastward at up to 10 cm/yr (Carlson and Melia, 1984).

Although they are situated in zones of lithosphere shortening, between plates with opposing relative motion, abundant evidence exists that backarc-arc-trench systems are loci of crustal extension and the generation of new crust, as illustrated in Figure 1. Fractional melting of the upwelling mantle diapirs forms new crust in the backarc basins, the island arcs, and part of the forearc. These melts are largely derived from the mantle wedge lying above the subduction zone, that is, the suprasubduction zone (SSZ) mantle (Pearce et al., 1984). Additional contributions may come from the counterflow of mantle into the SSZ region and from the subducted lithosphere plate (e.g., Tatsumi et al., 1986; Takazawa et al., 1992). The result is that new crust has varied enrichments in low-partition coefficient elements relative to crust formed at oceanic spreading centers. Ocean Drilling Program (ODP) studies in the western Pacific basin investigated several of these convergent margin systems with the objective of gaining a better understanding of their evolution. The Lau Basin–Tonga backarc-arc-trench system (Fig. 2) affords a classic area in which to study the complex petrologic and tectonic processes that operate at convergent oceanic plate margins. A major objective for Leg 135 was to sample the crust of the Lau Basin and the Tonga Ridge to assess the sequential development of crustal composition in both settings. This report presents a synthesis of the results of petrologic studies on Leg 135 samples that give new insight to the petrologic evolution of the Lau Basin.

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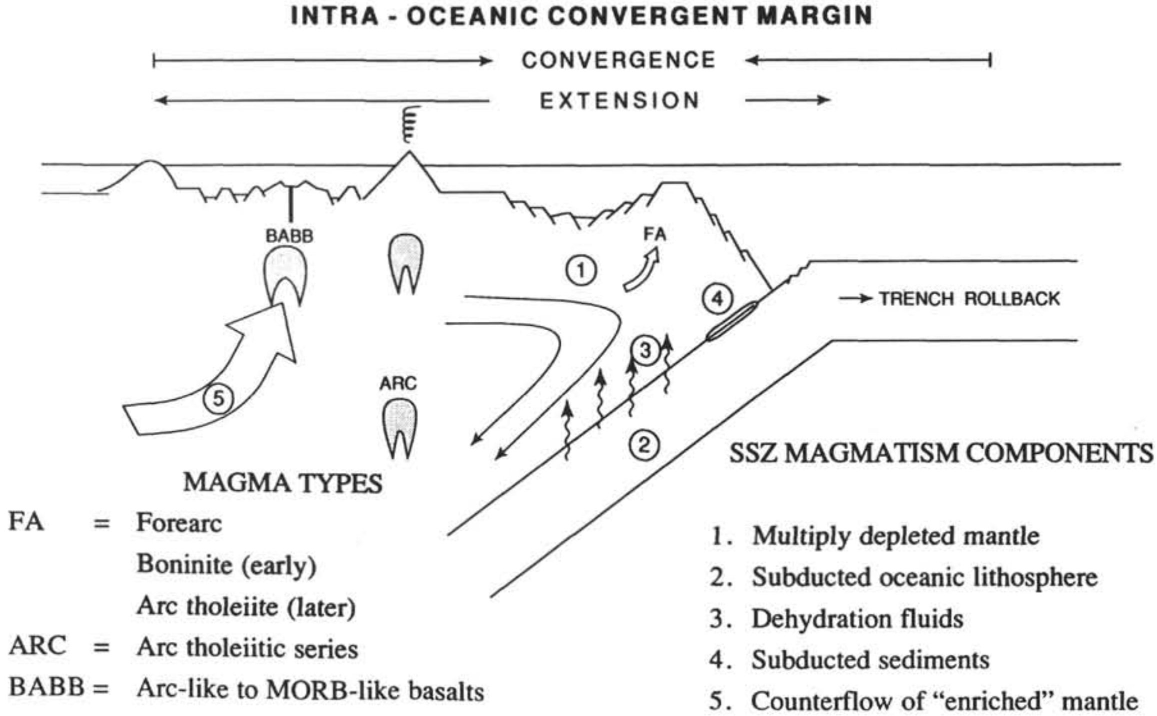

Figure 1. Schematic cross-section of an intraoceanic convergent margin showing the remnant arc, backarc basin, active arc, and forearc (FA). Sites of magmatic activity are shown by arrows and shaded crescentic patterns. Numbers refer to potential components that may contribute to SSZ magmatism.

SUMMARY OF THE REGIONAL GEOLOGIC SETTING

The Lau Basin is the backarc basin separating the inactive Lau Ridge remnant volcanic arc from the Tonga Ridge (Fig. 2). Opening of the Lau Basin, as a consequence of crustal extension above the Tonga Trench subduction system, is the most recent (i.e., <6 Ma) event in a long sequence of crustal extension episodes that may be traced back to Late Cretaceous time and the initial breakup of the eastern Australian continental margin. A comprehensive synopsis of interpretations of the evolution of this broad region, spanning nearly 35° of longitude, is given by Kroenke (1984) and is summarized here. Many of the episodes of extension appear to have been coeval with the development of volcanic island-arc systems. In the brief summary that follows, I will focus on the chronology of crustal extension in this region (Fig. 3) to set the stage for discussing the most recent episodes that we studied on the Leg 135 transect. A major point that will be made is that two styles of deformation can be recognized in the evolution of the Lau Basin. The initial stage was accomplished by crustal extension and rifting; the second stage has involved seafloor spreading and rift propagation.

The breakup of Gondwanaland in mid- to late Jurassic time, the initial rifting of Australia from Antarctica at about 110 to 90 Ma, and the subsequent commencement of seafloor spreading at 83 Ma, or Anomaly 34-time (Cande and Mutter, 1982), all represent a major disruption of the lithosphere in the southwestern Pacific. Crustal extension of eastern Australia started in Late Cretaceous time with the opening of the Tasman Basin (Hayes and Ringis, 1973). The central anomaly of the Central Tasman Basin has been identified as Anomaly 24 and the oldest as Anomaly 34. Initially, these were considered to correspond to an age range of 80 to 60 Ma (Hayes and Ringis, 1973), but they were reappraised as ranging from 82 to 60 Ma (Weissel and Hayes, 1977). Cande and Mutter (1982) subsequently have revised the age of Anomaly 34 to 86 Ma. The basin youngs northward where the oldest magnetic lineation was identified as Anomaly 32, that is, 70 Ma (Hayes and Ringis, 1973). The opening of the Tasman Sea rifted away the Lord Howe Rise, the Campbell Plateau, the Norfolk Ridge, and probably part of present-day New Caledonia from the Australian Margin (Burns and Andrews, 1973; Packham, 1973; Bentz, 1974). Both the Lord Howe Rise and the Norfolk Ridge are capped with basaltic volcanic islands, but these appear to be younger features superposed on continental crust. Shor et al. (1971) interpreted the crust of the Lord Howe Rise as having "a deep root comparable to those found under continental shelves in many parts of the world." The depth to the mantle under the Lord Howe Rise was estimated to be about 29 km. Drilling at DSDP Site 207 established that the Lord Howe Rise has an igneous basement of rhyolite overlain by a sediment cover of biogenic ooze that ranges in age from Maastrichtian to late Pleistocene. Basal sediments (glauconitic silty claystone and sandstone, of Maastrichtian [?] age) overlie rhyolitic rocks with textures that range from lapilli tuffs produced by explosive eruptions to vitrophyres, some of which are autobrecciated. Textural interpretations of the lapilli tuffs suggest that they were erupted either subaerially or in very shallow water. The rhyolites have been dated by K/Ar techniques as 93.7 ± 1.2 m.y. (Burns and Andrews, 1973; van der Lingen, 1973). An origin on the continental margin of Australia is inferred.

The Norfolk Ridge and New Caledonia were separated from the Lord Howe Rise by the opening of the New Caledonia Basin in Late Cretaceous time (Kroenke, 1984). Although the Norfolk Ridge has not been studied well, Shor et al. (1971) interpreted a crustal thickness of about 22 km and concluded that it probably was continental. Dubois et al. (1973) reported a crustal thickness of 20 to 25 km. New Caledonia is considered to be, in part, continental with a crustal thickness of up to 35–39 km under the central mountain chain (Dubois et al., 1973). The older parts of the island are considered to have once been part of a Late Jurassic convergent plate margin of Gondwanaland (Blake et al., 1977). Pre-Tertiary rocks comprise Permian, or older, quartzofeldspathic rocks and submarine basalts; Permo-Triassic blueschists; Triassic and Jurassic andesitic arc-derived volcaniclastic rocks; and Late Jurassic to Early Cretaceous high-pressure, low-temperature meta-

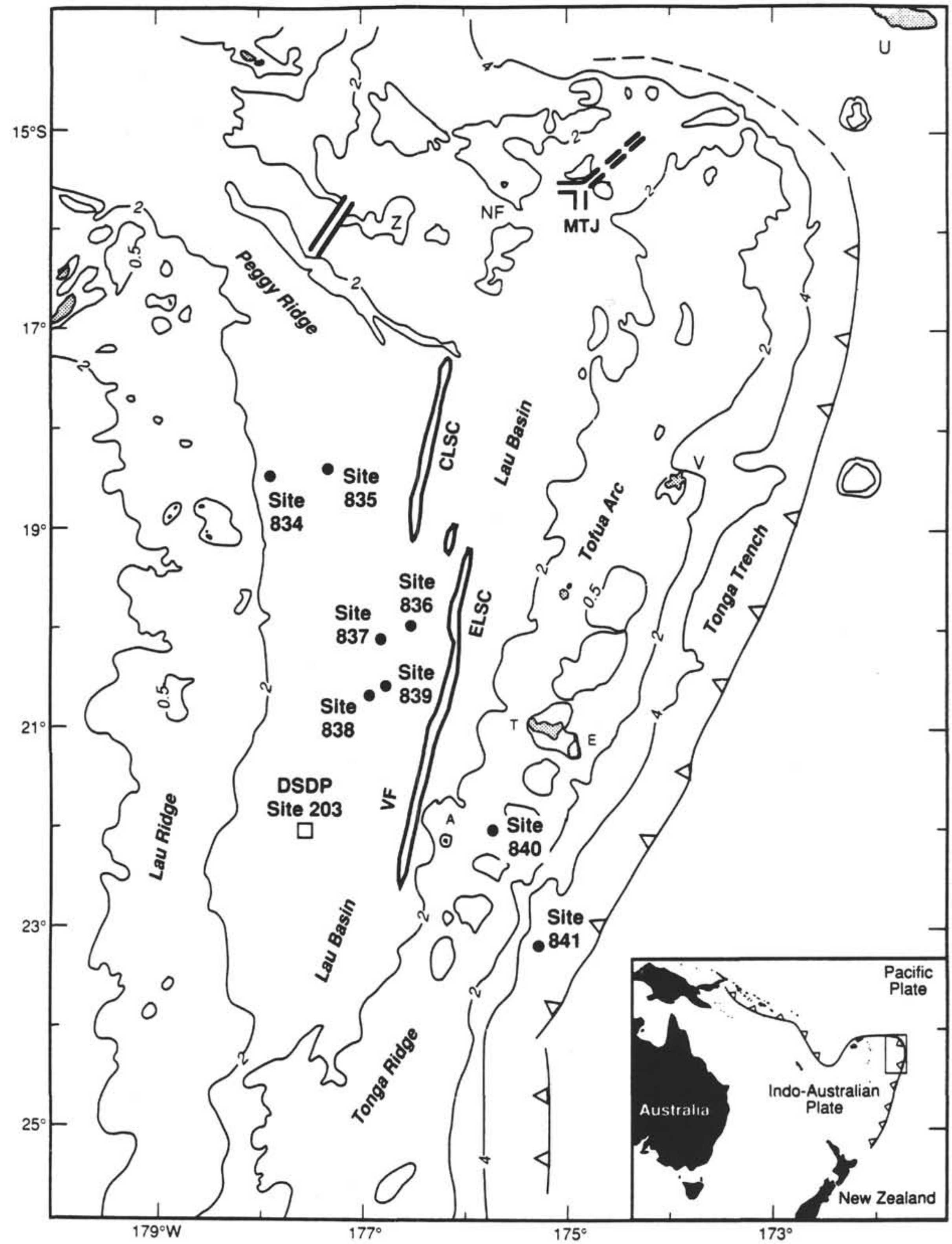

Figure 2. Lau Basin and Tonga Trench-Arc system showing locations of Leg 135 Sites 834–841 and DSDP Site 203. Major features shown include sites of modern volcanism in the Lau Basin: Mangatolu Triple Junction (MTJ) (Hawkins, 1989; Nilsson et al., 1989; Nilsson, 1993); Central Lau Spreading Center (CLSC), Eastern Lau Spreading Center (ELSC), and Valu Fa Ridge (VF) (von Stackelberg and von Rad, 1990; Parson et al., 1990). Islands and shoals are Upolu (U), Western Samoa, Niuafo'ou (NF), Zephyr Shoal (Z), Vava'u (V), Tongatapu (T), 'Eua (E), and 'Ata (A). Contour interval in kilometers.

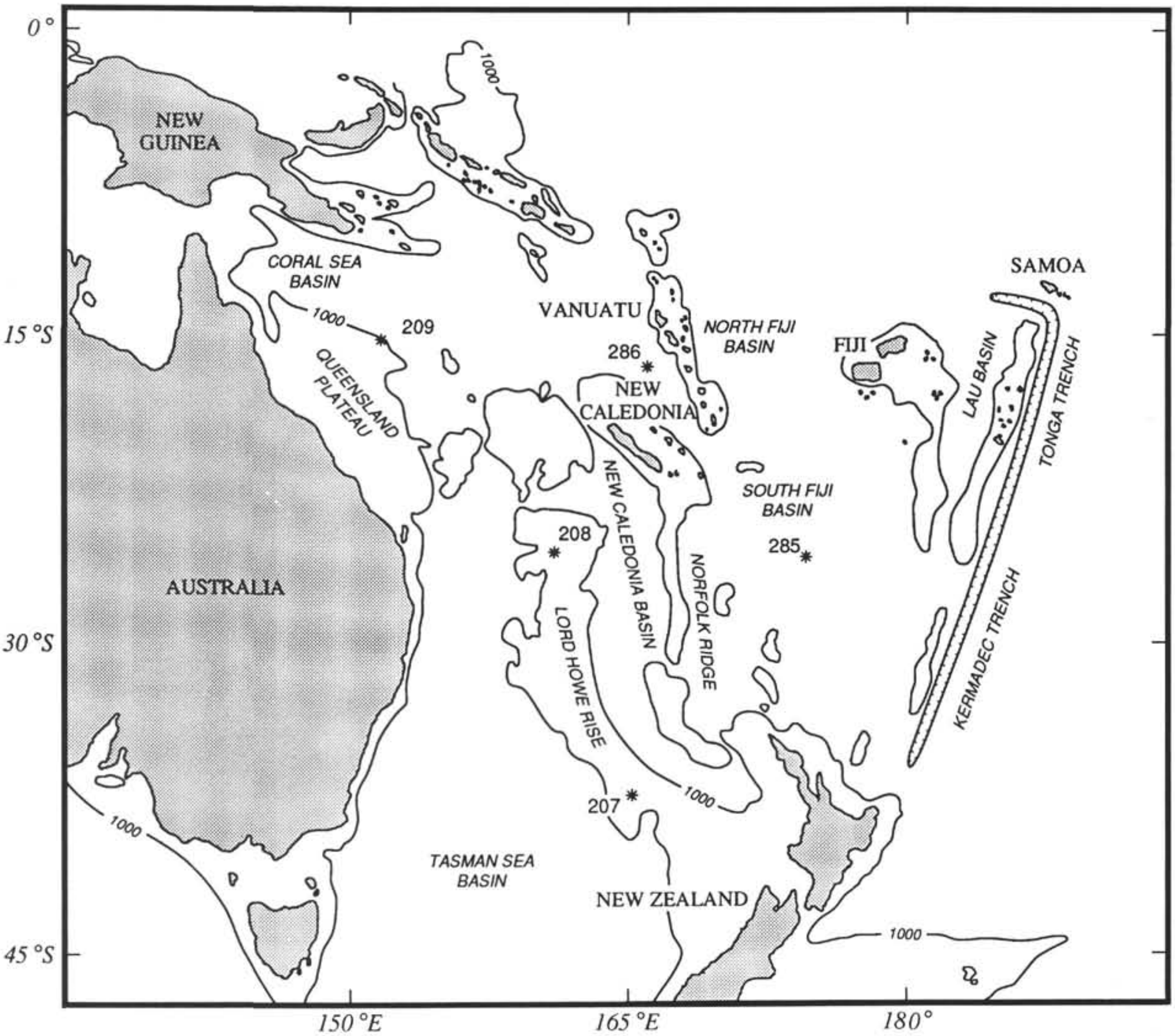

Figure 3. Regional distribution of marginal seas, backarc basins, island arcs, and trenches between Australia and the Tonga Trench. Numbered stars are locations of DSDP drill sites.

morphic rocks (blueschists and eclogite). Part of the island is oceanic rocks and includes peridotite overlying tholeiitic basalts; both have been thrust over boninites and Cretaceous to Eocene sediments. Emplacement of the thrust sheets in late Eocene time was followed by intrusion of granodiorite, adamellite, and hornblende quartz diorite also in the late Eocene (Dubois et al., 1973). The New Caledonia Basin has a seismic structure showing that the sediment layers are thicker than on the adjacent ridges and the crust is thinner; the depth to mantle seismic velocities (6.8 km below seafloor) is typical of oceanic crust (Shor et al., 1971). The timing of the opening of the New Caledonia Basin is controversial but, as discussed by Kroenke (1984), it may have been contemporary with Tasman Sea opening. However, rifting appears to have continued for a longer time. An alternate view is that it is slightly younger (e.g., early Paleocene, Anomaly 27 to 26, or 65 to 53 Ma). DSDP drilling at Site 206 in the New Caledonia Basin ended in mid- Eocene (Paleocene ?) sediments without reaching basement. Backarc basin extension in the North and South Loyalty Basins generated oceanic lithosphere through Eocene time (Lapouille, 1982; Maillet et al., 1982). Kroenke (1984) proposed that the same ridge system that formed these basins may have migrated to the southeast and began to form new crust in the South Fiji Basin during Oligocene time.

The South Fiji Basin is distinctive in preserving a crustal fabric, as marked by magnetic lineations, that indicates an origin by spreading from a ridge-ridge-ridge system during Oligocene time (Weissel and Watts, 1975). The magnetic patterns were interpreted by Watts et al. (1977) as recording seafloor spreading from anomaly 12 time (about 35 Ma) to anomaly 7A time (about 28 Ma). DSDP Site 285 drilled through about 565 m of nannofossil ooze, and an increasing amount of volcaniclastic material with depth, before encountering a diabase sill. The oldest sediments were identified as early middle Miocene (Andrews, Packham, et al., 1975). The wedge shaped basin is elongated in a northeasterly direction parallel to well-developed magnetic lineations; the direction of elongation meets the Lau Ridge at a high angle. This geometry puts important constraints on general models for backarc basin evolution that involve simple rifting and spreading between active and remnant volcanic arcs. The South Fiji Basin is separated from the North Fiji Basin by the Hunter Fracture Zone. This major linear feature may represent a trench-trench transform fault between the New Hebrides and Tonga trenches.

The North Fiji Basin is a broad (more than 800 km wide), high-standing region of oceanic crust. Bounded on the west by the Vanuatu–New Hebrides arc and trench system, the basin has a complex fabric that includes several active spreading ridge systems, and at least one ridge-ridge-ridge triple junction. The age of the basin is not well known, but it is considered to be no older than late Miocene (about 8 Ma; Malahoff et al., 1982). The Lau Basin overlaps in time the development of the North Fiji Basin.

Areas of crust separating the extensional basins listed above are either fragments of continental crust (e.g., Lord Howe Rise and parts of New Caledonia and the Norfolk Ridge); ophiolite complexes (e.g., part of New Caledonia); or volcanic island arcs (e.g., Three Kings Rise, Loyalty Islands, Lau Ridge, Vanuatu, and the composite microcontinent of Fiji). There is a general younging of the basins from west to east and a similar progression in age is seen in the island arcs. However, inasmuch as reversals in polarity of subduction direction are recognized, it is an oversimplification to suggest that the locus of subduction and basin extension migrated continuously from west to east.

VANUATU-FIJI-LAU-TONGA AREA

Magmatic Chronology

Evolution of the Lau Basin is tied closely to the evolution of the Tonga Trench arc systems. These are broadly related to the early Cenozoic Vitiaz Arc, for which the chronology of magmatic events in the Vanuatu-Fiji-Lau-Tonga region shows a more or less continuous record of arc volcanism from late Eocene to the present. The locus of activity has migrated, and there may be a small-scale episodicity, but in general this has been an area of plate convergence coupled with continual massive upwelling of hot mantle and related partial melts through Cenozoic time. A controversy remains as to whether the late Eocene convergence and arc volcanism were related to subduction at a northeast-facing Vitiaz Trench system that was subsequently replaced by the modern Tonga Trench system, or the Tonga Trench system has prevailed since Eocene time and formed the late Eocene arc material preserved in the Fiji-Lau-Tonga region (e.g., Gill, 1987). Deciphering the original tectonic setting and orientation of the Vitiaz Arc are beyond the scope of this paper, but the presence of Eocene volcanic rocks on the Leg 135 drill transect is important in tracing the evolution of magma types that is the major focus of this synthesis.

The convergent margin history has been well studied on the Fiji Islands where Gill (1970, 1987), Whelan et al. (1985), and Wharton et al. (1992) recognize the following chronology of events: (1) an early arc stage from 45 or 35 Ma to 10 Ma (this may have been caused by subduction at the Vitiaz Trench and accompanied by opening of the South Fiji Basin as a backarc basin in Oligocene time (Watts et al., 1977); (2) a mature arc stage from 10 to 5 Ma (this includes the main activity on the Lau Ridge and was probably the result of subduction at the Tonga Trench); (3) an early rifting stage of the arc from 5.5 to 3 Ma; and (4) a late-rifting stage of the arc since 3 Ma. Initial rifting of the Lau Basin began at about 6 Ma (or perhaps as early as 7 Ma; Parson and Hawkins, this volume) and continued during the arc-rifting stages.

Although the evidence for regional Eocene magmatism is spotty, it includes pre-late Eocene rhyolitic rocks at Site 841 (Parson, Hawkins, Allan, et al., 1992; Bloomer, this volume). The rhyolite has been dated by K-Ar techniques as 42.7 and 46.5 Ma, averaging 44 ± 2 Ma (McDougall, this volume). The Ar-Ar dates on basalt and gabbro cobbles from 'Eua yield ages of 40 to 46 Ma (Duncan et al., 1985). Rocks of similar age probably occur on Viti Levu, Fiji, where upper Eocene reef limestones with *Discocyclina* are interlayered with pillow lavas (Colley and Hindle, 1984; Colley et al., 1986). Some of these lavas are low-Ti, high-magnesian andesites (Gill, 1987). These have boninite affinities and may be the consequence of the earliest stage of convergence at an ancestral Tonga or Vitiaz trench (Kroenke, 1984; Gill, 1987).

Cole (1960), Colley and Hindle (1984), and Colley et al. (1986) report late Eocene and early Oligocene volcanic rocks in the Wainimala Group on Viti Levu, Fiji. Oligocene magmatic activity is recorded on the Tonga Ridge, at the island of 'Eua, where basaltic to andesitic flows have radiometric ages of 33 to 31 Ma (Duncan et al., 1985). No direct evidence exists for pre-middle Miocene volcanism on the Lau Ridge, but Oligocene and early Miocene volcanism and sedimentation have been postulated on the basis of the regional occurrences of rocks with these ages, as cited previously. As discussed in a subsequent section, the main record of volcanism on the Lau Ridge is of Miocene age. 'Eua also has andesitic dikes with ages of 19 to 17 Ma that are intrusive into older volcanic units. Paleontologically dated tuffaceous sediments on 'Eua indicate that volcanism continued until 5 to 3 Ma (Cunningham and Anscombe, 1985). Leg 135 drilling at Site 841 on the Tonga Ridge encountered a series of basaltic and andesitic sills or dikes intrusive into late Miocene arc-derived volcaniclastic turbidites (Parson, Hawkins, Allan, et al., 1992; Bloomer et al., this volume). Volcanic activity on the Lau Ridge continued into Pliocene time, as indicated by 4.5- to 2.5-Ma rocks of the Korobasaga Volcanic Group and <2.5-Ma rocks of the Mago Volcanic Group (Cole et al., 1985). The spectrum of radiometric ages indicates that the volcanic activity was more or less continuous throughout this time, and, as discussed more fully in a later section, overlapped with the beginning of crustal extension that formed the Lau Basin.

Summary of the Local Geologic Setting

Lau Ridge

The Lau Ridge lies on the western margin of the Lau Basin (Fig. 2); it comprises islands and atoll reefs as well as a submerged chain of conical features assumed to be volcanic seamounts. The islands commonly are surrounded by a barrier reef, a fringing reef, or both. A submarine extension to the south, the Colville Ridge, gives a total length of 2400 km for this remnant arc system. The emergent Lau Islands form a chain about 500 km long between latitudes 21°S and 16°45′S. They are maturely dissected volcanic cones or coral-capped parts of a remnant volcanic arc that was up to 80 km wide. Many of the islands are completely covered with Miocene to Pliocene age limestone and no volcanic rocks are exposed. Some of the younger volcanoes (e.g., Pliocene), lack a limestone cover and presumably postdate the main phase of reef development (Woodhall, 1985). No data are available for the submerged part of the Lau Ridge other than the bathymetry which shows generally conical forms rising to sharp peaks. Evidence from the Lau Ridge indicates that arc volcanism began at least by mid-Miocene time and was active until early Pliocene time, that is, from 3.5 to 2.5 Ma (Gill, 1976; Cole et al., 1985; Woodhall, 1985). The Lau Ridge–Colville Ridge system is postulated to be a remnant of a more extensive volcanic arc ("an outer Melanesian island arc"; Woodhall, 1985) that originally included part of the present Tonga Ridge, the Fiji Platform, and the New Hebrides Ridge. A detailed summary of the geologic history of the Fiji Island Group from the perspective of arc and backarc basin evolution was presented by Gill (1987). The Lau and Tonga ridges were separated during the extensional development of the Lau Basin (e.g., Karig, 1970; Woodhall, 1985; Cunningham and Anscombe, 1985). A similar separation of the Kermadec Ridge from the Colville Ridge has formed the Havre Trough (e.g., Wright, 1993). The geology of the Lau Ridge is pertinent to the results of Leg 135 drilling inasmuch as volcanism on the ridge overlapped with the initial stages of Lau Basin extension, and the Lau Ridge volcanoes were one of the sources for volcanogenic sediments that were deposited in the extensional basins.

The oldest known rocks exposed on the Lau Ridge are the pre-middle Miocene lava flows, pyroclastic rocks, and volcaniclastic rocks of the Lau Group. Radiometric ages for the Lau Group range from 14 to 6 Ma (Whelan et al., 1985; Woodhall, 1985). Basaltic andesite and andesite, including both tholeiitic and calc-alkaline low- to high-K magma series, are the major rock types of this unit. The range in rock types also includes basalt, dacite and rhyolite. Intrusive rocks, found on several islands, represent the same magma types as the volcanic rocks (Woodhall, 1985). Lesser amounts of clastic rocks, deposited in intra-arc basins, and probable air-fall tuffs, are interbedded with flows and pyroclastic rocks. The Lau Group has the high ratio of large-ion-lithophile elements (LILE) to high-field-strength elements (HFSE) that characterize island-arc magma systems, but it lacks the corresponding signature of high LILE/REE (rare-earth elements). The $^{87}Sr/^{86}Sr$ ratio (0.7030–0.7033) is also low for island-arc-series magmas and is among the lowest of the circum-Pacific arcs (Gill, 1976). This implies that their mantle source was not significantly enriched in a "subduction component" and is a major distinction between the Lau Volcanic Group and modern lavas of the Tofua Arc, the ratios of which range from 0.70361 to 0.70399 (Ewart and Hawkesworth, 1987). Cole et al. (1985) interpret the chemistry and petrology of the Lau Group volcanic rocks as indicating the early and mature stages of volcanic arc evolution. Lau Group volcanism predated much of the crustal extension that formed the Lau Basin.

Overlying the Lau Group is the early Pliocene Korobasaga Group (4.5 to 2.5 Ma, as per Cole et al. [1985]; also reported as 5.5 to 3 Ma, by Whelan et al. [1985]). It includes flows, pyroclastic rocks, intrusive dikes and plugs, and epiclastic rocks. Although broadly similar to the Lau Volcanic Group in mineralogy, Fe-enrichment, and alkalis (i.e., they are a medium-K tholeiitic series), the Korobasaga Group has a greater abundance of basalt, as well as hornblende-bearing differentiated rocks. The Korobasaga Group also is enriched in Sr, Ba, and LREE; the $^{87}Sr/^{86}Sr$ ratio is 0.7035 (Gill, 1976). It is depleted in HFSE relative to the Lau Group (Cole et al., 1985). These authors correlate the Korobasaga Group with incipient rifting and subsidence of the Lau Ridge. Rifting of the arc was coeval with the early phase of crustal extension in the Lau Basin that began before 5.5 Ma.

The youngest volcanic series recognized on the Lau Ridge is the Mago Volcanic Group which has rocks radiometrically dated as ranging from 2.2 to 0.3 Ma and which stratigraphically overlies a late Miocene or early Pliocene limestone unit (Woodhall, 1985). The Mago Volcanics are alkaline and include both basaltic and hawaiitic varieties. They are interpreted as being related to extensional stresses that led to rifting and block faulting (Woodhall, 1985). Their eruption coincided with the time of Lau Basin extension by seafloor spreading.

Many of the volcanic islands were surrounded or partly covered by Miocene-Pliocene reef systems that are now preserved only as erosional remnants; some islands are formed only of the Miocene-Pliocene limestones without any exposures of the volcanic rocks; and some islands have exposures only of Pliocene volcanic rocks that, presumably, buried the Miocene-Pliocene reefs (Woodhall, 1985). Pliocene to Holocene reefs, some of which are emergent, are common. Thus, the reefs of the Lau Ridge record episodes of submergence and emergence during the time of arc-volcanic activity and lithosphere dilation. A similar record of vertical movements on the Tonga Ridge has been postulated to be related to thermal processes caused by subduction and lithosphere extension and to tectonic effects caused by the subduction of the Louisville Seamount Chain (Packham, 1985).

The evolution of the Lau Basin involves crustal extension and the partial dismembering of the Lau Ridge. For some backarc systems, it has been suggested that backarc and arc magmatism are not synchronous. It is important to note that in the Lau-Tonga system evidence is present that backarc extension and magmatism were contemporaneous with arc volcanism on the Lau Ridge. The 4.5–2.5 Ma basalts of the Korobasaga Group erupted during the "early rifting stage" in the evolution of the Lau Ridge (Whelan et al., 1985). The timing of this rifting and the associated volcanism, which was dominated by tholeiitic basalt, broadly overlaps the beginning of crustal extension in the western Lau Basin.

Tonga Ridge and Tonga Trench

The Tonga Ridge (Fig. 2) is a massive feature that separates the 10,500-m-deep Tonga Trench from the 2,000- to 3,000-m-deep Lau Basin. The ridge comprises two separate, nearly linear, belts that extend for about 1100 km. The eastern belt is a broad region commonly referred to as the Tonga Platform. Geomorphically, it may be subdivided into northern, central, and southern platform segments. The northern segment is deeper (e.g., north of Vava'u, depths range from 1000 to 1500 m), whereas the central and southern segments are from 500 to 1000 m deep with many emergent banks and islands. The chain of active volcanoes of the Tofua Arc is aligned along the crest of the northern segment, whereas it is offset to the west in the region south of Vava'u. A more detailed description is in Scholl et al. (1985). The segments are uplifted blocks of Tertiary platform carbonates overlying volcanic rocks that range in age from middle Eocene to late Miocene; these are capped by Quaternary reef limestones. The uplifted blocks form the larger Tongan islands and island groups. Kroenke (1984) subdivided the Tonga Platform into an (eastern) 'Eua Ridge, which exposes remnants of an Eocene and Oligocene volcanic arc, and the (western) Tongatapu Ridge, which includes the island groups of Vava'u and Ha'apai.

Drill holes and exposures show that the Tongatapu Ridge comprises mixed pelagic sediments and volcaniclastic rocks derived from a Miocene volcanic arc. Locally, there are rare intrusions of arc rocks; it is highly probable that this arc was once part of the Lau Ridge. The oldest rocks known on the ridge are the pre-late Eocene igneous basement exposed on the island of 'Eua (Ewart and Bryan, 1972; Cunningham and Anscombe, 1985; Duncan et al., 1985; Hawkins and Falvey, 1985) and the late middle Eocene rhyolitic rocks we recovered at Site 841 (Parson, Hawkins, Allan, et al., 1992; Bloomer et al., this volume).

The rocks of 'Eua include beach boulders of arc-tholeiitic hypersthene gabbro; these have been dated as 46 to 40 Ma (Duncan et al., 1985). They are overlain by upper middle Eocene calcareous conglomerates and breccias (Cunningham and Anscombe, 1985). Volcanic activity, attributed to the Lau Ridge, is recorded by Oligocene arc-tholeiitic andesitic lava flows dated as 33 to 31 Ma and Miocene dikes, also of andesitic to silicic andesitic composition (Hawkins and Falvey, 1985), dated as 19 to 17 Ma (Duncan et al., 1985). The geologic record on 'Eua offers a condensed view of the geology of the Tonga Ridge that has been developed from more extensive sections on other islands (see Cunningham and Anscombe, 1985, for a more comprehensive discussion).

The western belt of the Tonga Ridge comprises the volcanic islands, seamounts, and shoals of the Tofua Arc. The Tofua Arc volcanoes are active, and many historic eruptions have occurred. They have erupted arc-tholeiitic series magmas that range in composition from basaltic to low-K rhyolite; the main magma type has been basalt to basaltic andesite (Bryan et al., 1972; Ewart and Bryan, 1972; Ewart et al., 1977; Gill, 1981; Ewart and Hawkesworth, 1987). There is a broad similarity to the main phases of volcanism on the Lau Ridge. As discussed in a subsequent section, the age of initiation of the Tofua Arc is uncertain; it may span several million years as a result of the southward propagation of arc-volcanic edifices. The volcanic island of Niuatopatapu, at the north end of the Tofua Arc, has been dated at 3 Ma (Tappin et al., this volume), but no evidence exists for similar ages at the latitude of our drilling transect. I do not think, therefore, that it was a major feature before 1 Ma or that it is petrogenetically related to the Lau Ridge.

The Tonga Trench wall exposes a cross-section of oceanic crust that is largely formed of arc-related mafic and ultramafic rocks. Rocks dredged from the inner slope of the trench include highly depleted serpentinite, dunite, harzburgite, clinopyroxenite, gabbro, diabase, arc-tholeiitic basalts, andesite, boninite, quartz diorite, tuffs, and volcanic breccia all derived from a volcanic arc (Fisher and Engel, 1969; Hawkins et al., 1972; Bloomer and Fisher, 1987; Vallier et al., 1985; Hawkins, 1988). Some rocks indicative of the accretion of seafloor material have been dredged as well but the bulk of the material suggests that the inner slope exposes the lower levels of an island arc as originally proposed by Fisher and Engel (1969). Bloomer and Fisher (1987) proposed a 4000 m stratigraphic reconstruction, based on depths of dredge hauls, that establishes an arc-like crustal section capped by mafic and intermediate composition volcanic rocks.

The Lau Basin

The Lau Basin shares many geologic characteristics with other western Pacific backarc basins, but it also has several features that may be unique. The present axial ridges are situated about 250 km above the inclined seismic zone of the Tonga Trench, which reaches depths on the order of 700 km to the west under the North Fiji Basin (Fig. 4; Isacks and Barazangi, 1977; Billington, 1980; Giardini and Woodhouse, 1984, 1986; Pelletier and Louat, 1989). The trapezoidal basin has opened more widely at the north, with its southern end apparently hinged at the present intersection with the Louisville Seamount Chain. The mantle under the basin has low Q (strong seismic wave attenuation) and is inferred to be hotter than the surrounding mantle (Fig. 5; Barazangi and Isacks, 1971). The average basin depth is about 2500 m (Hawkins, 1974), which is anomalously shallow in view of its probable age. These depths are in striking contrast to the Mariana Trough, in which depths as great as 4500 m are common near the axial ridge and the crest of this ridge is about 3500 m deep (e.g., Hawkins et al., 1990). Interpretation of the configuration of the Tonga Trench seismic zone suggests a half-spoon shape sharply curved to the west at its north end and abruptly terminated along the west-trending part of the Tonga Trench (Fig. 6; Billington, 1980). The seismic zone has a discontinuity at about 535 km depth, which has been interpreted either as a flexure (Billington, 1980) or a result of imbrication (Louat and Dupont, 1982). The basin has a complex seafloor morphology and, as discussed in a subsequent section, only a minor part of the crust has formed by seafloor spreading (see also Parson, this volume).

Karig (1970) proposed an extensional origin for the Lau Basin in one of the first papers to address the origin of western Pacific backarc basins. This was followed by detailed studies of the geology of the Lau Basin by Hawkins et al. (1970) and Sclater et al. (1972); these described the first samples of the basaltic seafloor. These papers, and subsequent work by Hawkins (1974, 1976), established that the young crust of the Lau Basin was basaltic, showed considerable similarity to MORB, and was distinctly different from arc tholeiites. Similar conclusions for the Mariana Trough were reached by Karig (1971) and Hart et al. (1972), but they emphasized that Mariana Trough backarc basalts, although MORB-like in many respects, showed distinct enrichment in volatiles and LILE. The modern spreading ridge system of the Lau Basin, what we now call the Central and Eastern Lau Spreading Centers, was discovered and sampled on Scripps Institution of Oceanography *7-TOW* Expedition in 1970 (Hawkins et al., 1970). We ran sufficient survey lines to delineate part of the ridge (Hawkins, 1974) and to recognize symmetric magnetic anomalies, but they proved difficult to trace throughout the basin. Additional data did not help solve the problem. We concluded that true seafloor spreading had not been an important process throughout the entire history of the basin and that some form of diffuse spreading on short ridge segments may have operated (Lawver et al., 1976; Lawver and Hawkins, 1978). Thus, it became clear early on that it would be difficult to explain the entire evolution of the basin by seafloor spreading, although several attempts to do so (e.g., Weissel, 1977) appeared to give satisfactory models for some of the data. As a result of extensive bathymetric surveys with single beam and Sea Beam profiling systems, seismic reflection profiling, shipboard and aeromagnetic surveys, and GLORIA imagery, we now have a good understanding of both the regional bathymetric fabric and the reason why the magnetic data have been so difficult to interpret (Parson et al., 1989, 1990; Parson, Hawkins, Allan, et al., 1992; Parson and Hawkins, this volume).

The Lau Basin comprises several tectonic-morphologic provinces that together constitute the extensional basin, still actively opening, as shown in Figure 2. These features are discussed more fully by Parson and Hawkins (this volume). The central part of the basin, where our drill transect was located, comprises a western extensional basin-ridge province (here informally called WEB) and a triangular region of young crust that has formed by seafloor spreading. The latter area includes the two actively spreading ridges—the Central (CLSC) and Eastern (ELSC) Lau spreading centers—and the relay zone between them. These ridges are asymmetrically located in the Lau Basin, being offset toward the eastern side, and much of the southern part of the ELSC lies within 50 km of the active Tofua Arc. The CLSC is shallower than 2300 m throughout its length. In detail, ridge crest morphology varies along axis from a narrow axial rift basin to being capped with small mounds and pinnacles. The ridge is flanked by basins as deep as 2800 to 2900 m that parallel the ridge trend. The ELSC has inward-facing scarps that define an axial rift valley. Small edifices are found within the valley. The floor of this rift shoals from about 3000 m at the dying northern end to about 2300 m near its southern end at 21° S. The

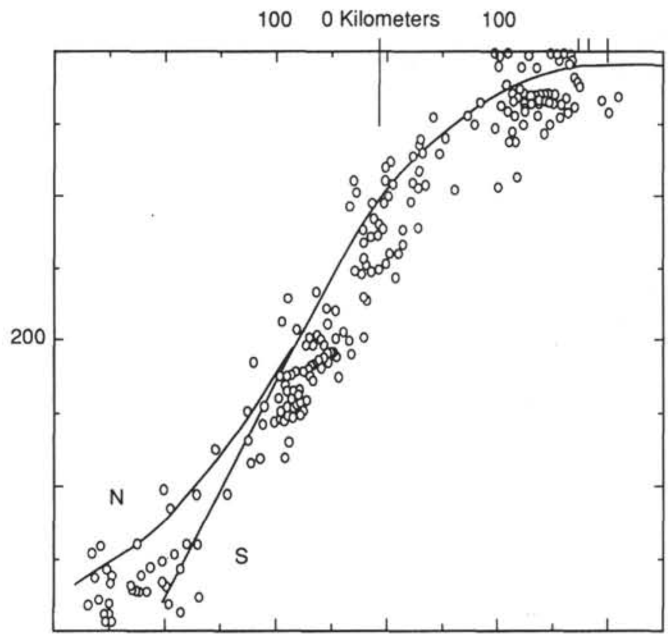

Figure 4. Composite cross-section of earthquake hypocenters for the Tonga Trench subduction zone. The "0 kilometer" mark on the horizontal axis corresponds to the location of the Tofua Arc. Locations of projections of the trench axis are shown by the vertical lines. Scale is in kilometers (Isacks and Barazangi, 1977).

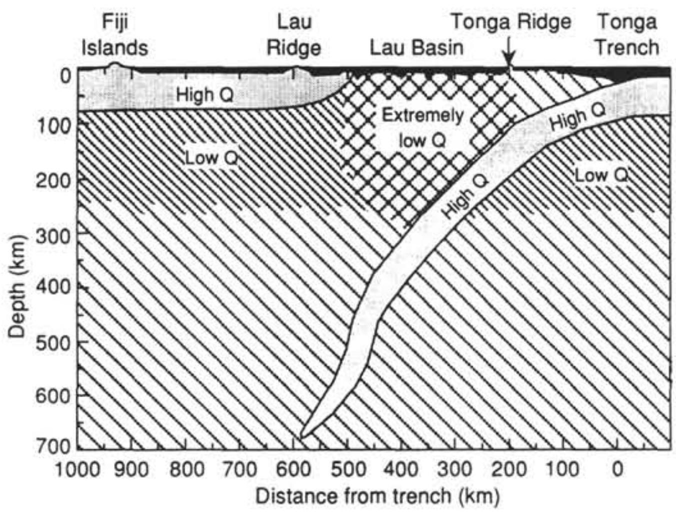

Figure 5. Cross-section of the Fiji Islands to Tonga Trench region showing inferred and extrapolated extent of the region of high and low attenuation in the uppermost mantle. Note that the Lau Basin is underlain by mantle with extremely low Q (high attenuation). From Barazangi and Isacks (1971).

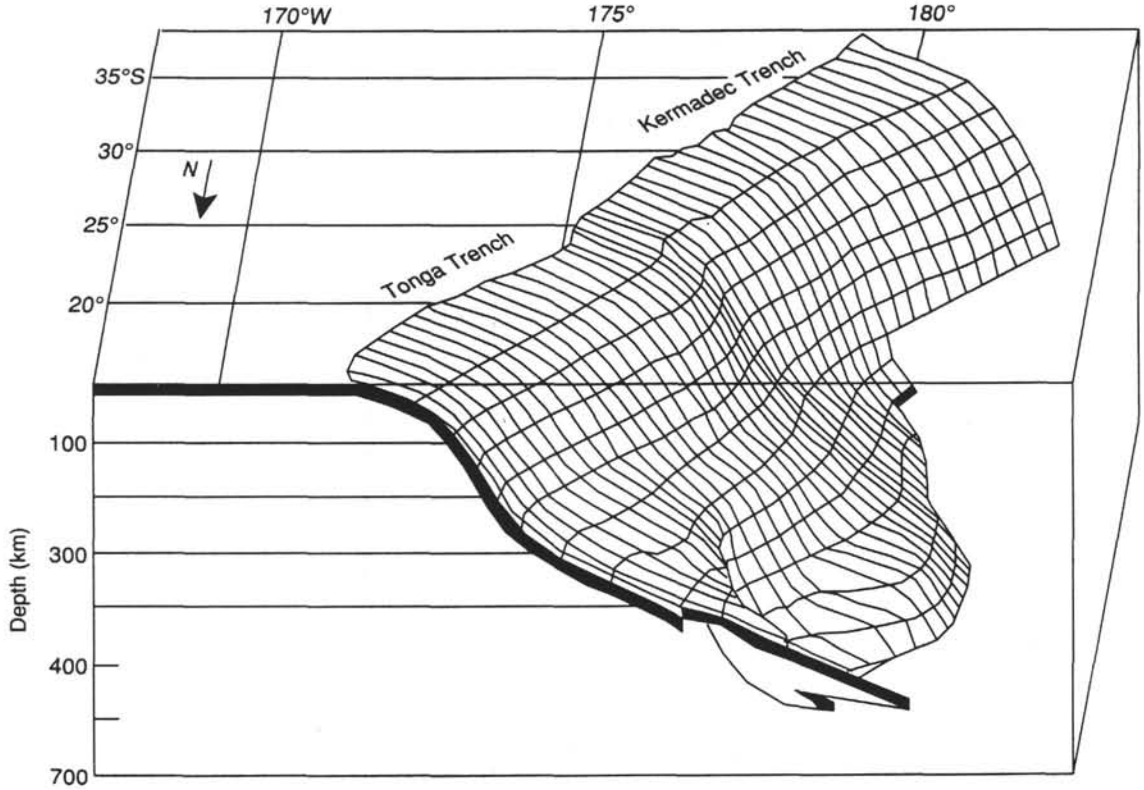

Figure 6. View, looking south, of a grid representing the upper surface of the Wadati-Benioff Zone of the Tonga-Kermadec subduction zone. The projection is slightly distorted in that it does not take into account the Earth's sphericity. From Billington (1980).

ELSC began as a propagating rift at about 5.5 to 4 Ma and was followed by the CLSC at about 1.5 to 1.2 Ma (Parson and Hawkins, this volume). Both have propagated southward from the Peggy Ridge. The two ridges form an overlapping system with the CLSC propagating southward at the expense of the ELSC. The southern end of the ELSC, the Valu Fa spreading center (Jenner et al., 1987; Vallier et al., 1991), appears to be propagating southward into older crust but essentially nothing is known about the age or composition of that older crust. Presently the northern Lau Basin may be opening at a rate of 100–110 mm/yr as determined by GPS data (see discussion in Parson and Hawkins, this volume). The complex magnetic fabric, and effects of propagating ridges, make spreading determinations difficult but it is likely that the axial ridges presently are spreading at 42 mm/yr.

The most extensively studied part of the Lau Basin is along the axial ridge systems. The ridges comprise a range in rock types that is mainly basaltic (Table 1, analyses A–G) with lesser amounts of fractionated types including Fe-Ti basalt, "oceanic andesite," and rocks approaching low-K rhyolite (e.g., Hawkins, 1974, 1976, 1977, 1988, 1989; Hawkins and Melchior, 1985; Sunkel, 1990; Boespflug et al., 1990; Ernewein et al., 1993; Falloon et al., in press). The basaltic rocks show chemical compositional variation from N-MORB-like varieties (e.g., Bryan, 1979b), which are the main rock type on the CLSC, to compositions that are transitional to arc-tholeiitic basalts (forming parts of the ELSC and flanks of the CLSC). The latter resemble many of the transitional basalts that are widespread in the Mariana Trough (e.g., Fryer et al., 1981; Hawkins and Melchior, 1985; Hawkins et al., 1990). A summary of representative rock types from the axial ridges and areas of older crust not sampled by drilling is in Table 1. The Leg 135 drilling results show that rocks from Sites 834 through 836 (Table 2) resemble those from the active ridges. At Sites 837 through 839 the rocks are more arc-like (Tables 2–3) and some are similar to modern Tofua Arc tholeiites (Ewart et al., 1977; Ewart and Hawkesworth, 1987; Ewart et al., this volume; Hawkins and Allan, this volume).

The northern part of the Lau Basin comprises at least four distinct provinces. Their southern boundary is the Peggy Ridge (Fig. 2). This high-standing feature of uncertain origin presently marks the location of many shallow earthquakes that have right lateral first motions (Eguchi, 1984; Hamburger and Isacks, 1988). It may be serving as a transform fault accommodating differential motion between the CLSC and an unnamed spreading system north of Peggy Ridge (Parson, this volume). As discussed in a subsequent section, the Peggy Ridge may have played a major role in the development of the ELSC and CLSC rift propagators. Rocks from the Peggy Ridge are MORB-like basalts that resemble the rocks of the active ridges (Hawkins, 1974, 1976, 1988, 1989; Hawkins and Melchior, 1985). Most of the area north of Peggy Ridge comprises a seamount province in which a wide range in rock types has been found including N-MORB, E-MORB, and dacite (e.g., Zephyr Shoal, Table 1). Volpe et al. (1988) and Poreda (1985) showed that many of the seamounts are anomalous for the Lau Basin in that they have isotopic signatures that indicate an affinity to the source of Samoan "plume" basalts. The eastern side of this province includes narrow basins with relatively thick sediment fill similar to those in the WEB (e.g., more than 0.2 s two-way traveltime; Hawkins, 1974), areas of high heat flow (e.g., up to 4.22 HFU; Sclater et al., 1972), and the recently active volcanic island of Niuafo'ou. The extreme northeastern part of the basin has the Mangatolu Triple Junction (Nilsson et al., 1989; Hawkins, 1989; Hawkins et al., 1989; Nilsson, 1993; Falloon et al., in press). At the Mangatolu Triple Junction, we charted and sampled a three-limbed feature with very fresh basalt on the axes of each of the limbs. Fragments of a zinc-sulfide-rich hydrothermal vent system were collected on the northeastern limb (Hawkins and Helu, 1986). Another probable spreading ridge system has been identified by backscattered sonar imagery (GLORIA) trending north-northeast from Peggy Ridge (Parson, this volume). Fresh glass-rich basalt samples from high-standing features on this presumed spreading system are MORB-like tholeiites (Hawkins, 1988).

Table 1. Rock and glass data from ELSC, CLSC, Valu Fa, and old crust.

	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q	R	S
SiO_2	50.95	52.79	52.79	49.15	52.04	50.81	53.76	56.43	55.86	56.43	50.24	53.59	50.89	49.33	51.57	55.97	50.79	55.29	63.26
TiO_2	0.95	0.92	1.85	0.99	1.24	1.94	2.07	1.53	1.2	1.25	0.76	1.05	1.32	0.51	1.11	1.04	1	1.79	0.53
Al_2O_3	15.41	15.13	14.08	16.97	14.46	13.41	13.01	14.06	14.95	14.14	15.8	14.56	17.1	11.89	16.03	14.45	17.41	13.86	14.56
FeO^*	9.36	10.47	13.62	9.19	10.7	14.1	14.44	13.14	11.32	12.19	9.13	11.55	8.01	9.07	8.38	10.35	7.28	13.36	5.33
MnO	0.18	0.19	0.21	0.18	0.2	0.26	0.24	0.2	0.22	0.22	0.21	0.21	0.17	0.16	0.18	0.21	0.15	0.23	0.09
MgO	7.52	6.52	4.28	8.48	6.71	5.63	3.92	3.91	3.87	3.72	7.39	5.27	6.35	12.91	6.23	4.4	7.37	3.57	3.42
CaO	12.64	11.47	8.9	12.14	11.37	10.21	8.34	8.53	8.32	8.14	13.37	10.22	11.06	12.29	11.77	8.87	11.86	7.85	5.16
Na_2O	2.03	1.92	3.03	2.57	2.53	2.88	3.27	1.68	2.91	2.41	2.1	2.34	3.23	1.83	3.03	2.81	2.67	2.92	4.03
K_2O	0.08	0.12	0.16	0.11	0.14	0.12	0.26	0.29	0.24	0.29	0.11	0.17	0.38	0.35	0.43	0.61	0.32	0.44	0.98
P_2O_5	0.08	0.08	0.19	0.07	0.1	0.16	0.28	0.15	0.15	0.14	0.09	0.11	0.23	0.13	0.2	0.13	0.14	0.28	0.05
Sum	99.22	99.61	99.09	99.88	99.67	99.55	99.56	99.92	99.2	99.11	99.2	99.07	98.79	98.47	98.93	98.84	98.99	99.59	97.41

ELSC: A = average of 20 samples, Mg# > 60; B = average of 7 samples, Mg# 50–59; and C = average of 2 samples, Mg# < 50.
CLSC: D = average of 5 samples, Mg# > 60; E = one sample, Mg# 50–59; F = average of 10 samples, Mg# 40–49; and G = average of 7 samples, Mg# < 40.
Old crust: H = RNDB 2-5GL, dredge site on ridge east of Site 836; I = RNDB 2-7GL, same location as H; J = RNDB 2-8GL, same location as H; K = 7-TOW 74-1, dredge site on small ridge; L = ANT 225-1, dredge site northeast of tip of ELSC; M = ANT 223-1, dredge site on ridge northeast of Site 839; N = RNDB 01-1, sparsely phyric olivine basalt, ridge west of Site 836; O = "MTB-type" basalt, 7-TOW-81, north of Site 834; P = basaltic andesite, 7-TOW 101-4, western Lau Basin Valu Fa and other rocks; Q = average of 43 glasses, Mariana Trough Basalt; R = Valu Fa Ridge, Dredge 2-3, rock; and S = Zephyr Shoal, average of two dacite vitrophyres.
Notes: References are as follows: A–H = Hawkins and Allan (this volume), I–J and N–P = Hawkins (1989), K–M = Hawkins and Melchior (1985), Q = Hawkins et al. (1990), R = Vallier et al. (1991), and S = Hawkins (1976).

Table 2. Representative rock and glass data from Sites 834 through 839.

Site	834 Unit 1	834 Unit 7	834 Unit 13	834 Prim.	834 Fract.	835 Unit 1	836 Prim.	836 Fract.	837 Ave.	838 Ave.	839 Ave.	839 BA
SiO_2	50.17	50.2	50.9	50.17	55.22	52.05	50.06	52.8	55.16	52.72	52.68	55.84
TiO_2	1.48	1.32	1.31	1.42	2.14	1.08	0.89	1.2	1.34	0.91	0.7	1.1
Al_2O_3	16.19	16.33	16.06	16.33	14.39	15.52	16	15.08	15.34	15.64	15.88	14.9
FeO^*	8.68	8.78	9.22	8.67	12.58	9.99	8.82	11.32	11.7	10.88	8.57	10.94
MnO	0.2	0.17	0.18	0.2	0.22	0.19	0.21	0.23	0.2	0.21	0.18	0.24
MgO	7	7.73	6.18	7.73	3.16	6.12	7.96	5.49	3.5	5.79	6.44	4.1
CaO	11.93	11.88	11.32	11.87	7.38	11.14	13.68	10.22	8.08	10.27	12.14	8.77
Na_2O	3.4	3	3.02	3.08	2.58	2.35	2.06	2.42	2.87	1.98	1.78	2.3
K_2O	0.12	0.1	0.14	0.09	0.23	0.21	0.07	0.26	0.69	0.34	0.31	0.49
P_2O_5	0.21	0.12	0.12	0.11	0.27	0.12	0.08	0.12	0.21	0.09	0.1	0.16
Sum	99.38	99.63	98.45	99.67	98.17	98.77	99.83	99.14	99.09	98.83	98.78	99.83

Site 834: Unit 1 = Sample 135-834A-12H-CC, 2 cm, Unit 1, Mg# = 62.3. Unit 7 = average of 17 samples, Mg# = 64.1. Unit 13 = average of 5 samples, Mg# = 57.6. Prim. = Section 135-834B-29R-1, 7–12 cm, Unit 7, Mg# = 64.6, least fractionated. Fract. = Section 135-834B-51R-1, 87–92 cm, Unit 12, Mg# = 34, most fractionated.
Site 835: Unit 1 = average of 27 samples, Mg# = 55.7.
Site 836: Prim. = Section 135-836B-7X-1, 8–14 cm, Unit 4, Mg# = 64.9, least fractionated. Fract. = Section 135-836B-7R-2, 56–62 cm, Unit 5, Mg# = 48.7, most fractionated.
Site 837: Unit 1 = average of 5 samples of sparsely phyric rock, Mg# = 38.
Site 838: Unit 1 = average of 5 samples, Sample 135-838A-9H-2, 95 cm. BA = Unit 2, Sample 135-839B-18R-1, 6 cm.
Site 839: Ave. = Unit 3, average of 6 samples.
Notes: All data from Hawkins and Allan (this volume). Site 837 data are for sparsely phyric rock; the rest are glasses. Prim. = primitive, and Fract. = fractionated.

RESULTS OF LEG 135 DRILLING

Introduction

This synthesis summarizes the extensive mineralogic, chemical, and isotope data presented in this volume by Allan, Bloomer, Bloomer et al., Bryan and Pearce, Bryan et al., Cawood and Fryer, Ewart and Griffin, Ewart et al., Forsythe and Fisk, Gaetani et al., Hawkins and Allan, Hergt and Hawkesworth, Hergt and Nilsson, McDougall, and Nilsson. A brief summary of the petrology of the volcaniclastic component of sediments filling sub-basins and deposited on the Tonga Platform is based on presentations by Bednarz and Schmincke, Clift and Dixon, Ledbetter and Haggerty, and Rothwell et al., also in this volume. In the following sections, I summarize the geologic setting and findings of major petrologic significance at each site. Additional petrographic details and shipboard chemical analyses are given in Parson, Hawkins, Allan, et al. (1992). A brief summary of the tectonic setting, and a model for the evolution of the basin, are included to provide some background for the discussion of petrologic evolution. The key to understanding the dynamic aspects of lithosphere deformation that are a major control on petrologic processes is given in the synthesis of tectonic evolution by Parson (this volume) and Parson and Hawkins (this volume).

Background

Before Leg 135 drilling, we knew little about the seafloor composition of the WEB province. A single drill hole, at DSDP Site 203, had penetrated 409 m without encountering igneous basement. The cores recovered mainly nannofossil oozes with variable amounts of arc-derived, glass-rich silts and sands. The abundance of volcanic material increased down section as did the grain size. Some cobbles of basalt were encountered near the bottom of the hole. The oldest dated sediment was middle Pliocene but a late Miocene age was inferred for the acoustic basement on the basis of sedimentation rates. Rock dredging in the western basin has been limited to a few sites on high-standing blocks that yielded rocks having compositions transitional between MORB and arc tholeiites (Table 1, analyses H, I, J, M, and N) and one locality that had arc-like andesitic rocks (Table 1, analysis K). There are no age data for these rocks; some may be remnants of the forearc to the Lau Ridge, some or all may be younger crust. Hawkins and Melchior (1985) noted the similarity between them and rocks that are common in the Mariana Trough (Table 1). They proposed that the crust was zoned on a regional scale from older rocks of transitional basalt types to basalts more similar to N-MORB that are the main rock type recovered in the central parts of the basin. They postulated a relation between this apparent regional zoning and temporal evolution of the mantle source.

Lau Basin morphology and crustal fabric was known from various single beam and multibeam surveys (e.g., Hawkins, 1974; Chase et al., 1982) and seismic reflection profiles had established that many of the sub-basins in the western Lau Basin had several hundred meters of sediment fill (Hawkins, 1974). Partial coverage with GLORIA imagery gave additional information about the regional fabric (Parson et al., 1989).

Table 3. Representative data for Tonga and Lau ridges and Site 841.

	A	B	C	D	E	F
SiO_2	53.21	54.87	58.29	55.16	51.52	77.58
TiO_2	0.51	0.71	1.11	1.14	1.05	0.22
Al_2O_3	17.09	18.3	15.32	18.08	16.33	12.04
FeO*	9.9	9.38	11.37	8.48	11.45	1.97
MnO	0.18	0.21	0.2	0.17	0.24	0.07
MgO	5.37	5.36	4.63	3.62	4.67	0.26
CaO	11.45	8.07	4.84	8.54	9.59	1.71
Na_2O	1.57	2.58	4.08	3.3	3.09	4.46
K_2O	0.34	0.2	0.15	0.84	0.28	1.17
P_2O_5	0.07	0.08	0.05	0.25	0.16	0.03
Sum	99.69	99.76	100.04	99.58	98.38	99.51

A = Tofua Arc, average basaltic andesite; B = Tonga Ridge, 'Eua island, andesitic flow; C = Tonga Ridge, 'Eua island, andesitic flow; D = Lau Ridge, average basic andesite of Lau Volcanic Group; E = Site 841, average arc tholeiitic intrusives, Unit 1; and F = Site 841, average low-K rhyolite, Subunit 2B.
Notes: References are as follows: A = Ewart and Hawkesworth (1987), B–C = Hawkins and Falvey (1987), D = Cole et al. (1985), and E–F = Bloomer et al. (this volume).

Petrologic Evolution of the Lau Basin: A Synthesis

Leg 135 drilling was done mainly in sub-basins of the WEB that record the early history of the Lau Basin during the extensional phase of its evolution (Fig. 2, and Hawkins et al., 1991). The modern spreading centers are an expression of the second phase which was dominated by propagating rifts that formed the CLSC and ELSC (Parson, Hawkins, Allan, et al., 1992; Parson, this volume). For these, we have data from one drill site (Site 836) and extensive data from conventional surveys. The drilling has given new insight to the early history of the basin by providing samples from beneath the sediment fill and by giving biostratigraphic data to establish a chronology of magmatic and tectonic events. Holes were drilled at six sites that, together with the extensive data from rock dredging, afford a three-dimensional view of the nature and complexity of the Lau Basin crust.

All of the Leg 135 backarc sites, as well as DSDP Site 203, were drilled in small, partly sedimented basins, elongated in a predominantly northerly direction, separated by north-trending ridges. Local relief typically is on the order of 100 to 200 m. These sites sampled successively younger backarc crust, from west to east, ranging from at least 5.5 Ma (Site 834) to about 0.6–0.8 Ma (Site 836). Age estimates were made by paleontologic dating of sediments overlying or interlayered with basalt flows, and by paleomagnetic determinations (Parson, Hawkins, Allan, et al., 1992; Chaproniere, Chaproniere et al., and Sager and Abrahamson, all in this volume). The age estimates are summarized in Figure 7. Sites 834, 835, 837, 838, and 839 were drilled on crust that we assign to the WEB and Site 836 is on crust formed early in the spreading history of the present ELSC. Our interpretation of the history of the WEB is based on cores from five sub-basins and a few dredged sites on the intervening ridges. The dredged rocks from the WEB are mainly transitional between arc and MORB and were termed Mariana Trough-type basalts by Hawkins and Melchior (1985). As discussed elsewhere in the text, the drill cores are different from the dredged rocks in being either MORB-like and "old," or arc-like and relatively younger than the intervening ridges. None of the sample sites penetrated old arc or forearc crust. The basins we drilled are insets within older crustal rocks and not part of a continuous crustal fabric that might have formed at a spreading ridge. From these samples, magnetic and seismic reflection data, and bathymetry, we have inferred that the WEB is largely rifted older crust, probably the forearc to the Lau Ridge, rather than "new" crust formed by seafloor spreading. Our interpretation of the geology at each drill site supports this interpretation. Clearly, there is much that remains to be learned about the WEB, especially concerning the geology of the interbasin ridges, however, the simplest explanation is that crustal extension formed all of the WEB and seafloor spreading is restricted to the narrow bands near the present axial ridges.

Petrologically, the drill sites may be separated into those with rocks similar to MORB and the present CLSC and ELSC (Sites 834, 835, and 836); rocks transitional in chemistry between MORB and arc (Sites 837 and 838); and rocks similar to those from the modern Tofua Arc (Site 839). Representative analyses are in Table 2. Several general points may be made about the Leg 135 samples that also apply to the dredged rocks. Except for Site 839, the crust of the Lau Basin is, for the most part, distinctly different from the rocks typical of island arcs. Their mineralogy and chemistry are more like MORB, as defined by Bryan et al. (1976), Melson et al. (1976), and Viereck et al. (1989) than any other basalt type. However, few of the data make a perfect match to MORB for all element abundances. All samples have at least some element abundances that hint of an arc affinity, even though few can be considered arc-like. This must reflect varied mixing of the SSZ component with true MORB constituents in all samples. Water, presumably of primary origin, is one of the more significant constituents in all of the Lau Basin rocks. This is expressed in the prevalence of highly vesicular rocks in the cores and the dredged samples, in measured high water in glasses, and to enrichments or depletions in some major elements relative to MORB that may be explained by the high-water content of the parental melts.

It has long been recognized that many backarc basin magmas are depleted in Fe and Ti, but enriched in Al and Na, relative to MORB at any given level of Mg (e.g., Fryer et al., 1981; Hawkins and Melchior, 1985). The enrichment in Al is a major distinction that has been shown to be an intrinsic property of the melts rather than due to accumulation of plagioclase. The most likely explanation is that this is the result of hydrous melting and petrologic evolution similar to that for island arcs (e.g., Baker and Eggler, 1987; Sisson and Grove, 1990a, 1990b). Many circum-Pacific arc systems have low-MgO, high-alumina basalts (e.g., >17% Al_2O_3, <54% SiO_2, <6% MgO). These rocks are not readily explained as direct partial melts of the mantle or subducted slabs, or the result of fractional crystallization of anhydrous parental basalts. However, they may be differentiates of parental basalts that had 2%–4% H_2O (Sisson and Grove, 1990b). Experimental studies on hydrous mafic melts (e.g., Baker and Eggler, 1987) suggest an explanation for the Al enrichment in arc magmas. The higher water content in the parental magma systems shifts phase relations and suppresses plagioclase crystallization (see also Michael and Chase, 1987; Bryan et al. and Gaetani et al., both in this volume; and the previous discussion about vesicularity). Addition of water to basaltic melts expands the primary phase volumes of both olivine and high-Ca pyroxene, shifting melt compositions toward plagioclase in the CMAS system, and lowering the crystallization temperature of anhydrous silicates (Sisson and Grove, 1990a). These authors point out that plagioclase, olivine, and Ca-pyroxene appear at high melt fractions close to the liquidii of high-alumina basalts. The high-water content destabilizes plagioclase as an early crystalline phase and increases the proportion of Fe-Mg silicates. Melting experiments on Leg 135 samples gave results similar to those for island-arc material (Bryan et al. and Gaetani et al., both in this volume) that demonstrate a shift in compositions toward plagioclase in the CMAS system. Primary melts are relatively enriched in Al, and plagioclase crystallization is delayed, thereby giving rise to high-Al melts. However, plagioclase is an abundant phase in all Lau Basin samples occurring as phenocrysts, a groundmass constituent, and xenocrysts. The most Mg-rich rocks and glasses (up to 8% MgO) have plagioclase, thus plagioclase was not suppressed in the erupted magmas. The parental melts may have had water contents high enough to delay the onset of plagioclase crystallization at higher pressures. Magma ascent caused rapid exsolution of water (vesiculation) at low P, from melts saturated with plagioclase, and promoted growth of plagioclase (Sisson and Grove, 1990b).

The elevated water content probably caused higher fO_2 and led to more oxidized conditions thereby depleting the melts in Fe and Ti. Nilsson's studies (this volume) on Site 834 samples indicates that they are more oxidized than N-MORB; this could be an important factor in controlling Fe and Ti in evolving melts.

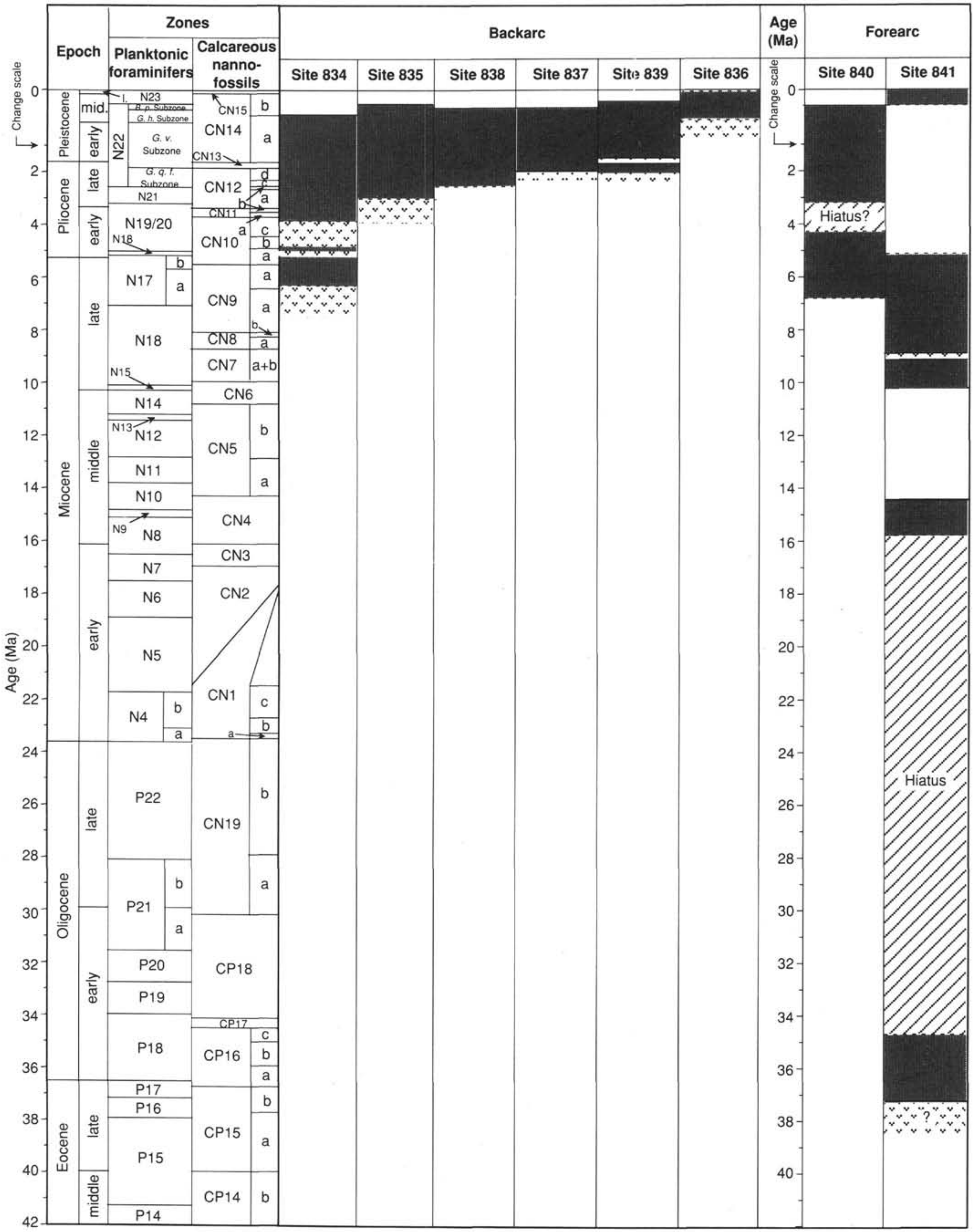

Figure 7. Biostratigraphic correlations for Leg 135 drill sites, based on planktonic foraminifers and calcareous nannofossils, showing the age constraints on minimum ages for the volcanic rocks at each site. Sediments are shown in black and igneous rocks are shown as V pattern. From Parson, Hawkins, Allan, et al. (1992).

Site 834

Site 834 was selected to give us information about the earliest stages in the opening of the Lau Basin. Here we recovered the oldest samples of Lau Basin seafloor from one of the numerous elongated half-grabens that form sub-basins in the WEB province. The drill site is about 100 km east of the Lau Ridge remnant arc in water depths of about 2700 m. Crust formed by the ELSC lies about 135 km to the east; the propagating rift tip of the ELSC would have passed the latitude of this site at about 3.6 Ma. The presently active CLSC is about 150 km to the east. The drilling penetrated about 106.2 m of sediment before reaching a basalt flow in Hole 834A and 112.6 m to basalt in Hole 834B. Drilling continued through a sequence of flows with minor sedimentary interbeds that extended to depths of at least 348.8 m below seafloor (mbsf). Probable continuous igneous basement was cored to a depth of 435.5 mbsf. Paleontologic dating of sediment layers interbedded with basaltic flows indicated that crustal extension and magmatism was underway more than 5.5 m.y. ago. The oldest dated sediment is a thin layer of calcarenite with foraminifers, at 310.6 mbsf, assigned to Subzone N17b (late Miocene or about 5.5 Ma). The youngest igneous rocks are from a probable flow overlain, at 112.5 mbsf, by sediments with fauna assigned to nannofossil Subzone CN11b, or older. This indicates a probable early Pliocene age or about 3.8 Ma as a maximum age for the youngest flow (Styzen, this volume). The sediment fill above the uppermost basaltic unit consists mainly of volcaniclastics with thick turbiditic volcanic sands and silts. The volcanic sands decrease upsection and deposits younger than about 2.4 Ma are almost entirely hemipelagic nannofossil ooze with interbeds of vitric ash and volcaniclastic sediments (Parson, Hawkins, Allan, et al., 1992; Rothwell and Parson, this volume). The sedimentary record is important as it records continuing arc volcanism coeval with Lau Basin extension. Evidence presented by Rothwell et al. (this volume) shows that the probable source was nearby *intrabasin* sources rather than the Lau Ridge, even though it was active until about 2.5 Ma.

Thirteen igneous units were recognized at Site 834 on the basis of mineralogy, texture, and the presence of sedimentary interbeds. All of the units recognized are MORB-like in composition, although they carry some subtle imprints of derivation from SSZ mantle. Mineralogically, they are olivine-plagioclase-clinopyroxene basalts with minor Fe and Ti oxides, and rare Fe-Cu sulfides; some samples have minor Cr-spinel. Nearly all samples have all three silicate phases as constituents of the groundmass (Bryan et al., Hawkins and Allan, Hergt and Nilsson, all in this volume). Microprobe and proton-probe analyses of the mineral constituents show that they are like those found in MORB (Ewart and Griffin, Hawkins and Allan, Hergt and Nilsson, all in this volume). Phenocryst proportions and abundances are varied and give rise to aphyric, sparsely to moderately phyric, and less common highly phyric textures. In most of the phyric samples plagioclase is the predominant phase with lesser amounts of one or both of the other silicates.

A detailed study of plagioclase zoning in Site 834 samples using Nomarski differential interference contrast microscopy was made by Bryan and Pearce (this volume). The zoning pattern in most of the Site 834 samples is much more simple than the intricate complex zoning found at some of the other sites and in arc lavas. They conclude that most of the Site 834 samples are typical of MORB but some have complex zoned cores more like the zoning developed in arc magmas. Mixing between arc and MORB magmas is a possible explanation for the plagioclase zoning and supporting evidence is found in Cr-spinel. Allan (this volume) presents data for Cr-spinel from Units 1 and 7 that are typical of MORB spinels, and he interprets these data as indicating similarity in host magma composition, mantle sources (eg. spinel lherzolite) and crystallization conditions. Groundmass spinel in Unit 1 shows compositional zonation. Cores of Cr-spinel crystallized from a melt significantly more rich in Mg than their rims and host rocks and must represent effects of magma mixing. Spinel in Unit 7 is generally similar to that in Unit 1 but the nature of zoning is different. It may reflect reequilibration with interstitial Fe-rich liquids during solidification.

Some representative compositional data for the youngest unit sampled (Unit 1), an average composition for the oldest (Unit 13), and an average for the most "primitive" (Unit 7) are given in Table 2. Unit 7 may be the best example we have of a probable parental magma. Also shown are one of the most fractionated samples and one of the most primitive. The units sampled show varied effects of low-pressure fractional crystallization but essentially all are low-K tholeiites or their differentiates. Silica content of glasses ranges from 49% in a high-Mg tholeiite of Unit 7 to 56% in basaltic andesite from Unit 12.

Major element data, plotted relative to MgO content, show that the Site 834 samples closely resemble glass compositions from the CLSC and many also overlap with the fields for Mariana Trough glasses (Hawkins and Allan, this volume). These data are paralleled by trace element data. One of the most significant findings at Site 834 was that the basalts sampled have LILE and HFSE abundances and ratios that resemble N-MORB (normal mid-ocean ridge basalt) in spite of the proximity of the site to the Lau Ridge and eruption ages that are coeval with arc magmatism. The REE data and chondrite-normalized patterns also support this interpretation. Hergt and Nilsson (this volume) give LILE data (Cs, Rb, K, Ba, Pb, and Sr) for glasses, all of which show enrichments relative to N-MORB, whereas the more compatible elements (HFSE and REE) are approximately like MORB abundances. Trace element data for each of the units recognized, normalized to N-MORB, are shown in Figure 8. Note the MORB-like HFSE abundances for the least fractionated samples and the relative enrichment in LILE. The depletion in Nb and Ta is a characteristic of all of the Lau Basin samples, and we consider it to be the signature of the SSZ mantle source.

Noble metal analyses (Cawood and Fryer, this volume) are interpreted as being MORB-like and reflect mantle-derived compositions not affected by low-P fractionation, alteration, or separation of immiscible sulfide liquids. However, Site 834 samples, as well as those from Sites 835, 836, and 839, all show high levels of Au. Platinum group elements (PGE) have strongly fractionated ratios and very low abundances of Rh and Ir. The anomalously high Au yields Au/Pd ratios for Site 834 samples that average 4.98; this is significantly higher than typical MORB values, which are close to 1. They suggest that the high Au/Pd ratios may be the result of the incorporation of a hydrothermal component (either fluids or derived precipitates) into the melt source. A small addition of such a component from subducted crust could give the observed ratios.

Unit 7, cored between 214.3 and 288.0 mbsf, is one of the more massive units and also the most "primitive." The age of Unit 7 is estimated to be about 5.5 Ma (latest Miocene to early Pliocene) from paleontologic dating of the thin sediment layers that bracket the unit. The upper layer, overlying igneous Unit 6, has nannofossils indicative of Zone CN10a and foraminifers of Zone N18. The lower sedimentary bed, at the base of Unit 7, has poorly preserved nannofossils identified as Pliocene-Miocene; no foraminifers were found. A maximum age of about 5.7 Ma for Unit 7 is indicated by the presence of Zone N17b foraminifers and early Pliocene–middle Miocene nannofossils in the sedimentary layer beneath Unit 8 about 22 m lower in the core. The glasses and rocks of Unit 7 have the least fractionated compositions and are the closest to N-MORB, when compared with the other drill-core data from Site 834. Some representative data and element abundance patterns for Unit 7 and other Site 834 units are in Table 2 and Figure 8. Most of the trace element abundances are nearly identical to N-MORB; however, the noble metal data show high-Au values. The Au/Pd ratios for two samples are 18.6 and 5.2, both well above MORB values of 1. Rocks of Unit 7, in particular, as well as other high-Mg samples from Site 834, probably represent a lesser extent of melting of a MORB-source mantle relative to other backarc samples, including those from the axial ridges. This inference is based on the higher Ti and Na in Unit 7 at comparable Mg content. Glasses from the deepest samples are from a moderately fractionated unit (Mg# =

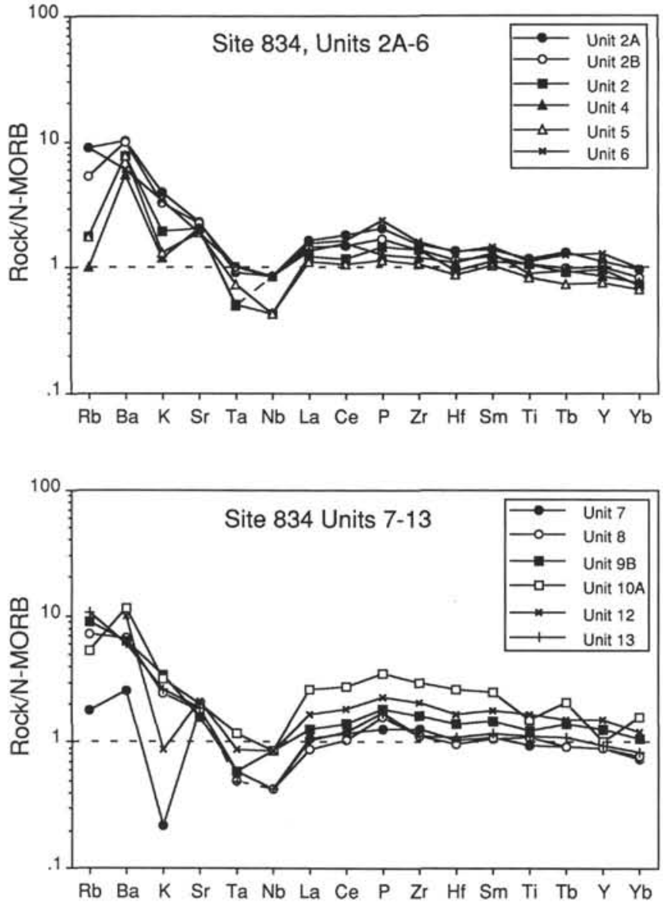

Figure 8. Trace elements normalized to N-MORB values, for Site 834. Normalizing values from Sun and McDonough (1989). Note the relative depletion in Nb and Ta and enrichments in LILE that characterize SSZ magmas. From Hawkins and Allan (this volume).

56.8 to 58.1), implying that there may be still deeper layers with accumulations of liquidus phases and rocks closer in composition to parental melts.

Chilled margins of flows or pillows are present in all but Unit 5, an aphyric diabasic textured unit. Glass compositions range from relatively unfractionated types with Mg# = 64.8 to Mg# = 34 (where Mg# = 100 * [atomic ratio Mg/(Mg + Fe^{2+})] and Fe^{3+}/Fe^{2+} = 0.15). No samples can be considered to be truly primitive melts (e.g., Mg# > 65), although many data are available that show relatively unfractionated samples with Mg# = 60 to 65 (e.g., Unit 7). Using the glass data as representative melt compositions, it is possible to investigate temporal variation in melt chemistry. The data show extensive differences through time that may be explained largely as the result of fractional crystallization.

Bryan et al. (this volume) discuss magma evolution using projections in the CMAS system and conclude that low-pressure fractionation of multiply saturated melts has given rise to the range in composition. The data, projected from plagioclase, define a linear band along the olivine/clinopyroxene cotectic. The same data, projected from silica, appear to follow a control line for evolved samples that is shifted toward plagioclase, consistent with equilibration at increased water content. Experimental studies of Site 839 samples at varied water content (Gaetani et al., this volume) support this argument.

Hergt and Nilsson (this volume) emphasize that although fractional crystallization can explain some of the LILE enrichments, it is inadequate to explain the abundances or the element/element ratios fully. They invoke another process such as magma mixing and propose that MORB and arc end-members may be involved. Mixing is suggested by the plagioclase and spinel data cited in a previous section, but the most compelling evidence comes from isotope data presented by Hergt and Hawkesworth (this volume). Some Site 834 units may be nearly "pristine" N-MORB (e.g., Unit 7), whereas others such as Units 2 and 8 may be mixtures of nearly equal amounts of the end-members. The isotope data for Site 834 (Hergt and Hawkesworth, this volume) gave new insight to the history of the basin and the evolution of the mantle source for magmas. It is established that the melts erupted on the CLSC and ELSC carry the isotopic signature of the Indian MORB-source (Volpe et al., 1988; Loock et al., 1990). This observation gave supporting evidence for the proposal that MORB-source mantle is streaming in above the Waditi-Benioff Zone (e.g., Hawkins et al., 1984; Hawkins and Melchior, 1985). However, as shown by Hergt and Hawkesworth (this volume), the isotope data for Site 834 samples are dominated by a source like that for Pacific Plate MORB. This implies that the early Lau Basin eruptions were fed by melts derived from a remnant of Pacific lithosphere trapped behind the Tonga Trench. As discussed in a subsequent section, the Pacific isotope signature shows up in other WEB province drill core samples.

The rocks of Site 834, and nearly all Lau Basin samples collected by drilling or dredging, are vesicular in spite of probable eruption depths of 2600 m or deeper (e.g., Hawkins, 1976). This indicates that their volatile content, probably largely water, was higher than MORB. Site 834 vesicle contents range from essentially zero in Unit 7 to as high as 40 to 50 modal percent in Units 1, 4, 6, and 8. Many of the larger vesicles are filled with dark colored, aphyric, highly vesicular, quench textured material. This material also occurs as vein fillings and patchy segregations. Bloomer (this volume) interprets these as segregation vesicles and veins formed by accumulations of residual liquids that migrated from the mesostasis into voids. They are present in seven of the thirteen units recognized at Site 834 as well as at Sites 835, 837, 839 and in many of the dredged rocks from the axial ridges.

Textural evidence for the high volatile content of backarc magmas supports the observations that backarc basin magmas have higher water content than otherwise similar MORB (e.g., Garcia et al., 1979). Relatively high fO_2 in backarc basin magmas has also been proposed (e.g., Hawkins and Melchior, 1985; Sinton and Fryer, 1987; Hochstaedter, Gill, Kusakabe, et al., 1990). The first comprehensive study of Fe^{3+}/Fe^{2+}, fO_2, and S content for backarc basin samples is given in this volume by Nilsson. She shows that, considering all of the data, there is a spectrum in fO_2 and S from MORB-like values to hydrous, high-fO_2, low-S, degassed samples, with the latter being more typical of arc magmas. In spite of other MORB-like chemical characteristics, most of the Site 834 glasses differ from MORB in showing no correlation between S and FeO^*. There is a negative correlation with S for LILE and fO_2. Unlike MORB, no correlation exists between fO_2 and MgO. Subunit 6A and Unit 7 are the only Site 834 samples having immiscible sulfide droplets, and these units have Fe-S relations that lie on the empirically derived sulfide saturation line for MORB. The calculated fO_2 values are within 1 log unit of the NNO buffer in contrast to MORB, which typically is 2 to 3 log units below NNO. Nilsson proposes that the strong correlation between LILE abundances, $^{87}Sr/^{86}Sr$ ratios, and fO_2 is a result of the influx of a hydrous fluid into the SSZ mantle source during subduction.

Site 835

Site 835 occupies a fault bounded half-graben, more than 2900 m deep, approximately 200 km east of the Lau Ridge remnant arc and about 90 km west of the axis of the CLSC. Crust that probably was formed by the ELSC lies about 35 km to the east. The site is on the eastern flank of a broad shoal, rising to about 1650 km depth, that lies on the regional north-south bathymetric grain. Drilling at Site 835 penetrated 155 m of nannofossil ooze, vitric ash, and volcaniclastic sediments before reaching basaltic rocks presumed to be igneous basement. Paleontologic dating with calcareous nannofossils assigned the lowest sediment samples to Subzone CN12a indicating a late Pliocene age of 3.5 Ma (Styzen, and Rothwell et al., both in this volume) for the end of volcanism in this basin. Magnetostratigraphic dating gave a similar age (Chaproniere et al., this volume). The propagating rift tip of the ELSC would have passed the latitude of the site at about 3.6 Ma. Deposition of thick volcaniclastic turbidites continued until about 2.9 Ma (Rothwell et al. and Parson et al., both in this volume).

We have no information about the age of the beginning of basalt eruption in this basin, but it ended about 0.3 Ma later than the last eruptions at Site 834 some 200 km to the west. This suggest that the termination of volcanic activity migrated eastward at about 16 cm/yr.

The rocks sampled at Site 835 are multiply saturated olivine-plagioclase-clinopyroxene basalts showing a wide range in textures from sparsely phyric to highly phyric and diabasic. Glass-rich chill margins from pillow rims or flow surfaces are common. The samples comprise varied proportions of the three main silicate phases; plagioclase is the most abundant mineral, being about twice as abundant as clinopyroxene. The rocks typically have a high vesicle content (e.g., 8%–35%) except for 2 to 3 cm wide zones, adjacent to chill margins. The magmas were erupted at 2900 m water depth; thus, the high vesicularity suggests a high volatile content. Vesicle size distribution is bimodal with from 5% to 15% comprising small (<0.5-mm diameter) vesicles uniformly distributed through the groundmass. Abundant microvesicles are present in parts of the core. Irregular, globular patches (up to several centimeters in diameter) of highly vesicular basalt (up to 70% or 80% vesicles) probably represent segregation vesicles filled with volatile-rich residual melt (Bloomer, this volume). All of the samples from Site 835 were assigned to a single lithologic unit on the basis of their mineralogical and chemical similarity (Table 2).

The chemistry of Site 835 basalts is more similar to MORB than to any other basalt type, but shows minor deviations toward arc-like characteristics. The time of the beginning of magmatism at this site is not known, but it ended shortly after the estimated time of passage of the propagating rift tip of the ELSC (Parson and Hawkins, this volume); therefore, a similarity to ELSC magmas would not be unexpected. Data fields for Al_2O_3, FeO^*, TiO_2, and Na_2O relative to MgO overlap with fields for the ELSC; these are displaced slightly toward arc values from the MORB fields (Hawkins and Allan, this volume). CaO and K_2O show little difference between the ELSC or MORB. All Site 835 samples are moderately fractionated low-K tholeiites. Glasses all have about 1.1% TiO_2 and 0.2% K_2O; they range from 6.0% to 6.2% MgO and from 2.2% to 2.4% Na_2O. The Mg# ranges from 54.9 to 56.3. The trends for TiO_2 and FeO^* vs. MgO also are MORB-like, although they too are displaced to slightly lower values at a given Mg content. The ratio CaO/Al_2O_3 is about 0.7 for all samples. This is a typical value for N-MORB and indicates that clinopyroxene crystallized later than plagioclase. The inferred crystallization sequence, olivine-plagioclase-clinopyroxene, is the MORB sequence. Figure 9 shows trace element data normalized to N-MORB. The data are limited but the same general pattern seen for Site 834 is developed with these rocks. A similar plot presented by Hergt and Hawkesworth (this volume) shows a shape similar to that for Valu Fa Ridge samples but having lower ratios. The trace elements and, in particular, LILE, Nb, and Ta, all are transitional to an SSZ or arc-like signature.

Cawood and Fryer (this volume) analyzed one sample for noble metals and, as for the other backarc sites, found an unusually high-Au content relative to PGE (e.g., Au/Pd = 5.95). They claim a strong fractionation for PGE but report Ir as zero and gave no ratios. In spite of the anomalously high-Au levels, Cawood and Fryer (this volume) conclude that the noble metals are more like MORB than arc.

The Sr- and Nd-isotope ratios indicate that the magma sources for Site 835 appear to have been generally MORB-like and that they lie within the Indian MORB field. The Pb-isotope ratios show more of an arc component than do the Site 834 samples (Hergt and Hawkesworth, this volume).

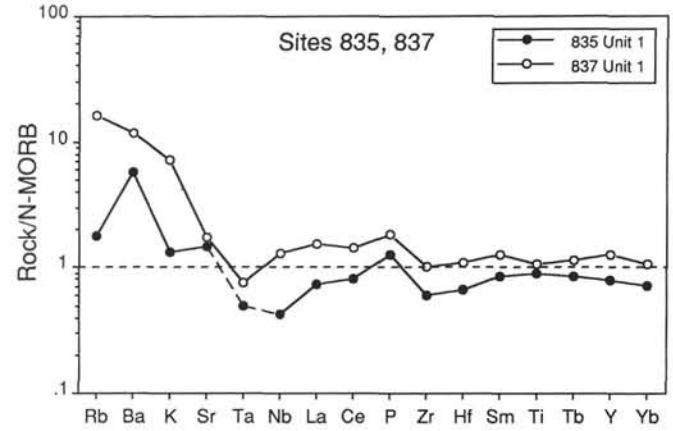

Figure 9. Trace elements normalized to N-MORB values for Sites 835 and 837. Normalizing values from Sun and McDonough (1989). From Hawkins and Allan (this volume).

Site 836

Site 836 is in a small oval shaped basin, elongated in a north-northeast direction. It is on relatively young crust that we interpreted as having been formed by the ELSC propagating rift. The overlying sedimentary sequence was sampled only through the mid-Pleistocene which was reached at a depth of 20.8 mbsf. The crustal age is well constrained by paleontologic dating of intercalated sediments and is estimated to be about 0.64 Ma. Magnetic anomaly data indicate that the ELSC, at this latitude, is no more than about 0.8 m.y. old. As we interpret the crust at Site 836 as having formed there, its age must be between 0.8 and 0.64 Ma. The present axis of the ELSC is about 48 km to the east and the southward propagating tip of the CLSC is about 100 km to the north. In contrast to the other backarc sites, Site 836 is on crust formed by seafloor spreading. Mineralogical, chemical, and isotopic data discussed below all point to a MORB-like composition for the crust at Site 836.

The presumed igneous basement comprises sparsely to moderately phyric tholeiitic basalts having phenocrysts of plagioclase, clinopyroxene and olivine typically in that order of relative abundance. Cr-spinel is present in trace amounts. Allan (this volume) found that the Cr-spinels are typical of MORB as discussed above for spinels from Site 834. Microprobe studies show that MORB-like silicate mineral assemblages are present (Bryan et al. and Hawkins and Allan, both in this volume). Two sparsely phyric andesitic hyaloclastite units, interbedded with ash layers, overlie the presumed basement. Site 836 rocks probably were erupted in about 2600 m water depth, however, like rocks at the other backarc sites, they are highly vesicular. Rocks of the presumed basaltic basement have from 15 to 40 modal percent vesicles and the two units intercalated with ash layers have up to 45 modal percent vesicles. Some of the vesicles are filled with highly vesicular, quench-textured mesostasis material (Bloomer, this volume). The deepest basement unit recognized is quartz normative tholeiite; two basement units above it are both olivine normative. The Mg# of glasses from the basement units range from 49 to 71, nearly all are higher than 65 and argue for them being relatively unfractionated magmas (Hawkins and Allan, this volume). Data for one of the most primitive samples and the most fractionated sample are in Table 2. All are low-K tholeiites (most have << 0.2% K_2O) and nearly all have less than 1% TiO_2 (the range is from 0.7% to 1.2%). Trace element data, normalized to N-MORB (Fig. 10), show a close parallel with rocks from the CLSC and ELSC. That is, they are MORB-like yet carry the variable, but slightly elevated, LILE signature of the SSZ magmas (Hawkins and Allan, this volume). As with the other backarc data, there is a relative depletion in Nb and Ta.

Bryan et al. (this volume) discuss the evolution of Site 836 samples and, using the CMAS diagram, show how the data lie along the 1-atm, multiply saturated, olivine-clinopyroxene phase boundary. Magma

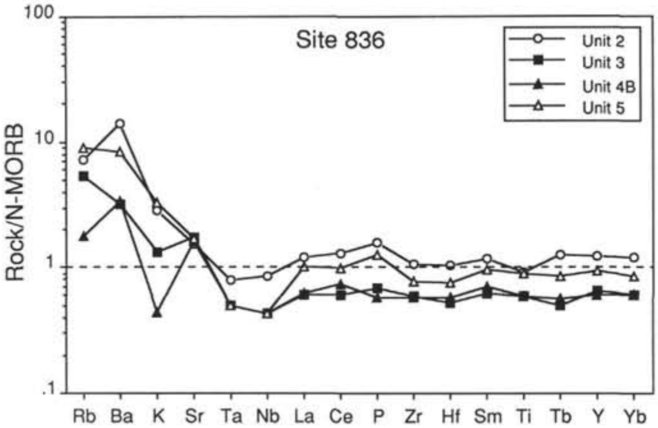

Figure 10. Trace elements normalized to N-MORB values for Site 836. Normalizing values from Sun and McDonough (1989). From Hawkins and Allan (this volume).

evolution along the calculated liquid line of descent can explain the range in compositions. They note that there is some offset from MORB multiple saturation boundaries, which they attribute to the elevated water content of these magmas.

Noble metal analyses (Cawood and Fryer, this volume) show that Au is anomalously high and "arc-like." The PGE are strongly fractionated (Pd/Ir is high; e.g., 40 to 88) as are the limited data for Indian Ocean MORB. Unlike MORB, which has an Au/Pd ratio of about 1, the average Site 836 Au/Pd ratio is 4.98, reflecting the high levels for Au. They note that high-Au contents have been reported from hydrothermal vent material at the Valu Fa Ridge at the southern tip of the ELSC, and, as discussed previously, they suggest that the high Au in Site 836 samples may be a signature of subducted hydrothermally derived sediment.

Sites 837 and 838

Sites 837 and 838 are both located on the eastern edge of the WEB in an area of narrow, partly sedimented troughs that trend northeasterly. Site 837 is about 69 km west, and Site 838 is about 87 km west of the axis of the ELSC. More extensive descriptions are in Parson, Hawkins, Allan, et al. (1992).

Igneous rock recovery at Site 837 was limited to about 4 m of sparsely to moderately phyric clinopyroxene-(orthopyroxene)-plagioclase basaltic andesite. The age is estimated to be about 2 Ma (Parson, Hawkins, Allan, et al., 1992). Orthopyroxene is present only in trace amounts; plagioclase phenocrysts are more abundant than clinopyroxene, but both rarely constitute more than about 2 modal percent. The samples are highly vesicular with from 15% to 20% vesicles. Some of the vesicles are filled with quench-textured, highly vesicular material (e.g., segregation vesicles, as discussed previously). Some representative compositional data are in Table 2, and chondrite-normalized trace element data are in Figure 9. Although they have some similarity to arc magmas (Table 3), the REE and the expanded trace element patterns show many similarities to the more evolved lavas from Site 834. The HFSE are MORB-like, whereas the LILE are enriched to as much as 10 × MORB values, and they have the SSZ depletion in Ta (but not Nb). Site 837 samples are fractionated, and we have no clear idea of their parental magmas. Assuming that olivine was the main phase removed, they are best described as having chemical signatures transitional between MORB and arc; therefore, they may have been derived from parental melts that resemble Mariana Trough backarc basin samples (Table 2). Hergt and Hawkesworth (this volume) present isotope data showing that the Pb-isotope systems overlap the fields for the Tofua Arc, but that Sr and Nd lie in the field for the modern Lau spreading systems and the Indian Ocean MORB. They interpret the Sr and Nd data as being similar to Valu Fa Ridge values and infer that, inasmuch as the Site 837 magmas were erupted when the ELSC propagator tip was nearby, the tectonic setting was comparable with that at the present Valu Fa Ridge.

Igneous rock recovery at Site 838 was limited to about 1.7 m of gravel, with clasts of aphyric to sparsely phyric clinopyroxene-plagioclase basalt. Fragments of rhyolitic ash also were recovered. A minimum age for the gravel of 1.8 Ma was estimated from the biostratigraphic data (Parson, Hawkins, Allan, et al., 1992). Some of the basaltic fragments are highly vesicular, with as much as 40%–60% tiny vesicles, which give the rocks a spongy texture. Some clasts have larger vesicles (e.g., up to 2.5-mm diameter) forming up to 10 modal percent. Compositional data are limited, but an average for four samples of "transitional" basalt glass, similar to Mariana Trough basalt, is in Table 2; major and trace element data for an andesitic and a basalt sample are in Parson, Hawkins, Allan, et al. (1992). Isotopic and bulk compositional data for a rhyolitic glass fragment are given by Hergt and Hawkesworth (this volume). The Pb-isotope data indicate an arc source and are the most radiogenic Pb compositions yet reported from the Lau Basin.

Site 839

Site 839 is different from all of the other sites drilled in the Lau Basin in that it has rocks similar in most respects to those of the modern Tofua Arc, yet they were erupted in the backarc basin. There seems to be little doubt but that Site 839 sampled an aborted attempt to form an arc edifice that was isolated shortly thereafter by the propagating ELSC and left as an inactive arc remnant. Site 839 is located about 55 km west of the ELSC on the eastern edge of the attenuated crust forming the WEB. The site is one of many elongated sub-basins separated by high-standing ridges presumed to be horsts. The basin is a half-graben about 15 km long and up to 5 km wide. It has up to 0.31 s TWT of sedimentary infilling that overlies an acoustic basement tilted to the southeast. Hole 839A drilled 206 m of sediment before encountering igneous rocks; at Hole 839B, 215 m of sediment was cored. Intra-lava sediment was recovered from as deep as 266.4 mbsf in Hole 839B. More detailed descriptions of this site are in Parson, Hawkins, Allan, et al. (1992) and Parson (this volume).

The 292 m of core recovered were separated into nine lithologic units that represent four distinct petrographic types (Ewart et al., this volume). The presence of abundant Cr-spinel, orthopyroxene (En_{59} to En_{74}), and calcic plagioclase (An_{88} to An_{90}) distinguish Site 839 rocks from those of the other backarc sites. The youngest unit (Unit 1) is an aphyric to moderately phyric clinopyroxene-olivine basalt distinguished by Cr-spinel megacrysts; Units 2, 5, 7, and 9 are moderate to highly phyric clinopyroxene-orthopyroxene-plagioclase basaltic andesites petrographically similar to Tofua Arc lavas. Units 3, 6, and 8 are sparsely to strongly phyric clinopyroxene-olivine basalts; some samples have abundant phenocrystal olivine, commonly with Cr-spinel inclusions, and loose grains of Cr-spinel. Unit 3 has Cr-spinel megacrysts. Unit 4 is a sparsely phyric olivine-clinopyroxene basalt and also has Cr-spinel megacrysts. Unit 9 has high ratios of plagioclase to pyroxene. In addition to having orthopyroxene, some of the Site 839 rocks are distinctive among Leg 135 samples in having abundant megacrysts, phenocrysts, and microphenocrysts of Cr-spinel. Allan (this volume) presents a detailed study of remarkable megacrysts of zoned Cr-spinels from Units 1, 3, and 4 that record the effects of magma mixing. The core compositions of these spinels indicate that they grew in primitive (Mg# = 75 to 80) low-Al melts of probable arc-tholeiitic composition. The parental melts would have been derived from highly depleted harzburgite or lherzolite mantle sources. Rim compositions of these spinel phenocrysts have crystallized from more evolved melts. Allan evaluates the possibility that the cores were xenocrysts from peridotite wall rock but concludes that magma mixing best explains relations. For example the composition of Cr-spinel inclusions in Mg-rich cores of olivine indicate a crystallization from high-Mg parental liquid, and the olivine phenocrysts also show strong zonation. He proposes that both minerals give evidence for hybrid melts formed by mixing of primitive melts and more evolved melts with Mg# = 54 to 56. The andesitic samples from this site are considered to be representative of the end-member postulated as a mixed component. Further support for these interpretations come from experimental work by Forsythe and Fisk (this volume), who show that the more Mg-rich spinel megacryst interiors would have been in equilibrium with low-Al and high-Mg# melts. Vitrophyric samples have xenocrysts of olivine with core compositions reaching Fo_{88} to Fo_{90}. These have narrow sharply zoned rims, a few microns wide, of Fo_{83} to Fo_{87}. The Fe-rich rims appear to be in equilibrium with the enclosing glass, but the cores clearly formed from more Mg-rich melts.

The most striking feature of rocks from Site 839 is their strong arc signature seen both in mineralogy and chemistry. Compositional data for a least fractionated sample from Unit 4 and a fractionated sample from Unit 5 are in Table 2. This arc signature also is apparent in trace element abundances normalized to N-MORB (Fig. 11), in particular, the relative enrichments in LILE and depletions in HFSE. The Nb and Ta depletions are especially distinctive as are the abundances of Ti, Na, and transition metals such as Cr and Ni.

Gaetani et al. (this volume) did experimental studies on basaltic andesite from Unit 1, an aphyric to moderately phyric clinopyroxene-olivine basalt. They determined the liquid line of descent under anhydrous conditions at 1 atm., and water-saturated conditions (approx. 6 wt% H_2O) at 2 kb in order to examine the effect of variable water content on arc magmas. Their results show that addition of water produces distinct shifts in the position of the clinopyroxene-plagioclase-olivine and clinopyroxene-plagioclase-orthopyroxene saturation boundaries. The crystallization sequence at 2-kb water-saturated conditions is Fo_{86}–Cr-spinel, then high Ca-clinopyroxene. Plagioclase is suppressed until about 20% crystallization; the first plagioclase to form is An_{94}. The residual liquid is an "arc andesite" with 56.3% SiO_2, 17.4% Al_2O_3, 8.2% FeO^*, and 4.82% MgO. They observe that the evolved residual liquid composition resembles the low-Mg andesites recovered at Site 839.

The high volatile content of the parental magmas is evident from the highly vesicular nature of the rocks and glasses. Vesicle abundances ranging from 10 to as high as 40 modal percent. Some of the vesicles are filled with mesostasis material.

The data supporting arc-like chemistry and mineralogy at Site 839 is paralleled by the isotope chemistry (Hergt and Hawkesworth, this volume). They show that isotope ratios for Site 839 samples resemble

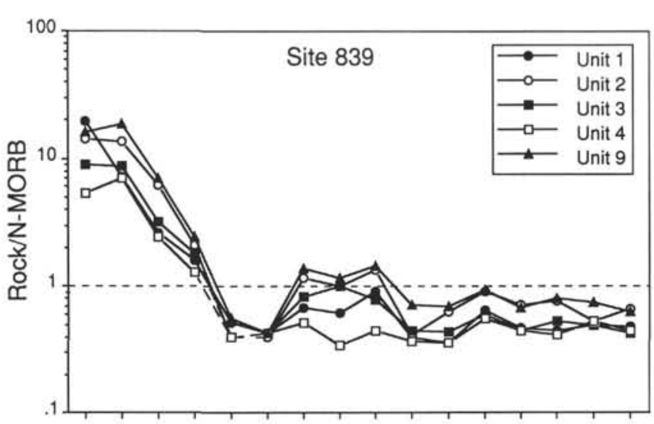

Figure 11. Trace elements normalized to N-MORB values for Site 839. Normalizing values from Sun and McDonough (1989). From Hawkins and Allan (this volume).

some of the modern volcanoes of the Tofua arc. It is likely that Site 839 has sampled part of an arc edifice that was abandoned after being cut off from the developing Tofua Arc by passage of the ELSC propagating rift (see also Ewart et al., this volume).

Discussion of the Isotope Data

Isotope data offer very important insights to the origin of magma systems; in particular, for studies of backarc basin, they are important in identifying potential end-members and studying the systematics of possible mixing relations. One of the most significant results of Leg 135 drilling in the Lau Basin has come from the isotope studies presented by Hergt and Hawkesworth (this volume) that bear on mixing of multiple mantle components in the evolution of Lau Basin crust. The isotope data, together with the mineralogic and geochemical data, have helped us understand the complex relations between tectonic style and magmatism during evolution of the backarc basin.

It has long been recognized that the Lau Basin has anomalously high $^{87}Sr/^{86}Sr$ ratios relative to MORB and to other western Pacific arc and backarc systems (Hawkins, 1976; Gill, 1976; Stern, 1982; Hawkins and Melchior, 1985; Volpe et al., 1988). Although Sr isotopes are higher than typical MORB values, the Nd isotopes resemble MORB (Carlson et al., 1978; Volpe et al., 1988). The isotope data, together with incompatible trace element abundances, have been used to suggest that several distinct mantle sources are required to explain the observed isotopic range of the active ridges and seamounts (e.g., Volpe et al., 1988; Loock et al., 1990; Boespflug et al., 1990). However, these observations, based on samples from the modern spreading centers, give a biased view as they have not taken into account the early history of the basin.

Evidence that the modern Lau Basin axial ridges and seamounts erupt basalts with isotope signatures requiring multiple mantle sources was first demonstrated by Volpe et al. (1988). They showed that Sr- and Nd-isotope data for basalt glasses from the CLSC and seamounts in the northern Lau Basin require (1) a MORB-like component depleted in incompatible elements; (2) a component similar to an E-MORB or OIB source that is enriched in incompatible elements; and (3), a distinct OIB-like component with high $^{87}Sr/^{86}Sr$ and low $^{143}Nd/^{144}Nd$ ratios. The Nd-Sr relations are more like MORB from the Indian Ocean than to the East Pacific Rise although all data are shifted to Sr-isotope ratios slightly higher than MORB. The Valu Fa Ridge (Jenner et al., 1987; Vallier et al., 1991) is a rift zone propagating into older oceanic crust at the southern end of the Lau Basin. Jenner et al. concluded that the trace element and isotopic data indicate a minor, but significant, contribution from the subducted slab. They proposed a mantle source for Valu Fa having both MORB and island-arc source isotopic signatures. Loock et al. (1990) presented additional data for the CLSC, for the Mangatolu Triple Junction in the northeastern Lau Basin, and for Valu Fa Ridge. Their data further support the interpretation that glasses from much of the northern Lau Basin and CLSC have the Indian Ocean signature ($^{87}Sr/^{86}Sr$ = 0.70309 to 0.70318 and $^{143}Nd/^{144}Nd$ = 0.51309 to 0.51312). The Mangatolu Triple Junction has ratios of 0.7040 and 0.51283, respectively, and the Valu Fa Ridge (propagating rift) has ratios of 0.70320 to 0.70339 and 0.51303 to 0.51306, respectively. Loock et al. (1990) and Boespflug et al. (1990) suggested that the arc-like isotope data for the Valu Fa Ridge could be explained by mixing between Pacific MORB and either a Tongan arc component or a flux of trace elements from the subducted slab. Hergt and Hawkesworth (this volume) note that the Valu Fa Ridge is the only part of the modern spreading ridge systems in the Lau Basin that lacks an Indian MORB mantle source signature.

Before drilling during Leg 135, the best interpretation of Lau Basin mantle magma sources was that the axial ridges were dominated by Indian Ocean MORB, propagating rifts in the southern and northeastern parts of the basin had a strong signature of an arc component mixed with Pacific MORB, and that the northern Lau Basin has an OIB component mixed with Indian MORB. The best evidence for each of these components is the Pb-isotope data but also is implicit in the Sr and Nd isotopes (see Hergt and Hawkesworth, this volume). There were no data for rocks older than about 1 Ma.

Hergt and Hawkesworth's data (this volume) gives further support, and better definition, to the idea that multiple mantle sources have been sampled during Lau Basin evolution. Even more significant is their recognition that magmas constituting the modern spreading centers are derived from a *recent* influx of asthenosphere having Indian Ocean mantle isotopic characteristics; this source has displaced an older source that had the isotopic signature of Pacific Plate asthenosphere. The range in isotope ratios for Leg 135 samples encompasses N-MORB values and those for the actively spreading CLSC and ELSC. Many of the new data are displaced toward lower Sr ratios from the linear trend between isotope ratios for the Tofua Arc and modern spreading centers, but they have high-Nd ratios similar to the active spreading ridges. The Pb data are most instructive in suggesting two different mixing trends. These require an "arc" component and a component from either a Pacific MORB or an Indian MORB source. Hergt and Hawkesworth (this volume) evaluate several possible explanations for the trace of the arc component. These are (1) magma mixing between upwelling mantle and arc magmas during the early stages of extension; (2) low but significant infiltration of slab derived flux into the backarc magma source; and (3) contamination by arc lithosphere during extension and upwelling. The "arc-like" component has low Ce/Pb and Nb/U ratios and relatively radiogenic Sr and Pb. The Pb composition of the "arc" end-member is more radiogenic than most of the data from the Tofua Arc but overlaps with Tafahi and Niuatoputapu, two of the northernmost Tofua Arc volcanoes, and with "old" Tonga Arc crust such as the Eocene rocks of 'Eua. No data are available to make a comparison for Pb isotopes of the Lau Ridge but, arguably, 'Eua may be equivalent. The N-MORB component is defined by Unit 7 at Site 834, which has relatively unradiogenic Pb and Sr like the Pacific N-MORB mantle sources. As discussed above, the trace element data for glasses from this unit have a smooth MORB-like normalized pattern.

Site 834 isotope data may be explained by mixing between arc and N-MORB components and offer insight to evolution of magma sources. The Nd isotope data are remarkably uniform (0.513109 to 0.513125) and are similar both to MORB and to the modern Lau spreading centers. The Sr data range from 0.702544 to 0.702958; they do not reach values typical of the modern Lau spreading centers (e.g., 0.7033). These are the most MORB-like Nd and Sr values reported to date for the Lau Basin. Good correlations exist between Sr- and Pb-isotope compositions. The Pb-isotope data define linear trends that indicate mixing and these mixing trends project into the field for Pacific MORB. As noted above, Unit 7, at mid-depth in the hole, is the most MORB-like unit and has the least radiogenic Pb. The Pb-isotope ratios become progressively more radiogenic upward to the youngest flow studied (Unit 2). Units 13 (deepest) to 8 also show a similar upward increase from less to more radiogenic; Unit 8 and Unit 2 are very similar. Units 7 and 8 offer the greatest contrast in Pb-isotope ratios. The variations with depth, and the apparent cyclicity of the variations, are interpreted as the result of mixing of different proportions of MORB-like and arc-like melts (Hergt and Nilsson, this volume). For example, they propose that the variation between Unit 7, the most MORB-like sample, and Units 2 and 8, the most radiogenic samples, may be explained by mixing between "essentially pristine MORB" and up to 50% of the "arc" end-member. The timing and nature of the mixing process are fairly well constrained and require mixing of mafic melts rather than melting of a heterogeneous source (Hergt and Nilsson, this volume).

Lavas from Site 837 and Site 839 have a strong arc signature in terms of mineralogy and chemistry (e.g., we found two-pyroxene basaltic andesite and andesite). They also have arc-isotopic characteristics. Site 837 samples are similar to the modern Tofua Arc, and Pb-isotope ratios lie on a trend between Indian mantle and arc. Site 839 Pb-isotope ratios are confined to a relatively small field, suggesting that they come from a well-mixed and homogeneous source. The Pb-isotope ratios lie on a trend between arc and the Pacific mantle and thus resemble Site 834 samples. The isotopic signatures for Site 839 resemble those for the modern Tofua Arc volcanic island 'Ata as well as those for the (Lau Ridge) arc-derived intrusive dikes at forearc Site 841. The arc rocks of Site 839 probably formed at an arc edifice that was cut off and isolated from the Tonga Ridge by the southward propagation of the ELSC.

Site 836 sampled crust formed early in the history of the ELSC. The propagating rift tip would have passed close to the site at about the time the Site 836 crust was formed (0.6 to 0.8 Ma). Probable igneous basement at Site 836 has a strong Indian mantle isotopic signature and resembles rocks of the present axial ridges. Samples from volcanic breccias or gravels are more radiogenic than the presumed basement and are similar to arc-like Site 837 samples.

Collectively, the data of Hergt and Hawkesworth, together with Loock's data, suggest that a remnant of "old Pacific" asthenosphere has been available under the easternmost part of the Lau-Tonga system for at least the past 6 m.y.. This inference helps explain a Pacific MORB-source input to the oldest drill site samples (Site 834) and to mix with Indian MORB-source and an arc component to form magmas of the modern spreading centers (e.g., Valu Fa Ridge). The remnant of Pacific mantle must have migrated eastward since about 5.5 Ma, perhaps in response to the eastward rollback of the Tonga Trench. Hergt and Hawkesworth (this volume) suggest that the influx of Indian MORB-source mantle under the Lau Basin replaced Pacific MORB-source mantle within the past 5.5 m.y. They postulate that the influx was coupled with the southward propagation of the ELSC. The rift propagation may have been the result or the cause of clockwise rotation of the Tonga Ridge and its underlying mantle. They note that the complex mixture of Indian, Pacific, and arc sources at the Valu Fa rift tip may be explained as a consequence of this process.

Mixing between MORB-source mantle and other components are not unique to the Lau Basin. For example, the Mariana Trough may have both a MORB and an arc component (Volpe et al., 1987, 1990). They interpreted the data as supporting a heterogeneous source rather than a mixing of melts, as proposed for the Lau Basin (Volpe et al., 1990). The northern Mariana Trough narrows to a propagating rift tip penetrating the Mariana Arc; mixing between MORB and either an arc component or a metasomatized (by fluid from the subduction zone) mantle is proposed by Stern et al. (1990). Nascent backarc basins behind the Shichito Ridge, in the Izu-Ogasawara Arc, are interpreted as showing evidence for mixing between E-MORB and island-arc *magmas* rather than derivation from heterogeneous magma sources (Ikeda and Yuasa, 1989). However, Hochstaedter, Gill, Kusakabe, et al. (1990) and Hochstaedter, Gill, and Morris (1990) invoke mixing relations between sources having E-MORB and a subduction component for the Sumisu Rift of the Izu-Ogasawara Arc system.

The question of whether the isotopic heterogeneity of backarc magmas is caused by selective melting of separate domains comprising a heterogeneous source, or by mixing with injections of different batches of melt, may be closely tied to the dynamics of backarc opening. Injection of new batches of MORB melt to mix with another type of mantle or melt (e.g., SSZ mantle) could be a consequence of mantle counterflow, upwelling, and melting in the extensional region above the subduction zone. Magmas should evolve toward more MORB-like systems as the "other" component is melted, mixed with, and overwhelmed by the influx of MORB-source mantle. Alternatively, a passive dynamic response may be visualized in which heterogeneous mantle is progressively melted with the more fusible (enriched) domains being consumed first. This too would lead to crust dominated by less enriched rocks. The second sequence of events would lead to a linear progression in composition rather than cyclicity and would not necessarily have N-MORB crust as the evolutionary end-point. The Lau Basin data suggest that the first model is a more likely explanation for Lau Basin evolution.

Backarc Basin Sedimentation

Comprehensive discussions of Lau Basin sediment studies and their tectonic significance are presented elsewhere in this volume by Bednarz and Schmincke, Bøe, Clift, Clift and Dixon, Hodkinson and Cronan, Ledbetter and Haggerty, and Rothwell et al. A brief synthesis of sedimentologic data relevant to the magmatic evolution of the Lau Basin and the adjacent island arcs is given here as it reflects on the basin's magmatic evolution. One of the more important discoveries is that intrabasinal arc-composition volcanism at numerous centers was a major source of the volcaniclastic sediment deposited in the sub-basins. These intrabasinal arc constructs were active synchronously with the nearby eruption of MORB-like lavas that formed backarc basin crust.

The backarc sites, as well as the two on the Tonga Ridge, display similar patterns of sedimentation. The oldest sediments recovered are late Miocene and the youngest are Holocene. The age of basal sediments differs at each site with a general pattern of younging eastward from Site 834. Except for Site 836, the sediments primarily comprise a lower sequence of volcaniclastic turbidites interbedded with hemipelagic clayey nannofossil mixed sediments, and nannofossil clays. These are overlain by an upper sequence of hemipelagic clayey nannofossil oozes, locally containing calcareous turbidites (Rothwell et al., this volume). They interpret the volcaniclastic turbidites as being derived from nearby arc fragments or constructs *within* the basin rather than from adjacent island arcs. They consider the Pliocene proximal volcaniclastic sediments to have been derived from intrabasinal seamounts. The Pleistocene record is mainly hemipelagic. It is important to note that they view intrabasinal "arc" volcanism as a major source of the volcaniclastic material. Site 836 is distinctive in having only 20 m of sediment overlying presumed igneous basement. The sediment has more than 50% volcaniclastic material in the lower half of the section. The glass shards constitute a limited compositional range from 54.3% to 58.1% SiO_2. These are best interpreted as being the result of fractional crystallization of magmas that formed the MORB-like basaltic basement at this site and the nearby ELSC. Similar basaltic andesite has been dredged from the high-standing ridges near Site 836 (Table 2; also Hawkins and Allan, this volume).

Rothwell et al. (this volume) propose that the Lau Basin had numerous intrabasinal seamount volcanoes in Pliocene time; these are the inferred sources for the volcaniclastic beds. Their interpretation that the volcanic centers were near the sub-basin depocenters is based on the preservation of delicate needle-like shards that could not have survived long distance transport, and the freshness of the glass. Neither the Lau Ridge nor the Tofua Arc are likely sources. The chemistry of the glass shards indicates that nearly all have low-K arc-tholeiitic compositions ranging from basaltic to rhyolitic. Some shards from Sites 834, 835, and 837 have medium-K arc chemistry. Glasses from Sites 834 and 835 span the full range from basaltic to rhyolitic, whereas Site 836 has a fairly unimodal composition (basaltic andesite, clustered around 55% SiO_2). Site 838 glasses also have a unimodal concentration of values ranging from 50% to 55% SiO_2. Bimodal suites are found at Sites 837 and 839 where basaltic andesites and rhyolites represent maxima. The Site 836 glasses are distinct in having trace element ratios similar to those of the nearby ELSC (e.g., MORB-like ratios for Ba/Zr of 0.9 to 1.4), in contrast to the much higher, arc-like values (3.9 to 10), found at the other sites. In general, all of the trace element data for glasses from the backarc sites have broadly similar patterns when normalized to N-MORB. These patterns show the SSZ magma signature, for example, moderate to strong enrichment in LILE, strong depletion in Nb, and minor depletion in other HFSE (Rothwell et al., this volume).

The provenance of the widespread volcaniclastic units is important to understanding the magmatic evolution of the Lau-Tonga system. The only epiclastic unit that can be tied to the Lau Ridge is a 3.3 Ma bed at Site 834 that closely matches the chemistry of the Korobasaga Volcanic Group. Rothwell et al. conclude that it was unlikely that there was widespread or voluminous volcanism on the Lau Ridge after extension and rifting was started in the Lau Basin. This idea is contradicted by the evidence that the Korabasaga Group was laid down between 4.5 and 2.5 Ma (Whelan et al., 1985) and was a potential source of sediment. Differences in eruptive style, sediment entrapment in the basin west of Site 834, a different pattern of dispersal, or a more limited locus of volcanism, all may have played a role in limiting volcaniclastic sedimentation. Cole et al. (1985) and Woodhall (1985) note that the Korobasaga Group is limited to six islands, forming a 35-km-wide belt, in the northern 200 km of the Lau Group. This may account for its relative unimportance as a source of volcanic detritus for infilling the sub-basins. Evidence discussed in a subsequent section suggests that the Tofua Arc was not a robust system; indeed, it may not have existed until about 1 Ma, yet there is an extensive record of Miocene volcanism on the Tonga Ridge. The most likely sources were (small and ephemeral?) intrabasinal eruptive sites (e.g., Clift, this volume). Zephyr Shoal (Table 2) is an example of such a feature. The low-K arc-series chemistry of many of the Tonga Ridge volcaniclastic units matches the compositions found in the backarc drill sites, and they too may have been derived from the intrabasin volcanoes.

The age of opening and sedimentation of the basins appears to young eastward. The time of transition from deposition of volcaniclastic sediments to sediments dominated by clayey nannofossil ooze varies from site to site but also tends to young eastward. The eastward age progression in timing of the locally derived voluminous eruptions of arc-like material suggests that a wave of arc-volcanic activity took place that "tracked" the eastward retreat of the Tonga forearc as it followed the rollback of the trench. This is a previously unrecognized aspect of backarc basin evolution that gives rise to "intrabasinal arc" edifices erupting alongside magma leaks that form MORB-like backarc basin crust. The arc-volcanic front, or zone, has been stabilized for about 1 Ma or less as the present Tofua Arc. Well-dated volcaniclastic units help constrain the longevity of the intrabasinal arc-like eruptive centers. For example, Site 835 records nearby activity from 3.5 Ma to 2.9 or 3.0 Ma. At Site 837, proximal sources laid down deposits from 2.1 to 0.34 Ma; at Site 838, proximal sources formed volcaniclastic beds from 1.92 to 0.50 Ma; at Site 839, the range is from 1.73 to 0.37 Ma (Bednarz and Schmincke, this volume). We infer that the "intrabasinal arc" constructs had eruptive life spans on the order of 1 to 2 m.y. before becoming inactive as the arc front migrated eastward.

The age of inception of the modern Tofua Arc has remained an important problem that bears on the nature of initial rifting of the ancestral Lau Ridge. Hawkins et al. (1984) proposed that the Tofua Arc was imprinted on the outboard edge of Lau Basin crust and probably was less than 1 m.y. old. The abundant evidence for Pliocene intrabasinal arc volcanism indicates that this idea is an oversimplification. Also, Tappin et al. (this volume) report an age of 3 Ma for the volcanic island Niuatoputapu at the north end of the Tofua Arc. There is no record of equivalent age Tofua Arc activity at the latitude of our drilling transect. However, abundant silicic volcanism is recorded in ash layers, all less than 840 k.y., that were cored on *Sonne* Cruise 30 (von Rad and Mohe, 1990). These ash beds are interpreted as fallout ashes, and ash remobilized as turbidites. Many ash layers less than 1 m.y. old were cored at Sites 837, 838, and 839 in the central Lau Basin, but only a few were found in the western basin at Sites 834 and 835. The ash layers from Leg 135 drilling all have an arc signature similar to the *Sonne* cores (see discussions by Bednarz and Schmincke and by Rothwell et al., both in this volume). These data suggest that a robust silicic volcanic arc became active in the eastern Lau Basin about 1 m.y. ago, but evidence for it before then is lacking. This probably was the time of origin of the present Tofua Arc on the drilling transect, but we cannot discount the possibility of older, less explosive, mafic volcanism. Tofua Arc volcanism may have developed first in the north and migrated south as the Lau Basin opened.

TONGA PLATFORM

No Miocene or early Pliocene volcanic edifices are known on the Tonga Platform; all of the record comes from dikes or volcaniclastic units (Exon et al., 1985); the latter were derived from a western source. Presumably, this was the Miocene Lau Ridge volcanic arc, although Cawood (1985) points out that the low-K arc series Miocene rocks of the Tonga Platform lack potential source areas in age-equivalent units on the Lau Ridge. He postulates along-strike, or cross-strike, compositional variation as a possible explanation. The low-K intrabasinal arc constructs (Bednarz and Schmincke, and Clift and Dixon, both in this volume) are potential sources for some of these units. The igneous rocks and volcaniclastics of the Tonga Platform are exposed on 'Eua (Cunningham and Anscombe, 1985; Hawkins and Falvey, 1985); as scarce interbeds in the uplifted platform carbonates, for example, in the Nomuka Group of islands (Cunningham and Anscombe, 1985); from drill cores from some of the Tonga Islands, as dredged rocks from their flanks (Exon et al., 1985); and in the ODP cores from the two Tonga Platform drill sites. Volcaniclastic sediments were sampled at Site 840, on the ridge crest about 45 km east-northeast of 'Ata Island, and at Site 841, 150 km east of 'Ata, on the upper slope of the Tonga Trench (Fig. 2). Before rifting of the Lau Basin, both sites were well to the west and part of the forearc to the Lau Ridge. Extension and rifting since 6 Ma has moved both sites away from the arc source on the Lau Ridge. Opening of intervening sub-basins has also formed sediment traps for detritus from western sources. The oldest volcanic rocks recovered from the Tonga Platform are the pre-late Eocene, low-K rhyolitic rocks from Site 841 (Bloomer et al., this volume).

Igneous rocks discovered in the Tonga Platform cores were among the most remarkable findings of Leg 135. They established the occurrence of intrusive activity in the Miocene *forearc* to the Lau Ridge as well as the occurrence of the Eocene age, low-K rhyolitic arc-volcanic complex and its subsidence of about 5200 m in 40 to 46 m.y.

Low-K Rhyolitic Complex

The deepest rocks recovered at Site 841 are from a 210.55- m section of low-K rhyolite, rhyolitic welded and nonwelded tuffs, breccias, and lapilli tuffs (see Parson, Hawkins, Allan, et al. [1992] and Bloomer et al. [this volume] for more details). The age of the rhyolite is well constrained by radiometric dating (McDougall, this volume), which yields an experimental K/Ar age range of 44 ± 2 Ma on glass from Subunit 2B. Basal sediments are calcareous volcanic sandstone of sedimentary Unit V, with planktonic foraminifers from Zone 15 (middle Eocene) to Zone P18 (early Oligocene) (Chaproniere, Nishi and Chaproniere, and Quinterno, all in this volume). The basal sediments are in fault contact with the top of the complex. A subaerial origin is interpreted for at least part of the rhyolite complex on the basis of welded textures, the lack of density sorting within the pumice breccias, and the lack of interbedded sediments or fossiliferous material. The present depth of the top of this formerly subaerial rhyolitic complex is 5200 m below sea level and 605 mbsf.

Five separate petrographic units were recognized; all are textural varieties of rhyolitic rocks comprising clasts, lapilli, pumice, glass shards and phenocrysts of andesine (An_{40-44}), and quartz with lesser amounts of augite, orthopyroxene, hornblende, and Fe-Ti oxides. Most of the samples are extensively altered to clay minerals, chlorite, pyrite, carbonate, quartz, and albite. Welded textures are a significant feature of the rhyolite, but nonwelded textures predominate. The welding could have occurred in a submarine environment, but this is discounted because of the lack of interbedded water-laid sediments or microfauna. The abundance of highly vesicular rocks and pumice is indicative of high volatile content, eruption at low confining pressure, or both. The lack of hydrous primary minerals as phenocrysts argues against melts with high water content, and the poorly sorted mixtures of very low-density pumice and more dense rock fragments precludes deposition in water. Subaerial or very shallow-water eruption seems to be the best explanation.

The least altered rhyolite is in a 19.51-m-thick unit of a quartz-plagioclase phyric rhyolite flow or dome (Subunit 2B). Parts of this unit have textures transitional to a poorly welded rhyolitic pumice. It was cored from about 612 to 632 mbsf. Subunit 2B (Table 3) has about 76.8% SiO_2, 4.35% Na_2O, 1.19% K_2O, and $Na_2O/K_2O = 3.66$. The soda-potash ratio sets it apart from "normal" calc-alkaline rhyolites, but it is typical of differentiated arc-tholeiitic series magmas. Tonga and Kermadec arc dacites share similar Na/K ratios and low-K relative to their silica content (Bryan, 1979a). A deeper unit (Unit 4, 747.2–757.4 mbsf) has clasts of basaltic andesite entrained in a 10.2- m-thick welded rhyolitic lapilli tuff. The deepest unit recovered (Unit 6, 776.2–815.6 mbsf) is welded rhyolitic lapilli tuff containing clasts of mafic greenstone and welded tuffs. A very significant observation is that some of the mafic clasts have Ti, V, Ni, and Cr abundances that indicate a MORB origin (Bloomer et al., this volume). The Site 841 rhyolitic complex must have been built on MORB-like crust rather than on arc or continental crust. Of the several possible explanations for the origin of this volcanic construct, two can be eliminated. These are genetic relation to the Late Cretaceous, rhyolitic Lord Howe Rise (van der Lingen, 1973), or derivation from a boninitic parental magma series. Bloomer et al. (this volume) conclude that it formed on rifted oceanic crust and is the product of extreme fractional crystallization of arc-tholeiitic magmas. It probably represents an early stage in evolution of the "Vitiaz Arc."

Post-Late Miocene Sills or Dikes

Forearcs of convergent margin systems have long been assumed to have very low heat flow and to be among the coldest parts of the Earth's crust. This view has been shaken by the discovery of a sill or flow of basalt in late Pliocene sediments at Site 781 in the Mariana forearc (Shipboard Scientific Party, 1990) and at Site 793 in the Bonin forearc where a sill of diabasic-textured basaltic andesite has intruded mid-Miocene sediments (Taylor, Fujioka, et al., 1992).

At Site 841, eleven separate sills or dikes of arc-tholeiitic basalt and basaltic andesite were cored (Table 3). These were intruded into distal facies of late Miocene volcaniclastic turbidites with fauna indicative of foraminifer Zone N16 to Subzone N17a, or 12 to 6 Ma (Parson, Hawkins, Allan, et al., 1992). Site 841 is presently 140 km east of the Tofua Arc axis and was at least 75 km east of the Lau Ridge volcanic-arc axis at the time of intrusion. A forearc setting is inferred.

The chemistry of the intrusives is arc-like; SiO_2 ranges from 50% to 54%, the average is 51.5%. The samples all show an enrichment in K, Ba, and Sr, a depletion in Nb and Ta, and a light depletion in REE relative to MORB (Fig. 12). Element ratios such as Y/Zr, Ba/Zr, Ba/La, La/Zr, and Sr/Nd are all similar to arc-derived rocks of the Miocene age (14–6 Ma) Lau Group (Bloomer et al., this volume). The isotope data for Site 841 basalts show $^{143}Nd/^{144}Nd$ that is slightly higher than for the Lau Ridge; however, Sr ratios are displaced to higher $^{87}Sr/^{86}Sr$, although not nearly as high as for the Tofua Arc. No Pb-isotope data are available for the Lau Ridge to make a comparison, but Pb isotopes from Site 841 lie in the Pacific MORB array for $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ and slightly above it on the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ plot (Hergt and Hawkesworth, this volume). In general, they are different from the data for 'Eua in that they are closer to the MORB field for Pb and closer to the Lau Ridge for Nd ratios.

The origin and mode of emplacement of the intrusions are problematic. Their chemistry establishes an affinity to the Lau Ridge volcanic arc rather than to forearc magmatism such as boninites. The long distance of inferred transport seems to pose a thermal problem as they are small bodies but not extensively fractionated (MgO is about 4.5% and Mg# is 41 to 43). Bloomer et al. (this volume) favor the idea that larger magma bodies have migrated outward from the arc and the intrusives represent smaller offshoots of these reservoirs.

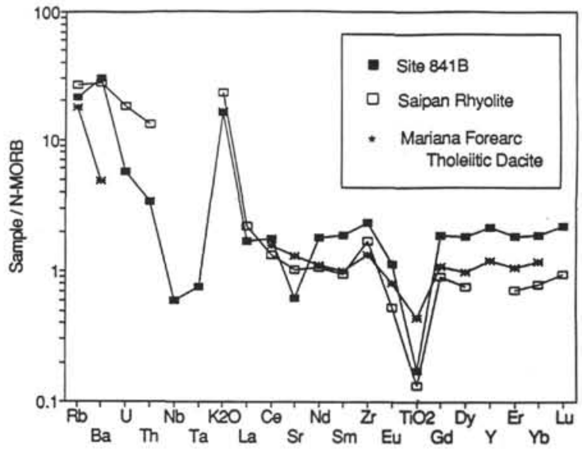

Figure 12. Trace elements normalized to N-MORB values for Site 841. Normalizing values from Sun and McDonough (1989). From Bloomer et al. (this volume).

SUMMARY AND A MODEL FOR LAU BASIN EVOLUTION

The evolution of the Lau Basin has been controlled by the kinematic response of the crust and upper mantle to oceanic plate subduction, mantle counterflow and upwelling, and extension of the suprasubduction zone (SSZ) lithosphere. The deformation has been accompanied by magmatic activity that involved both the SSZ mantle and "new" mantle brought in by convective overturn. A direct comparison to extensional provinces in continental settings may not be warranted, but there are general similarities to evolution of the Basin-Range Province of western North America where crustal extension and thinning have been accompanied by voluminous volcanism.

The opening of the Lau Basin has been accomplished by two modes of extension. Initially, rifting and extension formed basins separated by high-standing blocks; subsequently, seafloor spreading began and continues today. The seafloor spreading was promoted by propagating rifts. Extension was accompanied by extrusion of both arc-like and mid-ocean ridge type magmas that have formed new crust. The proximity to active island arcs, and the similarity in many respects between arc and backarc magmas, suggests that there may be similarities in petrogenesis. Here we will discuss a model that attempts to connect all of these observations.

Several requirements are necessary for any successful model for backarc basin evolution. These include an explanation for the spectrum of crustal rocks that range from MORB-like to arc-like; the cause for widespread extension at a convergent plate margin; the mechanism causing extension of the crust and its eventual rupture after which seafloor spreading becomes operative; and the apparent random location of the part of the remnant arc or forearc that becomes the site of extension and rifting.

The model shown in Figures 13 and 14 is an updated version of those presented by Hawkins et al. (1984) and Hawkins and Melchior (1985) and draws on more recent models of Stern et al. (1990), Parson et al. (1990), and Parson and Hawkins (this volume). Stern and Bloomer (1992) developed a similar interpretation to explain the early history of the Izu-Bonin-Mariana system and used it to explain the evolution of ophiolite remnants in California. Interpretations of data from ODP drilling in the Sea of Japan, the Izu-Bonin system and the Mariana Arc all have led to broadly similar models (e.g., Tamaki et al., 1992; Taylor, 1992; Fryer and Pearce, 1992; Pearce et al., 1992). All data such as ages, rock and sediment chemistry, geophysical data, and the regional geology used in this Lau Basin model are from the numerous contributions cited elsewhere in the text in addition to extensive data that I have collected.

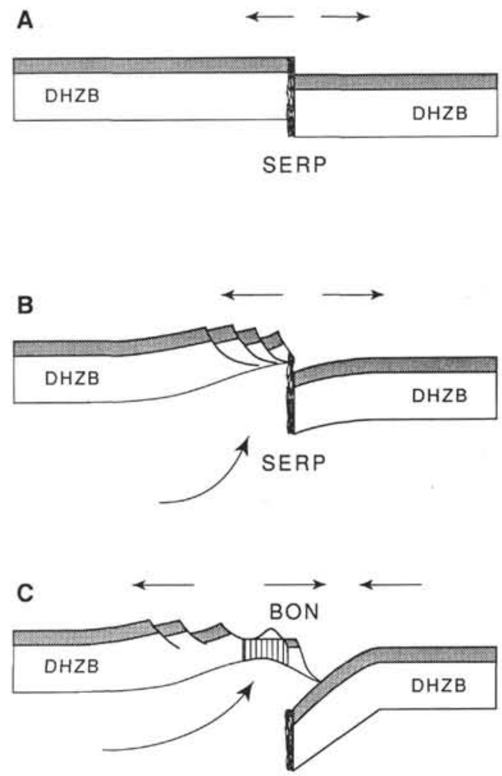

Figure 13. Schematic representation of a model for the initiation of boninitic magmatism (BON) in a precursor phase of development of a subduction zone. **A.** Shows oceanic crust (shaded), serpentinite (SERP), and depleted harzburgite (DHZB) residual from previous melting events. Oceanic crust, of different age and buoyancy is juxtaposed at a fracture zone. **B.** Uparching and extension are either causes, or results, of mantle upwelling. **C.** As discussed in the text, boninite forms by low-P, melting of strongly depleted hydrated peridotite. Subduction may not be required. (See also Stern and Bloomer [1992] and Hawkins et al. [1985]). The geometry of stage B is favorable for the beginning of subduction.

A discussion of the initial conditions at the beginning of Cenozoic time is beyond the scope of this paper, but we postulate that a dramatic change in rheology of the upper mantle, perhaps accompanied by a change in relative plate motion, brought into being a vast zone of extension and upwelling along the 4400 km trace of the West Melanesian–Solomon–Vitiaz trench and arc system (see Kroenke [1984] for an extensive discussion). A similar tectonic and petrogenetic scheme is required to explain the evolution of the Palau-Kyushu Ridge (e.g., Mattey et al., 1980) and was used by Stern and Bloomer (1992) for the early stages of the Izu-Bonin-Mariana system. The upwelling and extension promoted the development of boninitic magma systems early in arc history and at relatively shallow mantle levels; a similar mechanism is suggested for the boninitic magmatism in the Izu-Bonin Arc (Pearce et al., 1992; Stern and Bloomer, 1992) and for the Mariana forearc (Bloomer and Hawkins, 1987; Stern et al., 1991). Boninite magmas were followed by, and perhaps overlapped with, eruption of arc-tholeiitic series melts. There is a major problem in explaining the mismatch in the age of initiation of the western Pacific Eocene arc systems and the timing of the relative change in motion of the Pacific Plate (about 43 Ma) as inferred from the age of the bend of the Emperor-Hawaii Seamount Chain. Evidence for volcanism at least 5 to 6 m.y. earlier is found on the Palau-Kyushu Ridge and in the Bonin Islands (Taylor, 1992, and references therein). Thus, *initiation* of arc volcanism cannot be related to the change in motion even though plate convergence has been important subsequently. The boninitic and early arc-tholeiitic stages of an arc require melting of a suitable mantle source at relatively shallow depths (e.g., Stern et al., 1991; Pearce et al., 1992; Stern and Bloomer, 1992); plate convergence may not be required, and, in fact, lithosphere extension could well have been the control on early boninitic magmatism. The scheme outlined in Figure 13 gives an admittedly speculative view of how an early phase of extension could initiate the boninitic magmatism and set up crustal buoyancy differences that could be exploited by subsequent plate motion changes to cause subduction. A similar model was proposed by Stern and Bloomer (1992) and a less refined version was presented by Hawkins et al. (1984). Different, and equally plausible models, presented by Pearce et al. (1992) include subduction of a spreading ridge beneath young lithosphere. Whatever the mechanism for its initiation, the earliest stages of Eocene arc development in the northwestern Pacific resulted in an arc that was up to 300 km wide and 3000 km long formed of boninite and arc tholeiite (Taylor, 1992). A comparable arc developed concurrently in the southwestern Pacific.

Considering the present Papua-Vanuatu-Fiji-Lau-Tonga region, there is evidence for an Eocene Vitiaz Arc or West Melanesian Arc on the northeastern margin of the Indian Plate (Kroenke, 1984; Gill, 1987; Wharton et al., 1992) that may have been as much as 200 km wide and nearly 2000 km long. As summarized previously, and shown schematically as Figure 14A, the 40 to 46 Ma rocks include, arc-tholeiitic gabbro and gabbro-norite, pillow basalts, low-K rhyolite, and boninite. Although these ages are close to the inferred age for the Emperor-Hawaii bend, they are hard to reconcile with a cause-effect relationship. We would not expect an instant response of magmatism to the beginning of subduction. It is probable that there was a delay of a few million years between the onset of magma separation and upwelling from a mantle source and the beginning of eruption. The silicic rocks are voluminous (?), highly differentiated melts, that must have had mafic precursors. A further delay is likely to allow the extreme fractionation. Whether these rocks are related to extension or are the result of subduction, they constitute some of the oldest crust of the West Melanesian or Vitiaz Arc. Evidence for Oligocene arc magmatism is widespread and includes rocks on Fiji, 'Eua, and Vanuatu, as well as New Britain and New Ireland (Gill, 1987). These were no doubt a response to subduction after the change in Pacific Plate motion. The Eocene-Oligocene arc is shown on Figure 14B and is labeled as "E-O" in the following figures. Opening of the South Fiji Basin as a backarc system appears to have been a response to convergence at the Vitiaz Trench in Oligocene time; however, for the sake of simplicity, it has not been included in the figures.

The Miocene arc (Lau Ridge) shown in Figure 14C probably faced the present Tonga Trench, although we have no control on its actual orientation and it is likely that there may have been some rotation. It is uncertain whether or not a break occurred in arc volcanism in the early Miocene; neither outcrops nor drill-site data offer an indication for continuity. The Eocene-Oligocene arc probably formed part of the forearc to the Lau Ridge, as evidence from 'Eua and other Tongan islands indicates that Miocene dikes and volcaniclastic material were emplaced on the older forearc rocks. Nearly all of the rocks of the forearc have an affinity to arc magmas, which suggests that the underlying crust is largely arc material. Evidence for the presence of older MORB-like crust under the forearc can be found in the form of basaltic clasts in the low-K rhyolite at Site 841 (Bloomer et al., this volume). These are not unequivocal evidence for trapped older deep seafloor, inasmuch as MORB-like rocks are among the earliest magma types formed in arc rifting (e.g., Taylor, Fujioka, et al., 1992). MORB-like lavas in the Mariana forearc were reported by Johnson and Fryer (1990), who suggested that they may represent injection of magma during forearc rifting. We conclude that it is likely that much, or all, of the forearc is underlain by crust formed in an island-arc system.

The SSZ source for the Lau Ridge volcanic arc is considered to be depleted harzburgite that had been metasomatized by fluid derived from subducted, altered ocean crust (e.g., Pearce et al., 1984; Tatsumi et al., 1986; Tatsumi, 1989) and possibly from subducted sediments. The depleted mantle source carried the isotope signature of the source of Pacific MORB. Late in the evolution of the Lau Ridge, at about 6 Ma, it began to be rifted, and at about the same time crustal extension began to form rift basins in the forearc to the Lau Ridge. An imbrication or a flexure of the subducted plate may have been formed at the trench at about 7 to 8 Ma, as shown schematically in Figure 14D (see Billington [1980] for a discussion of this important feature). This change in the subducted plate may have been either a cause or an effect of this extension. The progressive deepening of the location of the flexure, presently at about 535 km depth, is shown in Figures 14D and 14E. The extensional development of the Lau Basin included an extension-and-rifting phase followed by seafloor spreading. As discussed elsewhere in this synthesis, the extension and rifting formed the basins drilled at all of the backarc sites except for Site 836, which we relate to the seafloor-spreading phase. Site 834 was fed mainly by a Pacific MORB source, and Sites 837–839 had an SSZ source, with some mixture of the Pacific source; Site 836 magmas, and the modern spreading centers, however, were derived from the Indian MORB source. The two-dimensional sketch does not show the complexity of the probable mixing pattern, as the Indian MORB source would have swept down the axis of the basin from the north in the wake of the propagating rift tip of the ELSC. The Pacific MORB source either was consumed, bypassed, or overwhelmed by the Indian source as the basin evolved. The final phase of evolution included the initiation and southward propagation of a second rift system at about 1–2 Ma that formed the CLSC. The development of a new volcanic arc (Tofua Arc) on the outboard edge of the Lau Basin probably occurred within the past 1 m.y. A complexity not shown because of limitations of the scale is the seaward march of the small, intrabasinal, arc-composition volcanoes that contributed to the infilling of the rift basins. These may have been no less significant at any time than the scattering of arc volcanoes that constitute the present Tofua Arc. Thus, the longstanding question of whether or not there was synchroneity of backarc magmatism and arc magmatism has an indefinite answer in the Lau Basin. Both phenomena have been operative more or less continuously *within* the basin as it opened. However, a well-defined arc on the Tonga Ridge did not appear until after about 6 Ma of backarc spreading.

The Lau-Tonga system serves as a model for understanding relict convergent margin rocks found as terranes accreted to continental margins and as ophiolite series assemblages. It seems likely that eventually plate reorganization will lead to the amalgamation of the arc and backarc system of the western Pacific and their eventual accretion to a continent. There is ample evidence in areas of composite terranes such as the Philippines for amalgamation of arc-backarc-forearc systems like the Lau-Tonga system (e.g., Hawkins and Evans, 1983; Hawkins et al., 1985; McCabe et al., 1985; Evans and Hawkins, 1989). There is also general acceptance of the idea that many ophiolites are remnants of SSZ magma systems. In many ways the Leg 135 transect, together with the other western Pacific ODP studies, provide important new data and insight to understanding continental evolution through terrane accretion.

ACKNOWLEDGMENTS

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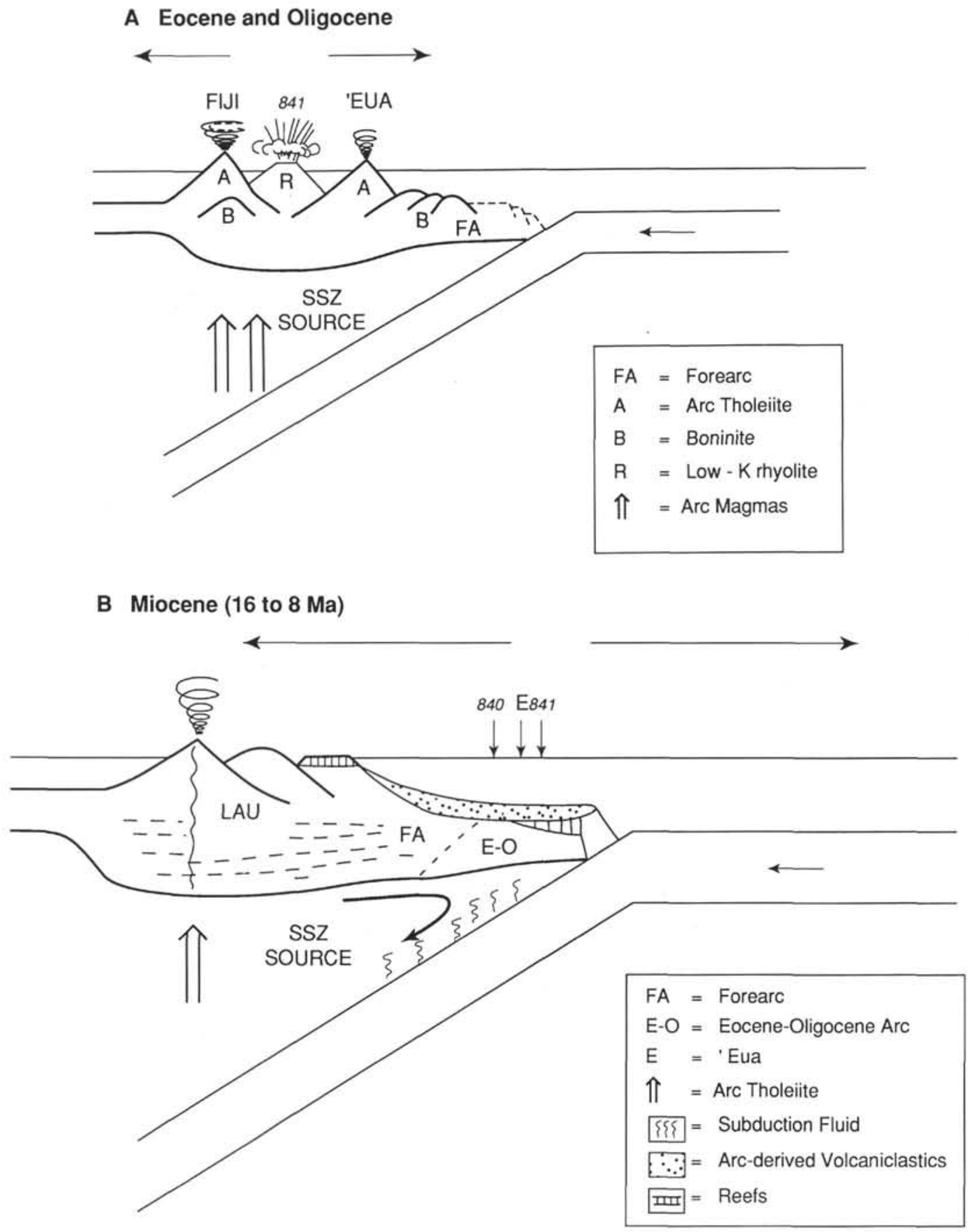

Figure 14. A model for the evolution of the Lau Basin. **A.** An early phase of the Vitiaz or West Melanesian Arc system forming arc rocks now found in the Fiji Islands and the Tonga Platform. The rhyolitic rocks of Site 841 are postulated to be part of this system. Arc polarity is not known but is assumed to have faced north or northeasterly. **B.** The Lau Ridge volcanic arc and possible locations of exposures on 'Eua and Sites 840 and 841. This view and subsequent views are projections on to latitude 20°S. **C.** The beginning of crustal extension to form the WEB and rifting of the Lau Arc. Magmatic activity has not yet filled the extensional basin but mantle upwelling is setting the stage for it. Note the insertion of the flexure or imbrication in the subducted lithosphere. **D.** Effects of extension, mantle upwelling, and magmatism in the backarc basin. The ELSC has been propagated southward and is fed by the Indian MORB source. Not shown are the ephemeral intrabasin arc-composition volcanoes that formed progressively from west to east. **E.** The present configuration with the Tofua Arc and the ELSC both active. Not shown is the CLSC that is propagating southward. The vertical dimension in all sketches has been exaggerated to show details of crustal geology.

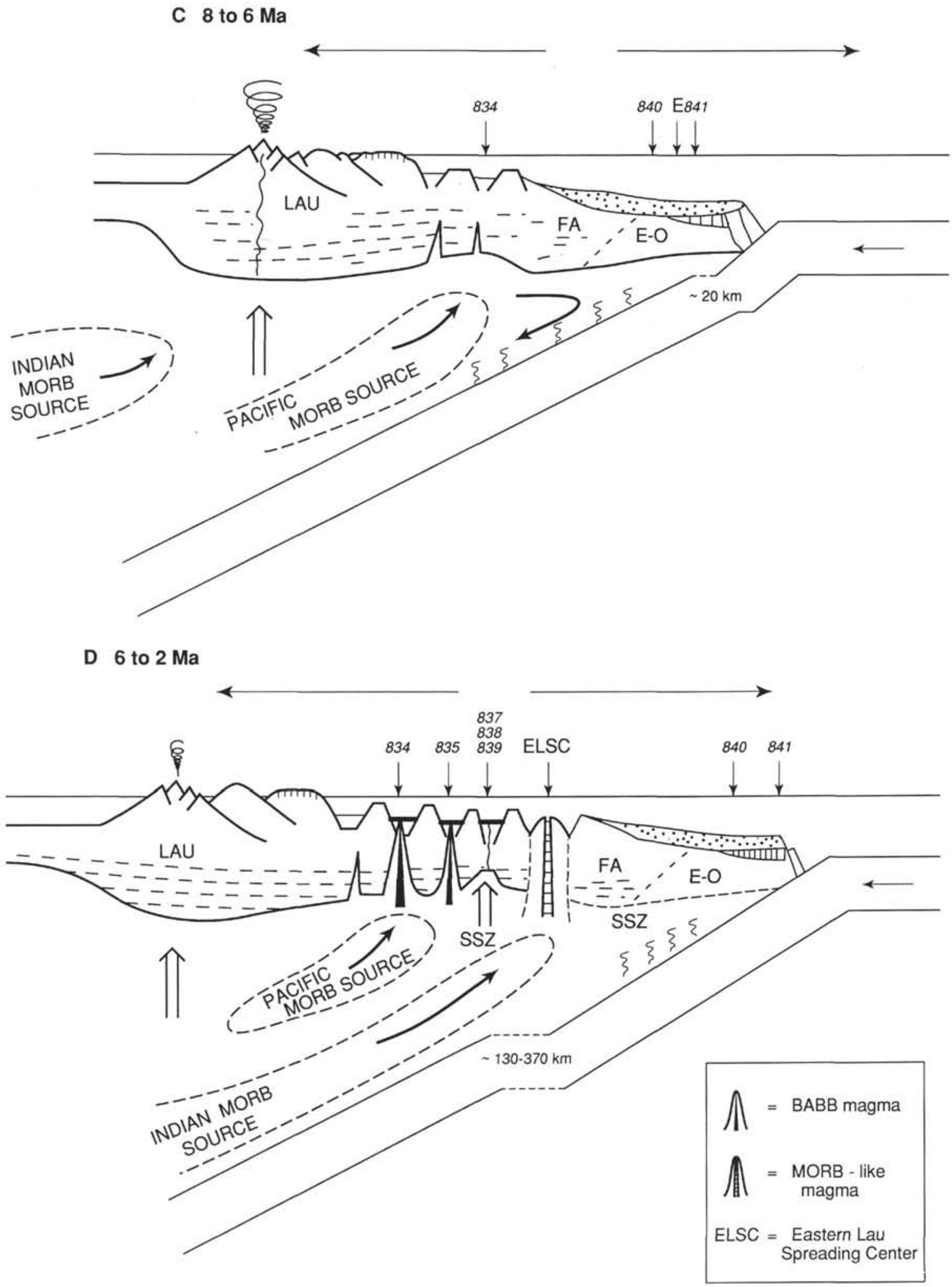

Figure 14 (continued).

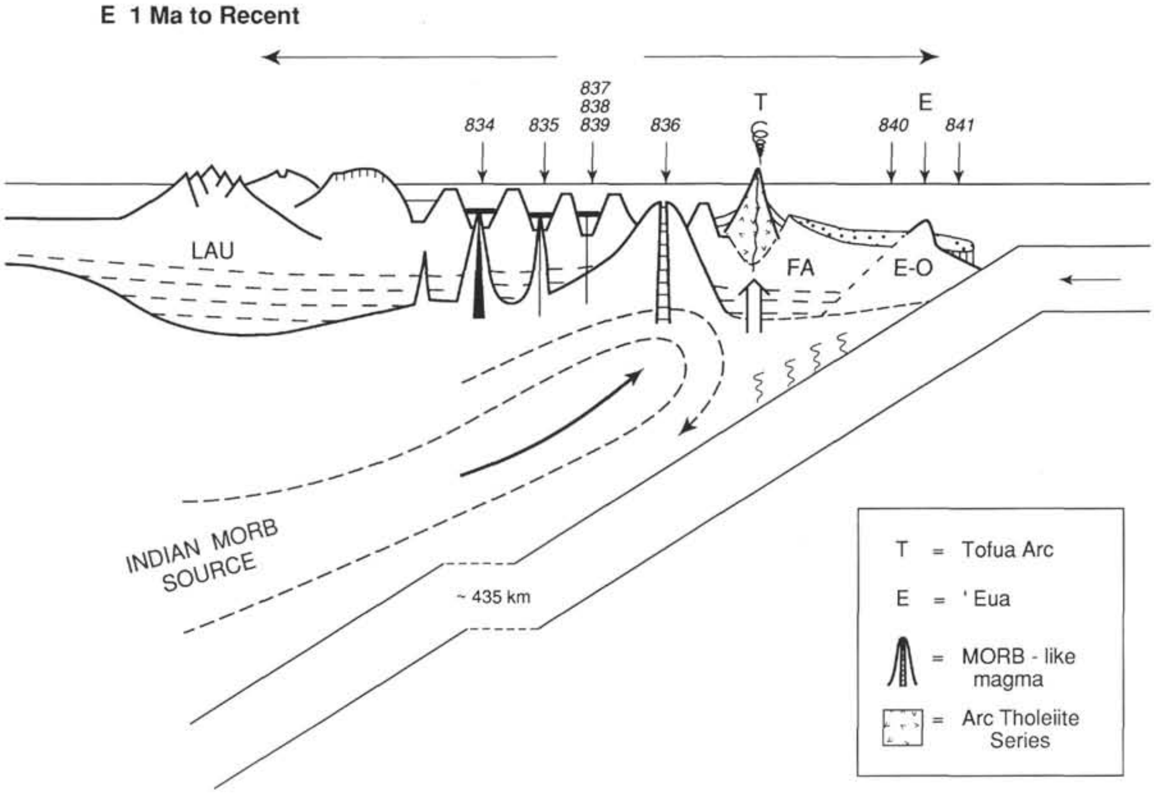

Figure 14 (continued).

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