Geological Society of America Special Papers

Geology of supra-subduction zones-Implications for the origin of ophiolites

James W. Hawkins

Geological Society of America Special Papers 2003;373;227-268
doi: 10.1130/0-8137-2373-6.227

Geological Society of America
Special paper 373
2003

Geology of supra-subduction zones—Implications for the origin of ophiolites

James W. Hawkins
Geosciences Research Division, Scripps Institution of Oceanography, La Jolla, Califorinia 92093-0220, USA

ABSTRACT

Ophiolites are rock assemblages recognized as relict fragments of oceanic crust and upper mantle. They provide insights into petrologic and seismic layering of in situ oceanic lithosphere at mid-ocean ridges and into deep oceanic crustal layers not easily sampled. Ocean crust also forms at other settings. At convergent oceanic plate margins, aged oceanic lithosphere is subducted into depleted mantle. Above the subduction zone, new oceanic crust is formed in the forearc, volcanic arc, and backarc basins. The magmas, products of supra-subduction zone (SSZ) processes, carry chemical and isotopic signatures reflecting SSZ sources and can be distinguished from mid-ocean ridge magmas. Ophiolite assemblages within belts of accreted terranes (e.g., the North American Cordillera) indicate an oceanic origin for some terranes, but their close geologic association with arc volcanic and volcaniclastic material, granitoids, and silicic extrusives makes a mid-ocean ridge origin suspect. There is much similarity between Cordilleran ophiolites, many other ophiolites that are associated both with tectonically juxtaposed volcanic arc complexes, and the Cenozoic SSZ systems of the Western Pacific Basin. The latter includes immature SSZ systems, such as the Tonga-Kermadec, Vanuatu, Mariana-Izu-Bonin, and also more evolved systems, such as Palau-Kyushu Ridge, Fiji, Luzon, and Mindanao. All geologic units that characterize ophiolites are found in SSZ rock assemblages, as well as distinctive rock types, such as boninite and island arc series rocks. Backarc basin crust is predominantly formed of rocks similar to mid-ocean ridge basalt, but commonly includes variants gradational to island arc tholeiites. Forearc and backarc basins may have thick ponds of arc-derived tuffs and siliciclastics. These may lie on nearly coeval oceanic crust. Parts of backarc basins, starved of clastics, may have pelagic sediments, metalliferous sediments, “umbers” and protoliths for cherty argillites. Mature arcs may be underlain by plagiogranites and tonalites that are plutonic complements of silicic arc volcanism.

Key Words: ophiolite, boninite, backarc basin, trenches, subduction.

INTRODUCTION

It is widely accepted that ophiolites are fragments of oceanic crust and depleted peridotitic mantle, the latter being the depleted residue from which magmas that formed the crustal units were derived (cf. Moores and Vine, 1971; Moores and Jackson, 1974; Coleman, 1977). Recognition of ophiolite assemblages in orogenic belts and collages of accreted terranes (Coney et al., 1980) establishes the incorporation of former oceanic crust into continental margins. Many geologists accept the notion that this ocean crust formed at deep-water “mid-ocean ridge” spreading centers (e.g., Bailey et al., 1970; Moores and Vine, 1971; Moores and Jackson, 1974; Coleman, 1977; Hopson et al., 1981; Moores et al., 2000). Under this notion, oceanic crust, destined for recycling into the earth’s mantle at subduction zones, is inferred to have been emplaced onto sialic crust by a process termed “obduction.”

*jhawkins @ucsd.edu

Hawkins, J.W., 2003, Geology of supra-subduction zones—Implications for the origin of ophiolites, *in* Dilek, Y., and Newcomb, S., eds., Ophiolite concept and the evolution of geological thought: Boulder, Colorado, Geological Society of America Special Paper 373, p. 227–268. For permission to copy, contact editing@geosociety.org.

Miyashiro (1973) proposed the highly controversial idea that the Troodos ophiolite formed in an island arc. Since then, the number of papers in which a subduction zone origin for ophiolites is presented has increased steadily. This discussion summarizes some of the evidence supporting supra-subduction zones (SSZ) as likely sources for many, if not most, ophiolites.

ON SUPRA-SUBDUCTION ZONES

Our understanding of SSZ systems has greatly benefitted from results of the Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) studies focused on the Tonga, Mariana, and Izu-Bonin arc systems. Many of the facts and interpretations summarized here come from observations and syntheses of drilling results presented by Kroenke et al. (1980); Hussong and Uyeda (1981); B. Taylor et al. (1992); Fryer et al. (1992); and Hawkins et al. (1994), and from regional syntheses of data by Hayes (1980, 1983); B. Taylor (1995); B. Taylor and Natland (1995); Fryer (1992); Stern and Arima (1998); and Stern (2002). There are many modern SSZ systems, as illustrated in Figure 1. All are potential sites for generating ophiolite-type rock series.

I will use the Mariana and Tonga arc-trench systems of the Western Pacific Basin as type examples of regions where new oceanic crust is formed by magmatism in the forearc and backarc-basin, as well as in the nascent stage of arc volcanism, in the region above the Wadati-Benioff zone. This is the SSZ tectonic setting, a term first proposed by Pearce et al., 1984. Figure 2 is a schematic cross-section of a generic intra-oceanic subduction system. It is based largely on the distribution of rock series, morphologic features, and geophysical data for the region between the Lau Ridge remnant arc and the Tonga Trench, but also applies generally to most other regions having island arc-trench systems. The entire region between the trench and backarc basin lies above the array of deep seismic events, which define the Wadati-Benioff zone. In the Lau-Tonga system, the seismic zone extends under the remnant arc. This arrangement, however, is not always observed in other SSZs. The seismic zone marks the presumed location of the subducted plate of oceanic lithosphere; hence, this is the supra-subduction zone environment. Some may choose to call this a "convergent plate margin," but in fact, the geologic processes distinctive for this setting extend across the entire region and are not limited to the site of convergence at the trench that marks the convergent margin. An important observation about SSZ environment is that although it is spatially associated with the *convergent* margin, the entire region between remnant arc and the outer trench swell is undergoing *extension*, and the hinge line outboard of

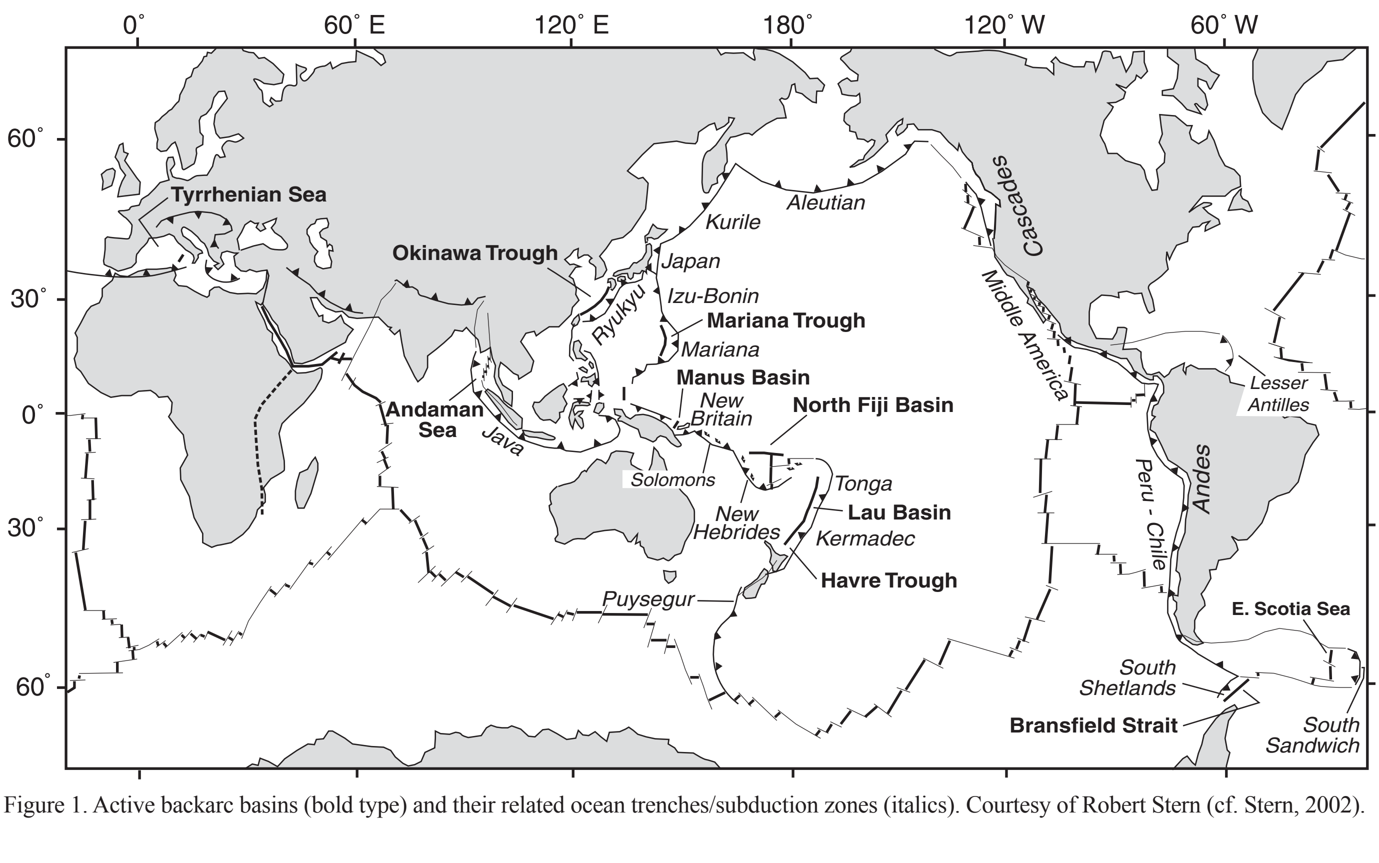

Figure 1. Active backarc basins (bold type) and their related ocean trenches/subduction zones (italics). Courtesy of Robert Stern (cf. Stern, 2002).

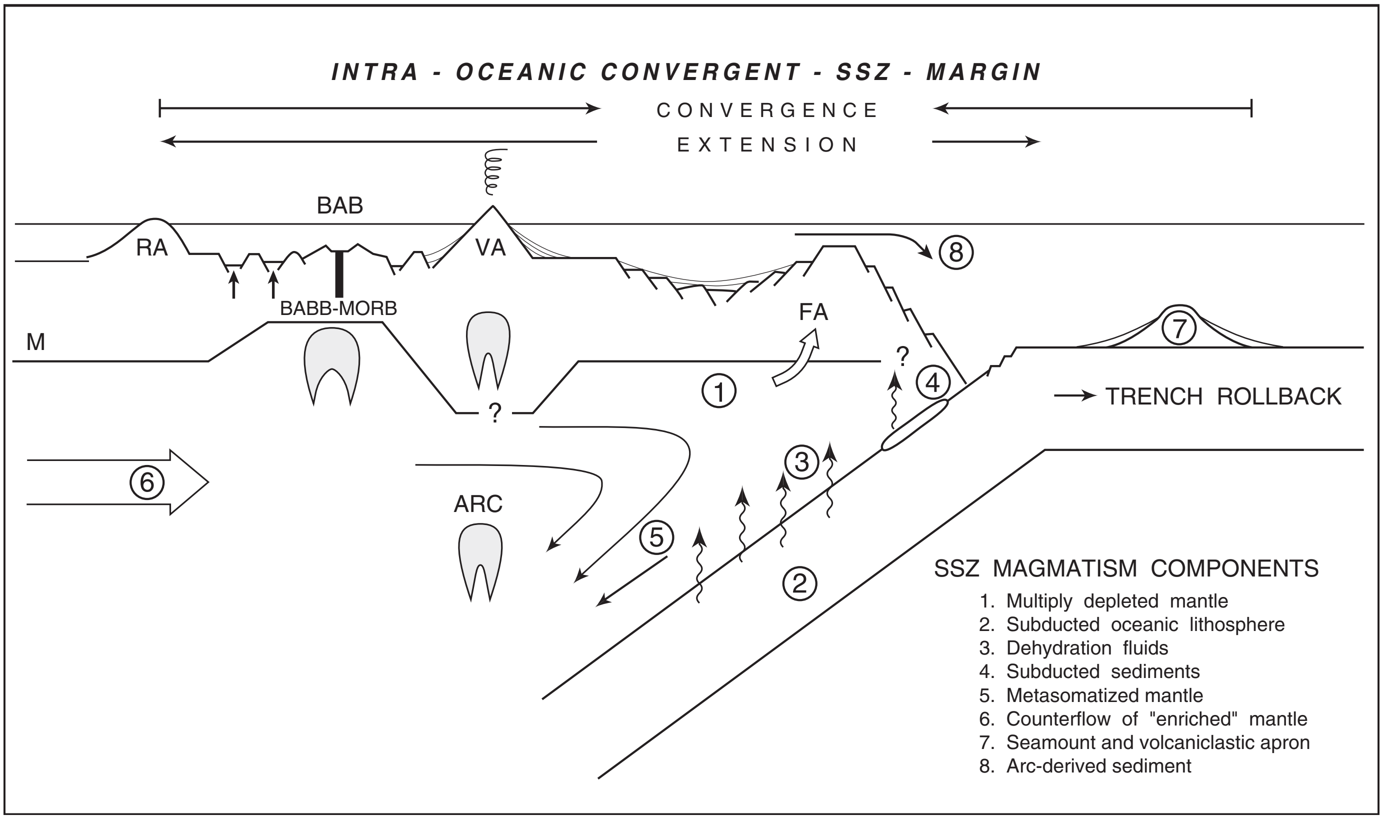

Figure 2. Schematic cross-section of an intra-oceanic SSZ system based on the Lau Basin-Tonga Trench region. Vertical dimension is at varied scale in order to conveniently show all parts. Crustal features are: RA—remnant arc, e.g., Lau Ridge; BAB—backarc basin, e.g., Lau Basin; VA—active arc, e.g., Tofua Arc; and FA—forearc. The arcs mainly erupt low-K island arc tholeiitic basalt (IAT) and their differentiates, whereas the BAB is dominated by tholeiite ranging from MOR-type basalt to basalts having chemistry transitional to arc tholeiite, i.e., backarc basin basalt (BABB). Initial stage of development, in which boninites form in the forearc, is not represented. Forearc magmatism shown here represents limited later stage dikes of arc magma intruded into forearc sediments. Potential source materials for melts are identified on the sketch. Mantle counterflow provides a continual supply of fertile mantle. Not shown are backarc basin seamounts, which may include OIB, MORB, and IAT.

the trench retreats by rollback (cf. Oxburgh and Turcotte, 1970; Elsasser, 1971; Uyeda and Kanamori, 1979; Hamilton, 1995, 2002). Furthermore, geophysical data indicate that there is a broad region of strong seismic wave attenuation (low-Q), probably from high temperature (*T*) and some partial melting, underlying the backarc basin (Barazangi and Isacks, 1971; Crawford et al., 2003). Sleep and Toksoz (1971) presented a model for opening of backarc basins showing convective overturn in SSZ mantle. They stated that "The down going slab generates a flow pattern by 'dragging' the low velocity layer material in the aesthenosphere. The convective cell thus driven exerts a stress at the bottom of the lithosphere behind the island arc forcing it in the direction of the trench." They also observed that "Heat transport by conduction alone cannot explain the high heat flow in the marginal basin." Influx and upwelling of relatively enriched mantle, driven by convective drag of the subducting plate (cf. Ida, 1983; Ewart and Hawkesworth, 1987), provides a continuous source for basaltic melts. These features, as well as the various components potentially available to form SSZ magmas, are discussed in more detail below.

The main purpose of this essay is to summarize the nature of the rocks formed in the SSZ setting and the processes important in their petrogenesis. Inasmuch as there is an extensive literature on details of petrogenesis, I will emphasize here the petrology of certain rock types and rock associations that are distinctive for the SSZ setting and that distinguishes these rocks from those formed at mid-ocean ridge spreading centers. The goal is to show how we can distinguish between ophiolites and associated rocks of the SSZ environment from ocean crust formed at mid-ocean ridges. Most backarc basin rocks could successfully masquerade as "true" mid-ocean ridge basalt (MORB), especially if they have experienced low-grade metamorphism. Rocks formed in arcs and forearcs have some fundamental differences from most in backarc basins, yet there is some gradation and overlap. Collectively, the arc and forearc assemblages are distinctive and are unlike crust formed at mid-ocean ridge spreading centers. A key point I wish to make is that in order to distinguish between ocean crust (ophiolite) formed at a mid-ocean ridge and ocean crust formed in an SSZ setting, one must evaluate all aspects of the geology of the ophiolite assemblage.

A Brief History—The Nature of the SSZ Setting

Introduction

The concept of the SSZ as a location of generation of new crust and recycling of aged oceanic lithosphere back to the mantle forms an important part of our present understanding of plate tectonic processes and crustal evolution. There was considerable knowledge of the geology of what we now call the SSZ setting, long before the advent of the modern plate tectonics model in 1961, but the data had not been integrated to form an understanding for, let alone to frame a model for, the geologic processes occurring in these regions. As with any formulation of a model for earth processes, there were many studies that laid the groundwork. What follows here is a brief summary of the key observations and interpretations pertinent to the SSZ setting, dating back to 1911, but mostly from between 1940 and 1980. My examples are mainly from the Western Pacific Basin, although related research was conducted in all major oceanic trenches. These years can be considered the "era of pioneer exploration" in arc-trench-backarc systems. The examples cited here were important antecedents to the extensive work of the last 20 years, which has produced a voluminous database and body of literature. Elsewhere in this volume, Julian Pearce (Chapter 15) presents a complementary discussion of the historical development of thought on the SSZ origin of ophiolites.

The first presentation of data and a description for the SSZ setting was by Marshall (1911), in which he summarizes: "the Tonga-Kermadec-New Zealand line is an important structural feature in the southwestern Pacific. The soundings throughout the ridge are shallow, the summits that project above the ocean along its crest are nearly all volcanic on the west side of the axis and raised limestone on the east. The whole line is bounded by a profound deep on its eastern side…in all of these islands the volcanic rocks of Cainozoic age are mainly andesite as in Tonga, Kermadec, and northern New Zealand. Everywhere to the east of this curved line the whole unbroken depths of the Pacific extend to the American coast…" Marshall also notes that "the island arcs….are largely formed of andesitic rocks, though they are usually associated with basaltic rocks" (1911). He describes the known great deeps of Kermadec-Tonga, New Hebrides, and Mariana, and suggests that they are "foldings" or a "synclinorium."

In 1961, Dietz (1961) and Hess (1962) presented their landmark papers proposing that new seafloor crust was being generated at ocean ridges, that seafloor spread from these ridges, and that older crust was "disappearing" (not being subducted!) into the mantle at the ocean trenches. According to Dietz (1961) oceanic "median rises mark the up-welling sites of divergence and the trenches are associated with the convergences or down-welling sites." Prior to 1961, the location and spatial association of volcanic island arcs, zones of deep seismicity, and deep ocean trenches had been recognized. The common development of arcuate arrays of island arc volcanoes is one of the more striking features of the SSZ setting. As mentioned above, Marshall (1911) recognized an "andesite line." This line separated "more acid lavas, especially andesites" from "the truly basaltic central area of the deep Pacific seafloor" (Keunen, 1950). The trace of the andesite line in the Western Pacific followed the eastern limit of the great trench systems. Hess (1948) proposed that drawing the andesite line on the outer, convex, side of the trenches would make it a useful structural, as well as petrographic, boundary.

Alfred Wegener (1915, see also 1966) gets credit for the first formal presentation of a model for continental drift. In his book, he also presented a model pertinent to opening of back arc basins and showed a cross-section sketch not unlike many that have been drawn for SSZ systems. Wegener proposed that "the island arcs, and particularly the eastern Asiatic ones, are marginal chains which were detached from continental masses, when the latter drifted westwards and remained fast in the old seafloor, which was solidified to great depths. Between the arcs and the continental margins later, still-liquid areas of seafloor were exposed as windows" (Wegener, 1966). These "windows" of seafloor are the backarc basins.

Our understanding of trenches, especially those of the Western Pacific, profited by the thousands of kilometers of bathymetric data collected in naval operations in World War II. Harry Hess (1948) presented a synthesis of seafloor bathymetry and fabric for the Northwestern Pacific Basin in which the curvilinear array of trenches, known to be "deeps," were identified as "axes of tectogenes" (down-buckling of the crust), whereas their adjacent island arc systems were labeled as "geanticlines."

Trenches and Benioff Zones

The exploration of the Western Pacific arc-trench systems in the 1950s led to a better understanding of the bathymetry of trenches and of the geophysical data for crust and upper mantle (see summary in Fisher and Hess [1963]). Hess (1948) presented profiles across the Mariana-Bonin, Kamchatka-Kurile, and Tonga-Kermadec trench-arc systems. He noted the previously unrecognized continuity of the deep trenches and their parallelism to island arcs; the close spatial association of seismic activity, gravity anomalies, and volcanism along the arc-trench trends; and the common occurrence of peridotite and blueschist metamorphism. Benioff (1949) showed the relation between ocean trenches and deep seismic zones that were delineated by an array of seismic events that he interpreted as representing the traces of great faults. "The oceanic deeps associated with these faults are surface expressions of the downwarping of their oceanic blocks" (Benioff, 1949). He noted that in some regions, such as Tonga, the array dipped uniformly at ~45° to nearly 700 km depth, whereas in Kamchatka–Kuriles, the array defined a zone convex upward that varied from nearly horizontal near the surface to ~60°. In a discussion of Dietz (1961), Bernal (1961) presented an interpretative sketch showing an ocean-continent interface with a thrust plane separating the plates, and proposed that the Pacific trenches formed by overthrusting along an inclined planar zone represented by deep earthquakes. He also suggested that "vulcanicity associated with mountain building is assumed to arise from activity of the fault plane itself" (Bernal, 1961).

Early models for plate tectonics (e.g., Morgan, 1968; Isacks et al., 1968) recognized three types of plate boundaries, one of which is "trench or young fold mountain." Morgan noted that "the compressive boundary seems to be the most difficult to delineate." For the Tonga–New Zealand–Macquarie system, he proposed that "fast compression leads to the trench-type structure" whereas "the [Macquarie] ridge is the result of slow compression." Neither backarc nor forearc basins were recognized as significant components of early plate tectonic models.

Extensive exploration of the crustal structure and rock types of trenches began in the 1950s. For example, Raitt et al. (1955) made seismic profiles parallel to the axis of the Tonga Trench and estimated the depth to the Moho (P-wave velocity >8.1 km sec^{-1}) along the trench axis at 20 km below sea level, whereas outboard of the trench, the same velocity was measured at 12 km below sea level. The highest velocities recorded beneath the Tofua Arc were 7.6 km sec^{-1}. They then extrapolated to a depth of 20 km under the arc before reaching mantle velocity. Samples recovered from the Tonga Trench wall at depths of 9 km included both unaltered and partly serpentinized peridotite (Fisher and Engel, 1969). Bowin et al. (1966) previously had collected serpentinite and altered basalt from the Puerto Rico Trench.

Backarc Basins

As noted above, Wegener's prescient suggestion for the origin of backarc basins was published before much was known about the trenches themselves. Holmes (1931, 1965) presented a trench arc system cross section that showed but did not discuss convective overturn in the mantle under arc systems, and a rift basin, equivalent to a backarc basin. For an interesting discussion of early evolution of ideas about mantle convection see Allwart (1988).

The first specific suggestion for extension and formation of new seafloor inboard of a trench-arc system appears to have been Rodolfo's (1969) proposal that the Andaman Sea had formed as a rhombochasm resulting from south to southeastward rifting of the southeast Asia margin. Karig (1970) observed that the Tonga-Kermadec, New Hebrides, and Mariana island arc systems displayed similar geologic characteristics. He concluded that inter-arc basins (e.g., Lau Basin and Havre Trough) are "underlain by crust of oceanic character and…[are] the locus of high heat flow." Karig probably was first to postulate "a process of island arc tectonism…during which the frontal arc and trench position migrate, generally away from the continents, concomitantly creating new oceanic crust, without mid-ocean ridges, in the ridge bounded basins behind the arcs." Packham and Falvey (1971) presented a similar observation; their main new contribution was data for heat flow, magnetics, and sediment thickness. They concluded that "genesis of oceanic crust takes place in a way that is analogous to seafloor spreading, but not at a mid-ocean ridge"; however, they suggested that "upwelling behind the Benioff zone, by a mechanism similar to…mid-ocean ridge crustal generation" was involved.

Moberly expanded on the backarc basin model and showed how various possible tectonic configurations for relative plate motions could develop different crustal fabric in the backarc basin. He, too, recognized the importance of trench and arc migration, proposing that "trench and arc migrate seaward against the retreating line of flexure of the suboceanic lithosphere…warm aesthenosphere…migrates up…forms new lithosphere in the extensional region behind the advancing arc" (Moberly, 1972). He also discussed causes and consequences of arc migration, and made some reconstructions of past trench configurations for the Western Pacific.

Barazangi and Isacks (1971) determined that a zone of strong seismic attenuation, possibly the result of high-temperature mantle, lay beneath the Lau Basin. The relation between subduction and mantle counterflow above the subduction zone was addressed by Sleep and Toksoz (1971), who showed how "marginal basin" extension and elevated heat flow may be related to convection in low-velocity aesthenosphere.

All backarc basin models postulated that some type of ocean crust underlies backarc basins. The first verification that the backarc basin crust was indeed basaltic and had MORB-like chemistry was in 1970, when we sampled the axis of what is now called the East Lau Spreading Center, while on the Scripps Institution of Oceanography 7-TOW expedition (Hawkins et al., 1970). Soon afterward, Hart et al. (1972) established that Mariana Trough basalts also were like MORBs.

Island Arc Magmas

The notion that the arc systems include andesitic rocks goes back at least to Marshall (1911) as cited above. In their synthesis of data for trench systems, Fisher and Hess (1963) suggested that "andesitic magmas might be produced where deformation is enough to force the basalt, or lower density crust layer, downward sufficiently to cause partial fusion." The most significant advances to understanding petrogenesis in arc systems came in the 1970s when Ringwood and colleagues made fundamental contributions based on laboratory melting experiments. Ringwood (1974) succinctly observes that "The most important petrologic problem relating to the development of island arc systems is the origin of the basalt-andesite-rhyolite volcanic suite."

In my opinion, the work of Coats (1962) was one of the most significant early advances in relating arc magma chemistry to the addition of water into the region of arc magma generation via altered seafloor rocks. Coats' work predated publication of the plate tectonic hypothesis. Owing to time lags in publication, his work followed the concept of geosynclines and was not integrated with plate tectonics concepts; however, in hindsight, we can recognize that his model anticipated many modern ideas about relations between subduction and arc volcanism. He observed that Aleutian Arc volcanoes ranged in composition from basalt to rhyolite, but were largely pyroxene andesite. He proposed that "andesitic rocks (formed) through addition of water and hyperfusible materials from eugeosynclinal deposits to eruptible basaltic material in the mantle" (Coats, 1962). The *eruptible basaltic material* was in a deep mantle layer containing "vitreous interstitial basalt." The water-bearing phases were dragged down along inclined faults systems, which he interpreted as reverse faults dipping under the arc. At that time, first motion studies of P waves were "inherently ambiguous…two possible orthogonal planes either of which might be the actual fault plane…two solutions for each earthquake" (Coats, 1962). Coats based his model on the interpretation that allowed N-S compression. His ideas were still couched in the now defunct geosynclinal dogma, and he proposed that the entrained ("subducted") material included eugeosynclinal assemblages of serpentine blocks and breccia, basalt, eclogite, greenstone and eugeosynclinal sediments.

Dickinson and Hatherton (1967) compared magma chemistry with the depth to the seismic zone beneath arcs. Data for different arc systems showed a positive correlation between K_2O and SiO_2. Using comparable values for SiO_2 (e.g., 55 and 60%), they plotted K_2O against depth to the seismic zone for which they found a positive correlation. They proposed that volcanic arcs were "located above the intersection of the Benioff zone with sub-horizontal zones favorable for melting within the Gutenberg low velocity zone" (Dickinson and Hatherton, 1967). Baker (1968), in an early synthesis of arc petrology, proposed that the relative maturity of island arcs was reflected in their chemistry; relatively young arcs such as Tonga, Mariana, and South Sandwich had a spectrum of magma types, from basaltic to rhyolitic but mainly basalt and basaltic andesite, whereas more mature arcs such as the Indonesian, Aleutians, and Kamchatka displayed the same range in composition, but were mainly andesitic.

Ringwood and co-workers made fundamental contributions to subduction system magmagenesis; they addressed the distribution of arc volcanoes relative to Benioff zones and the apparent temporal evolution from arc tholeiite to calc-alkaline melts, a rock series he called "orogenic volcanic series." The mantle wedge and basalt, sediment, or both, of the subducted plate were considered as sources. Green and Ringwood (1968) proposed that the (parental) basalt series of island arcs results from melting of pyrolite in the mantle wedge above the subduction zone. Considering the role of subducted crust, Ringwood (1969) also proposed that "partial melting, at high *P* [pressure] and shear stress, of subducted ocean crust leads to generation of calc-alkaline magmas." The role of water was recognized in Ringwood (1974), who noted that it is "most plausible that water is introduced along the Benioff zone via the subsiding lithosphere plate" and that it is "unlikely that the source of this water is to be found in subducted sediments" but rather it is "introduced as hydrated minerals occurring in the subducted oceanic crust" (Ringwood, 1974).

Gill (1970) used his data for Viti Levu, Fiji to show how during Cenozoic time there had been three successive periods of volcanism and plutonism, all of which could be related to subduction. Another landmark paper on arc volcanism was Miyashiro's (1974) presentation, which discussed variations in magma chemistry of island and continental arcs. He showed that they could be separated into calc-alkaline, tholeiitic, and alkaline series. These compositional differences could be related to crustal thickness, and crustal thickness could be correlated with the relative abundance of calc-alkaline series to tholeiitic series in different ocean arcs.

Perfit et al. (1980) presented a compendium of island arc magma chemistry and discussed implications for mantle sources. Jim Gill (1981) summed up the state-of-the-art thought on arc magmatism in his book, *Orogenic Andesites and Plate Tectonics.* The literature on SSZ petrology has expanded greatly in the last 20 years. Key papers are cited in the section on island arc petrology.

OCEAN CRUST

The fundamental magma type that has erupted to form earth's crust since mid-Proterozoic time has been Mg-rich melts ranging in composition from komatiite to picrite and basalt. We find the evidence for this in ancient rocks preserved within and on the margins of continents. Direct sampling of present day ocean crust shows that the ocean basins have been floored with basalt since at least 200 Ma, and most would agree that this process has prevailed throughout most of earth's history.

The primary site of origin of ocean crust has been at mid-ocean ridge spreading centers. It is estimated that ~80% or more of the basaltic crust of present ocean basins was generated at mid-ocean ridge systems. Perhaps another 10% of this crust was generated at SSZ in the backarc or forearc of intra-oceanic convergent plate margin systems. The remainder formed as oceanic plateaus, linear volcanic chains, and isolated seamounts. Basalts that form mid-ocean ridge crust are referred to as mid-ocean ridge basalts (MORBs); a subset of this broad group is termed normal mid-ocean ridge basalt (N-MORB) (cf. Viereck et al., 1989). N-MORB represents relatively unfractionated tholeiitic basalts having trace element and isotopic chemistry indicative of small amounts of melting of a Ca- and Al-enriched peridotite source that, relative to the composition of estimates of the bulk silicate earth, is depleted in Th, U, alkali metals, and light rare earth elements. (cf. Ringwood, 1975; Hart and Zindler, 1986). N-MORBs, as defined, have distinctive trace element and isotopic characteristics and generally low concentrations of magmaphilic elements. They are distinctive in their marked depletion in alkali metals, alkaline earths, light rare earth elements, and water, relative to other basalt types. Mid-ocean ridge crust is not formed exclusively of N-MORB; commonly, there are regional compositional variations that may be attributed to variation in source chemistry, extent of melting, and fractionation, but on most ridge segments the least fractionated rocks are similar to the model N-MORB chemistry. The petrogenesis of MORB is not the focus of this discussion, but inasmuch as their chemistry gives insight to their mantle source and melting processes, MORB serves as a useful point of comparison for the magmas of the SSZ setting and for understanding their mantle sources and melting processes. The chemistry of SSZ magmas indicates that their sources may include mantle even more depleted than the mid-ocean ridge source. There are rare SSZ occurrences of rocks resembling intra-plate, or off-axis, magmas that are indicative of mantle sources enriched, compared to the MORB source.

Present annual mid-ocean ridge basalt magma production is estimated to be ~24 $km^3\ yr^{-1}$. It is estimated that, since solid earth accreted, nearly three-fourths of the upper mantle has been partly melted, and that crust formed from those Mg-rich melts has been recycled back to the mantle (O'Neill and Palme, 1998). Since Paleozoic time, the maximum residence time of ocean crust has been ~200 Ma. Lithosphere formed at mid-ocean ridges ages in transit to the subduction zones; it thickens, subsides, gets altered by hydrothermal processes, is injected by off-axis magmas, and receives a thin veneer of pelagic, terrigenous, and meteoritic sediments. In the grand geologic scheme of crustal evolution, ocean lithosphere is destined to return to the mantle via subduction in ocean trench systems. Yet, we find remnants of old ocean crust and its subjacent melt-depleted upper mantle preserved on margins of continents, as uplifted masses of ocean lithosphere in some arc systems, and on micro-continents in ocean basins (e.g., New Caledonia, Fiji, Luzon, Mindanao). These are the ophiolites. Are these truly remnants of crust formed at mid-ocean ridges? I will present evidence to support the hypothesis that these allochthonous fragments of fossil oceanic lithosphere are predominantly from the SSZ environment.

THE NATURE OF OPHIOLITES

Ophiolites, first defined as a rock assemblage by Steinmann (1927), have been the topic of many extensive discussions (e.g., Coleman, 1977; Panayiotou,1980; Peters et al., 1991; Dilek et al., 2000), all of which support the contention that they are fragments of oceanic lithosphere. The strongest evidence for their oceanic origin is their siliceous and carbonate sedimentary rocks bearing oceanic fossil assemblages, and the biochemical precipitates interbedded with and overlying pillow basalts. Umbers and metalliferous sediments (Robertson and Hudson, 1973) similar to those that form at ocean ridge hydrothermal systems (Bostrom and Peterson, 1966) also are common. Pillow basalts of ophiolites, morphologically identical to those dredged from and photographed on the seafloor, overlie a sheeted-dike series. This same petrologic layering is found on the seafloor, as evidenced by ocean drilling (e.g., ODP Site 504). Rocks from deeper levels of ophiolite assemblages resemble dredged samples of gabbros, cumulate rocks, and fresh or partly serpentinized peridotites. The peridotites show varying degrees of depletion in the components of the basalts (e.g., they are depleted lherzolite, harzburgite, and dunite) that comprise the probable lower oceanic crustal and upper mantle rocks.

Interpretations of oceanic seismic velocity layering, based on seismic velocities determined for ophiolite samples (e.g., Salisbury and Christensen, 1978; Brocher et al., 1985), have supported a lithologic and velocity stratigraphy as shown in Figure 3. This is a somewhat circular argument that ophiolite equals ocean lithosphere, which in turn is ophiolite, but most geologists will agree that it is qualitatively correct; however, "ophiolites are defined largely in terms of petrologic and structural field relations, whereas the majority of the oceanic crust, owing to its inaccessibility, is best defined in geophysical terms, specifically, its velocity structure as revealed by seismic refraction" (Salisbury and Christensen, 1978). A major problem is that many ophiolites have rock types, stratigraphic and temporal relations, and details of rock chemistry that are contrary to what is known about ocean lithosphere formed at mid-ocean ridge

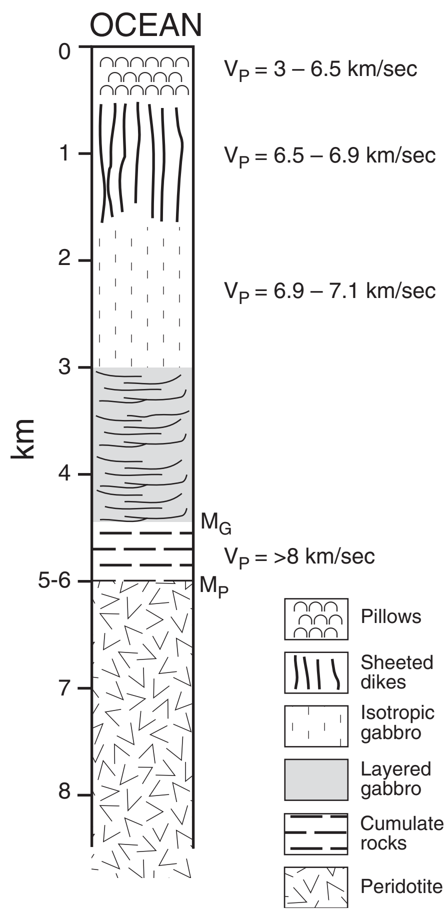

Figure 3. Ideal rock column for ocean crust and upper mantle. P–wave velocities (V_p) from Crawford et al. (2003).

spreading centers (e.g., Saunders et al., 1980; Hawkins, 1980; Pearce et al., 1981; Harper, 1984). Thus, we have the "ophiolite conundrum" (Moores et al., 2000). The problem that gives rise to this conundrum is the a priori assumption by many geoscientists that ophiolites are formed at mid-ocean ridges, when in fact we well know that ocean crust may be formed at other sites, as described above. With a proper understanding of SSZ geology, the supposed ophiolite conundrum can be eliminated.

GEOLOGY OF SUPRA-SUBDUCTION ZONE SETTINGS

Introduction

Convergent plate margins in intra-oceanic settings have a broad region that is several hundred kilometers wide across strike and that extends from the trench axis to the backarc basin. The major features of intra-oceanic convergent plate margins, the tectonic environment for SSZ systems, are shown in Figure 2. In addition to the trench, the major features are the backarc basin, active volcanic arc, and forearc. The forearc may include fragments of older convergent margin systems, old deep-sea floor ocean crust, or both. This region is situated above the trace of the inclined array of seismicity that defines the Wadati-Benioff Zone (WBZ), or subduction zone. The modern SSZ systems of the Western Pacific Basin form a nearly continuous array of trenches, arcs, and backarc basins extending across more than 100 degrees of latitude from Kamchatka to the nascent Puysegur and Hjort trenches south of New Zealand. In some settings (e.g., Tonga), the WBZ may extend beneath the remnant arc. In others (e.g., Mariana), the WBZ steepens to near vertical under the backarc basin or (e.g., New Hebrides) the WBZ may extend only to beneath the volcanic arc (Isacks and Barazangi, 1977). The term supra-subduction zone aptly describes the region in which there is concurrent extension, magmatism, and sediment accumulation in the backarc, arc, and forearc, as well as serpentinite diapirism in the forearc, and extension on the inner wall of the trench.

It is important to note that in Figure 2, the entire region from the remnant arc to the outboard side of the trench is under extension, even though on a regional scale it is the locus of plate convergence. The subducted plate itself is extended and broken up into a series of en-echelon rift basins as the incoming plate bends into the subduction zone (Hawkins et al., 2000). Note also the lack of an accretionary sedimentary prism in this generic cross-section. Accretionary prisms do, indeed, exist at some convergent margins (e.g., Sumatra, Antilles, eastern Aleutians), but the largest are in those regions where large drainage systems (e.g., Brahmaputra and Orinoco Rivers) bring in massive amounts of terrigenous sediments that funnel into the trench. Active volcanic arcs are close to newly formed crust of the backarc basin; siliciclastic arc-derived sediments are deposited directly onto crust having essentially the same age (e.g., <3 Ma) as the sediments. This would be most unusual for "normal" mid-ocean ridge crust. Extensional basins of the forearc serve both as depocenters for arc-derived clastics and as potential sites for the emplacement of mantle-derived melts. An important factor is that basaltic oceanic crust, arc rocks, granitoid roots of the arcs, and arc-derived clastics all form in close proximity and are essentially coeval.

The crust and mantle of SSZ settings constitute dynamic systems that evolve through the complex interplay of subduction of cold "rigid" lithosphere, upwelling hot mantle, transfer of heat and mass by magmatism, and crustal extension and rifting of SSZ lithosphere, coupled with, or driven by, rollback of the hinge line on the incoming subducted plate (e.g., Elsasser, 1971; Uyeda and Kanamori, 1979; Hawkins et al., 1984; Hawkins, 1994, 1995a, 1995b; Hamilton, 1988, 1995, 2002). Heat and mantle mass transfer may be accomplished by "ascending diapir (that) induces a descending return flow and results in a circulation" (Ida, 1983). Hamilton (2002) proposed that the incoming lithosphere plate "falls" into the mantle as the hinge line rolls back, that "the slab sinks more steeply than it dips…the magmatic arc follows the retreating hinge, extension affects the arc and backarc, and new hot mantle rises into the wedge from the backarc." Unloading of the crust by thinning and rifting has a feedback relationship to the rise and decompression melting of fertile mantle diapirs (McKenzie and Bickle, 1988; Hawkins, 1995a, 1995b). The tectonic setting is different, but the dynamic processes suggest similarities to the rise of mantle, ductile deformation of crust, tectonic denudation, and the close association of rifting and volcanism seen in detachment structures of the Basin and Range Province of North America (cf. Rehrig, 1986; Gans et al., 1989).

The first use of the term supra-subduction zone for this type of setting was by Pearce et al. (1984). Alabaster et al. (1982) used "supra-Benioff zone" for the same type of setting. In both papers, the term was used in reference to the site of origin of ophiolites and specifically for ophiolites derived from immature arc systems. Since 1984, there have been numerous uses of the term supra-subduction zone, both to describe present arc-trench system geometry and for interpretations of sites of origin for ophiolites. Hawkins and Evans (1983) and Hawkins et al. (1984) expanded the implications of the term to include an array of crust/mantle material that included forearc and backarc crust, as well as crust formed in a nascent arc. Part of the support for this extended use of the term was owed to recognition of the role and site of origin of boninites in the early stages of SSZ evolution and to the range in chemistry of backarc basin crust that ranged from MORB-like to island arc tholeiitic compositions.

In the northern hemisphere, the Western Pacific Basin (Fig. 4) has three concentric SSZ systems dating from mid- to late-Eocene time to the present, indicating that for this region, the present is the key to the past, at least through Cenozoic time (cf. Fryer, 1992). There has been sequential development of volcanic arcs and backarc basins on the eastern edge of the Philippine Sea since the inception of the ancestral "Mariana Trench" that formed the Palau-Kyushu Ridge. From oldest to youngest,

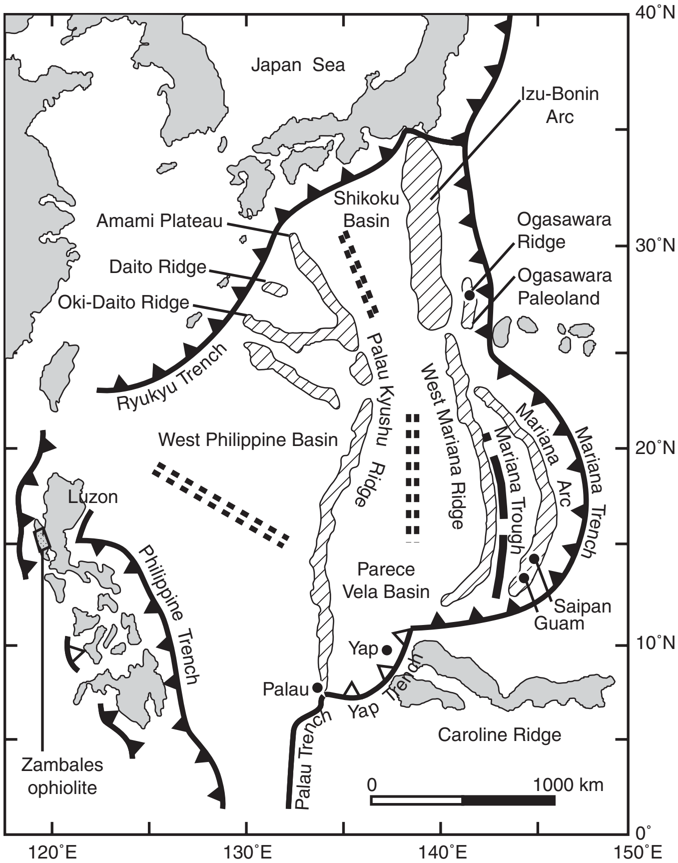

Figure 4. Schematic map of part of the northwestern Pacific, showing features of subduction systems discussed in text. Solid barbs denote active subduction systems. Open barbs are dormant or fossil systems. The actively spreading axis of the Mariana Trough is shown as solid lines. Inactive "fossil" spreading centers in Parece Vela-Shikoku Basin and West Philippine Basin are dashed.

from west to east, these features are: (1) Palau-Kyushu Ridge, active from mid-Eocene to late Oligocene (~48–30 Ma), and now a remnant arc; (2) Parece Vela and Shikoku backarc basins, which opened from 32 to 17 Ma and 25 to 15 Ma, respectively, and which both are now inactive; (3) West Mariana Ridge, active from late Oligocene to early Miocene, and now a remnant arc; (4) Mariana Trough backarc basin, active since ~6–7 Ma (?) to present; Fryer (1995) proposes 3.67 Ma for the beginning of seafloor spreading, but this was preceded by a rifting phase; (5) Mariana Arc, active since ~3–6 (?) Ma to present. Note the overlap in times of volcanic activity between "remnant arcs," backarc basins, and "frontal volcanic arcs," and that backarc spreading precedes appearance of the "frontal volcanic arcs." Outboard of these arcs is the Mariana forearc, which includes composite fragments of older arcs (e.g., Eocene and Oligocene volcanic units of Guam and Saipan), equivalent to volcanic rocks on the Palau-Kyushu and West Mariana Ridges. The volcanic rocks are partially capped by Miocene carbonate bank deposits and Pleistocene reef, bank, and lagoonal deposits (Cloud et al., 1958). The southwestern Pacific basin has at least three systems with similar configuration, petrology, and age ranges, but the geometry is more complex. A record of these is preserved on Fiji (cf. Gill, 1970). At present, there is an east-dipping subduction zone at the Vanuatu-New Hebrides Trenches and a west-dipping subduction zone at the Tonga-Kermadec Trenches. We find mid- and late-Eocene arc material and ophiolites on Fiji (Gill, 1970), and on the outer part of the Tonga Ridge, e.g., 'Eua (Ewart and Bryan, 1972), and at ODP Site 841 (Bloomer et al., 1994). This great assemblage of SSZ systems spans >45 Ma of history and lies waiting to be accreted to a continent, where it will constitute an assemblage of terranes like those of the western North America Cordillera (cf. Coney et al., 1980). Some of these Cordilleran terranes constitute ophiolites (e.g., Harper, 1984). It is reasonable to infer that pre-Cenozoic SSZ systems developed similar concentric arc-backarc systems. The Cenozoic Western Pacific Basin may be a mirror-image analogue for the Eastern Pacific in Paleozoic-early Mesozoic time (cf. Hawkins et al., 1984).

SSZ Magma Systems

The geologic and geometric features listed above help characterize active SSZ systems; however, there are additional factors that are critical for identifying remnants of SSZ systems preserved in the geologic record. Magmas erupted in modern SSZ settings have distinctive chemical, isotopic, and mineralogic characteristics that help distinguish them from major magma types such as N-MORB or ocean island basalt (OIB). These differences reflect the more complex history of the SSZ source relative to the MORB sources, and offer constraints on the *nature* of the SSZ magma sources. These data give useful insights into the extent of melting, chemistry, and mineralogy of the mantle source, and into eruption processes; however, the mineralogy and chemistry, if used alone, are probably insufficient for inverting the problem and understanding the tectonic setting in which magmas developed. For example, rocks from the Chile Rise carry an SSZ trace element signature, but they formed at an oceanic spreading center (Sturm et al., 2000). When petrology and geochemistry are coupled with other data, such as field associations, rock series, age, stratigraphy, and regional structures—in short, the geology—we should be able to make well-reasoned inferences about the tectonic setting in which an ophiolite series was formed.

Sources of SSZ Magmas

There are three main source components important for SSZ petrogenesis (Pearce et al., 1984; Hawkins et al., 1984; Hawkins and Melchior, 1985; Ellam and Hawkesworth, 1988; Stern et al., 1991). The primary source is the SSZ mantle wedge situated above the Wadati-Benioff Zone (Fig. 2); in nearly all SSZ settings, this should be depleted mantle (cf. Woodhead et al., 1993).

The subduction component is the oceanic lithosphere, of which the most important contributions are from altered ocean crust, altered upper mantle, and a thin sediment carapace, all of which are enriched in volatiles (mainly H_2O, CO_2, and CH_4) and elements mobile in aqueous fluids (cf. Tatsumi and Eggins, 1995). The fluid-mobile elements are the large ionic radius lithophile elements (LILE). These are also known as low-ionic potential elements (i.e., they have a small value for charge/ionic radius). A third potential component is mantle material convectively entrained above the subduction zone by viscous drag exerted by the subducted plate, and brought in under the backarc basin and arc (cf. Ewart and Hawkesworth, 1987). This "new" mantle may be relatively fertile and capable of generating N-MORB, enriched MORB (E-MORB), or OIB (Hawkins, 1976, 1995a, 1995b; Hawkins et al., 1990; Volpe et al., 1987, 1988, 1990; Ikeda and Yuasa, 1989).

The melting process preferentially extracts magmaphilic elements from the mantle source, as controlled by their bulk distribution coefficients (D_O) for mantle source materials. Both high field strength elements (HFSE) and LILE have low values for D_O and thus will be concentrated in melts, with the further constraints being element abundance in the source and the extent of melting. There are abundant data from active SSZ systems that show that HFSE are depleted in all magma types relative to N-MORB, whereas the LILEs show variable enrichments. This variable enrichment in LILE is attributed to the subduction component (fluids, melts, or both, derived from subducted sediments and lithosphere), whereas the HFSE depletion stems from previous melt extraction.

Mantle wedge. The petrology of the SSZ mantle wedge (SSZ-MW) is arguably the fundamental control on the chemistry of magmas erupted and on the mineral composition of the rocks formed on cooling. Initially, the SSZ-MW probably is dominated by extensively depleted peridotitic mantle from which basalt previously was extracted (e.g., at a mid-ocean ridge or in a backarc basin). Relative to the MORB-source mantle, which itself is depleted relative to bulk earth mantle, it is even more depleted in all magmaphilic or low partition coefficient elements, including both LILE and HFSE. We find examples of this depleted SSZ-MW in rocks dredged from trench walls, such as the abundant dunite and harzburgite, both serpentinized to various degrees (e.g., Fisher and Engel, 1969; Dietrich et al., 1978; Bloomer and Hawkins, 1983). Peridotites of ophiolite terranes exhibit similar depleted mantle material (e.g., Hawkins and Evans, 1983; Harper, 1984; Wallin and Metcalf, 1998). Because of this previous depletion, the depleted peridotites of the initial SSZ-MW are unlikely sources for extensive melt generation because they have low abundances of Al and Ca (which reside mainly in clinopyroxene) and alkali metals. This greatly limits the amount of melt they can form. The most likely melts would be the class of high-Mg SSZ rocks known as boninites, discussed below, but boninites require high thermal gradients and addition of water to the source to allow further melting of already depleted mantle (cf. Crawford, 1989).

Partial melting is promoted by convective transport of hot mantle, adiabatic decompression from upwelling, and lowering of solidus temperature *(T)* of hydrated (e.g., serpentinized) harzburgite mantle.

Subduction component. Subduction of altered oceanic crust, serpentinized upper mantle peridotite, and sediment all serve to convey water, as well as mobile, low ionic potential elements into the SSZ-MW (Peacock, 1990; Tatsumi and Eggins, 1995). The low ionic potential trace elements include Li, K, Rb, Cs, B, Ba, Sr, La, Ce, U, and Pb. Th is not considered to be mobile in fluids, and Th enrichments observed in some SSZ magmas are likely from mobilization in melt. The chemistry of the LILE-enriched fluids is controlled by dehydration of structural water from hydrous minerals and by fluid/mineral partitioning of mobile elements. Extent of re-enrichment may be time-dependent or may be controlled by abundance of fluid-bearing and LILE-bearing minerals in the subducted crust. In addition to metasomatizing the depleted mantle with fluid-mobile elements, the added hydrous phase affects the melting relationships of minerals and has important influence on Al, Fe, Ti, and Na abundances in SSZ magmas (cf. Fryer et al., 1981). "Melting of the mantle wedge is induced by the influx of H_2O released from pressure-sensitive dehydration reactions in hydrous peridotite as it is dragged downwards by subduction along the base of the mantle wedge—preferential transport of elements soluble in H_2O-rich fluids gives rise to the distinctive geochemical characteristics of subduction zone magmas and their mantle wedge source regions" (Tatsumi and Eggins, 1995). Much of the original fluid is lost by compaction and dewatering of sediments at the onset of subduction; progressive dehydration occurs as the subducted material is heated and compressed. Much of the subducted fluid escapes and is returned to the ocean; some is taken up by newly formed metamorphic minerals (cf. Peacock, 1990). Hydrous minerals (e.g., zeolites, chlorites, micas, serpentines, epidote, chloritoid, amphiboles, lawsonite, and clinohumite) persist to different pressures and temperatures, but some are stable, and fluids may be retained to depths of 80–100 km, depending on thermal gradients and rate of heat transfer. Calcite may be an effective medium for deep subduction of CO_2. The progressive dehydration of hydrous phases transfers the low ionic potential elements to the depleted mantle peridotite.

Sediment subduction may be important in some settings, especially those having large accretionary prisms (e.g., Sunda Arc), but most intra-oceanic subduction systems lack large accretionary prisms of sediment. This is likely the result of minimum input of sediment to the trench and effective subduction of the small amount of sediment that is present. Subduction of small amounts of sediment need not result in delivery of sediment to depths and temperatures where the (nearly anhydrous) sediment would melt; however, direct or indirect (via fluids) contributions may explain some SSZ Sr and Pb isotope data. For example, Meijer (1976) proposed that Pb isotope data for Mariana Arc lavas may be explained by <1% sediment contribution. Pb data for the Tonga Arc may permit <2% sediment contribution (Oversby and Ewart, 1972; Ewart and Hawkesworth, 1987). Sr isotope data for both arcs are consistent with similar small contributions, although fluid phase transfer rather than direct melting of sediments is highly probable. For the Tonga Arc, influx of more radiogenic mantle sources may be involved (Hergt and Hawkesworth, 1994). Cosmogenic ^{10}Be (half-life 1.6 Ma) has proved useful to show sediment contributions in many arcs (Tera et al., 1986; Ben Othman et al., 1989; Morris et al., 1990). The short half-life requires rapid transfer from ocean sediments, where it accumulated from atmospheric precipitation, to deeper levels having low ^{10}Be, where it is incorporated into the magma source. The transfer mechanism is debated, but many geoscientists consider that Be is relatively insoluble, and thus immobile, at near surface T and P (Leeman, 1996). Hydrothermal experiments (You et al., 1996) show that Be is mobile in fluids at 300 °C, thus obviating the need to invoke sediment melting to explain Be data. Sediments clearly play a role in SSZ magmatism, but most data support a role for sediment-derived fluid, rather than direct melting of sediment (e.g., Stern et al., 1991); however, the aqueous fluids cannot transport HFSE. Melting is required for that.

Partial melting of the subducted ocean lithosphere must be considered as another source and, in view of its "limitless" supply, a potentially significant one; however, many geochemical data indicate that the subducted lithosphere rarely is a source of melts in most subduction systems (cf. Gill, 1981). Melting of the subducted basalt would mobilize HFSE, but SSZ magmas are low in HFSE. In some places, a case can be made for melting of young, and hot, subducted ocean crust (cf. Defant and Drummond, 1990; Maury et al., 1996), but the resulting magmas (adakites) are rare in the intra-oceanic SSZ setting (e.g., they should have exceptionally high Sr/Y and La/Yb, and show HFSE enrichments). Thermal configurations of subduction systems apparently limit the situations in which low-P melting of the subducted slab can occur to settings where very young, hot, crust is subducted. Evidence for high-P melting (eclogite field of stability) in oceanic arc settings is not supported by rare-earth element (REE) data, which should show strong negative slopes on chondrite-normalized plots. It is unlikely that melting of subducted oceanic crust plays a major role as a magma source for the intra-oceanic SSZ, although it may be important in magmatic arcs of Andean-type margins.

Mantle counterflow. Counterflow of hot mantle above the subduction zone (cf. Fig. 2) was proposed by Holmes (1931, 1965) "to illustrate a speculative arrangement of convection currents which might account for the origin of andesitic lavas….and for the concentration of volcanoes over the region of intermediate earthquakes." See also a discussion by Allwart (1988). Most modern ideas about the dynamics of the SSZ mantle utilize some form of convecting mantle (cf. Ewart and Hawkesworth, 1987). Sleep and Toksoz (1971), Toksoz and Bird (1977), and Davies and Stevenson (1992) proposed that counterflow responsible for extension and arc magmatism is a consequence of viscous drag induced by the subducted ocean lithosphere. Alternatively, the driving force may be upward circulation of hot mantle driven by descent of cooled mantle (Ida, 1983) or a passive response to seaward retreat of the subducted lithosphere accompanying rollback of the hinge (Hamilton, 1995, 2002).

The entrained "new" mantle may be N-MORB source or more enriched material, such as a source for E-MORB or OIB magmas. Evidence for mantle counterflow into the SSZ comes from isotope studies of Western Pacific Basin systems (e.g., Hergt and Hawkesworth, 1994; Hickey-Vargas et al., 1995), that show that young SSZ systems having Pb and Sr isotope systematics more typical of an "Indian" MORB source have replaced older sources with "Pacific" MORB source isotope signatures. For SSZ systems with active backarc basins, this convective overturn of mantle above the WBZ is significant because it brings in mantle that is fertile enough to generate the MORB-like melts characteristic of backarc basin crust (Hawkins, 1995a, 1995b). Most models envision counterflow in a direction opposite to the incoming subducted plate; however, complex flow under the Lau Basin is indicated by seismic anisotropy (Smith et al., 2001). Their data suggest that in addition to trench-normal flow directions, there is north-south flow. Independent evidence for the latter is found in ^{3}He/^{4}He data showing high ratios at some northern Lau Basin seamounts, which may represent a "plume" component from Samoa (Poreda and Craig, 1993).

Extraction of MORB magmas from the convectively overturned mantle beneath backarc basins leaves a partially depleted mantle residue that may provide continuous influx of mantle that is still capable of making melts to form island arc tholeiite basalts and basaltic andesite (cf. Ewart and Hawkesworth, 1987).

BACKARC BASINS

Introduction

It is my thesis that few, if any, ophiolites represent fragments of true mid-ocean ridges. The Macquarie Ridge ophiolite (Varne et al., 2000) is a rare occurrence that clearly is derived from crust formed at a mid-ocean ridge, but there are few other unequivocal examples. In the case of Macquarie Ridge, rock chemistry ranges from N-MORB to E-MORB and all rocks have relatively high alkali metal content. Few ophiolites have this enriched chemistry, but in this location it has been emplaced by crustal contraction on a former fracture zone and clearly represents an ophiolite assemblage.

If most ophiolites are not from a mid-ocean ridge, then from where do they come? The geology of backarc basins may be the key to understanding the source for the MORB-like crust of many ophiolites. In this section, I summarize the petrologic characteristics of backarc basin crust and show its close geochemical and mineralogic similarity to crust formed at mid-ocean ridges. I use the Lau Basin and Mariana Trough as type examples, although any of the backarc basins shown on Figure 1 would suffice. The North Fiji Basin (Price et al., 1990) and Bransfield Strait (Weaver et al., 1979) are nearly identical in terms of petrology and crustal fabric. Backarc basin crust forms at spreading centers that, in many backarc basins, erupt tholeiitic melts that have mineral and chemical signatures ranging from N-MORB to island arc tholeiite (Weaver et al., 1979; Hawkins and Melchior, 1985; Hawkins, 1995 a, b; Gribble et al., 1996, 1998). Price et al. (1990) showed that backarc lavas from the North Fiji Basin range in composition from N-MORB to rocks more enriched in LILE and having "transitional alkalic chemistry (up to 0.5% Ne in the Norm)." This range in rock chemistry may be related to the maturity of the basins, which in turn may be owing to the relative importance of subduction-derived fluids in the mantle source. In their early stage, basalts retaining an SSZ signature may form, but as the backarc system evolves, the main magma becomes basalt similar to MORB in nearly all respects. The early SSZ or "alkalic" signature (cf. Price et al., 1990) may be a consequence of source heterogeneity on various scales that retain memory of the SSZ setting (e.g., Volpe et al., 1987, 1988, 1990). An alternate explanation, well developed by Gribble et al. (1996) for the southern part of the Mariana Trough, is that the source is fluids derived from the subduction component. The southern Mariana Trough lies 40–60 km west of the trace of the steeply inclined seismic zone; this suggest that the the subduction-derived fluids must be entrained in convecting mantle material.

It is important to note that relatively "mature" backarc basins such as the Lau Basin presently erupt MORB on the axial ridges, whereas in the Mariana Trough, we find a range from arc-like (IAT) basalt to MORB on the present axial ridge. Both basins have about the same chronologic age, but clearly have different mantle sources (Hawkins, 1995 a, b). Can this be related to differences in the geometry of the WBZ, as discussed above? There is no *a priori* reason why magma evolution must follow a common pattern; mantle source composition, including the nature of mantle convectively entrained above the WBZ, must be a controlling factor.

Seamounts in the backarc basins range in composition from island arc chemistry to ocean island basalt. Backarc basins may be blanketed with arc-derived clastic sediments on their flanks. This variety of geologic features is part of the key to recognizing the backarc basin component in SSZ ophiolites. I postulate that if crust from the Lau Basin or Mariana Trough were incorporated among rocks of an orogenic belt, they would be identified as ophiolite and, if considered out of context with related rocks, might well be interpreted as having formed at a mid-ocean ridge spreading center. Important to the study of ophiolites is the observation that backarc basins may have heterogeneous compositions in which MORB may dominate, but in which both arc-like and OIB-like rocks also are present.

Lau Basin—Form and Geological Setting

The Lau Basin separates the Lau Ridge remnant arc from the active Tofua (Tonga) Arc (Fig. 5). It has a trapezoidal shape, with the Tofua Arc and Tonga Trench on its eastern side. The Tonga Trench curves sharply westward along the northern edge, where it becomes part of the proposed transform system linking the Tonga and Vanuatu-New Hebrides trenches. The backarc basin narrows southward to near 26° S, where it becomes a narrow basin, the Havre Trough, which is the backarc basin of the Kermadec trench-arc system. Farther south, the Havre Trough merges with the Taupo graben of New Zealand. The shape of the Lau Basin may in part be a consequence of backarc crustal extension being hindered by the thickened lithosphere of the Louisville Seamount Chain, which enters the Tonga-Kermadec Trench near 26° S. Rapid extension (165 mm/yr^{-1}, Bevis et al., 1995) at the unconstrained north end of the Lau Basin is a consequence of rapid trench rollback manifested as crustal extension on both the North Lau spreading center (NLSC) and the Mangatolu Triple Junction (MTJ). The central part of the Lau Basin, near 19° S, is situated ~250 km above the seismic zone (Isacks and Barazangi, 1977) and above a zone of extremely low-Q (seismic attenuation) that presumably is because of elevated mantle temperature (Barazangi and Isacks, 1971; Zhao et al., 1997). Seismic refraction studies (Crawford et al., 2003) show that the crust of the basin varies from 5.5 to 6.5 km thick on the east side, with a relatively thick "sheeted dike" section (2–3 km) overlying a thin section (2 km) of lower crust. The crust thickens to the west, being 7–8 km thick near the Central Lau spreading center (CLSC) and 8.9–9.5 km thick at ~50 km west of the CLSC.

The Lau Basin began to open at ~6.5 Ma, as determined from ODP Site 834 data (Hawkins et al., 1994). At that time the Lau Ridge volcanic arc (>14 Ma to 2.5 Ma) was still active. The forearc was extended and small rift basins began to open. At ~4.5 Ma, a southward propagating ridge—the present East Lau spreading center (ELSC)—began to replace former crust with new seafloor having N-MORB composition and developing symmetric magnetic anomalies. A second propagator started at 1.5–1 Ma and formed the CLSC, which also has symmetric magnetic anomalies. At about this time, the North Lau spreading center (NLSC) and the complex of three separate ridges that radiate from the MTJ also formed. As discussed below, all of these spreading centers are dominated by N-MORB-type chemistry, although the MTJ lavas show a subduction component (Table 1). All but the CLSC have a distinctive depletion in Nb and Ta, which points to their SSZ heritage. Lau Basin seamounts and some of the seafloor include both arc-like lavas, as evidenced by a strong SSZ signature in rocks and detritus (e.g., ODP Site 839, Table 1), as well as some have an enriched or E-MORB signature (e.g., Rochambeau Bank and Niua Fo'ou Island, Table 1).

The western side of the Lau Basin has many small rift basins, up to 25 km long and 8–14 km wide, partly filled with volcaniclastic gravity flow deposits (turbidites) derived from local high-standing fragments of rifted arc crust or nearby submarine arc-composition volcanoes (Parson et al., 1994). The basins have on the order of 100 m of fill overlying basaltic basement. Lower in the section, clastic sediments are interbedded with and intruded by tholeiitic flows and sills having the general

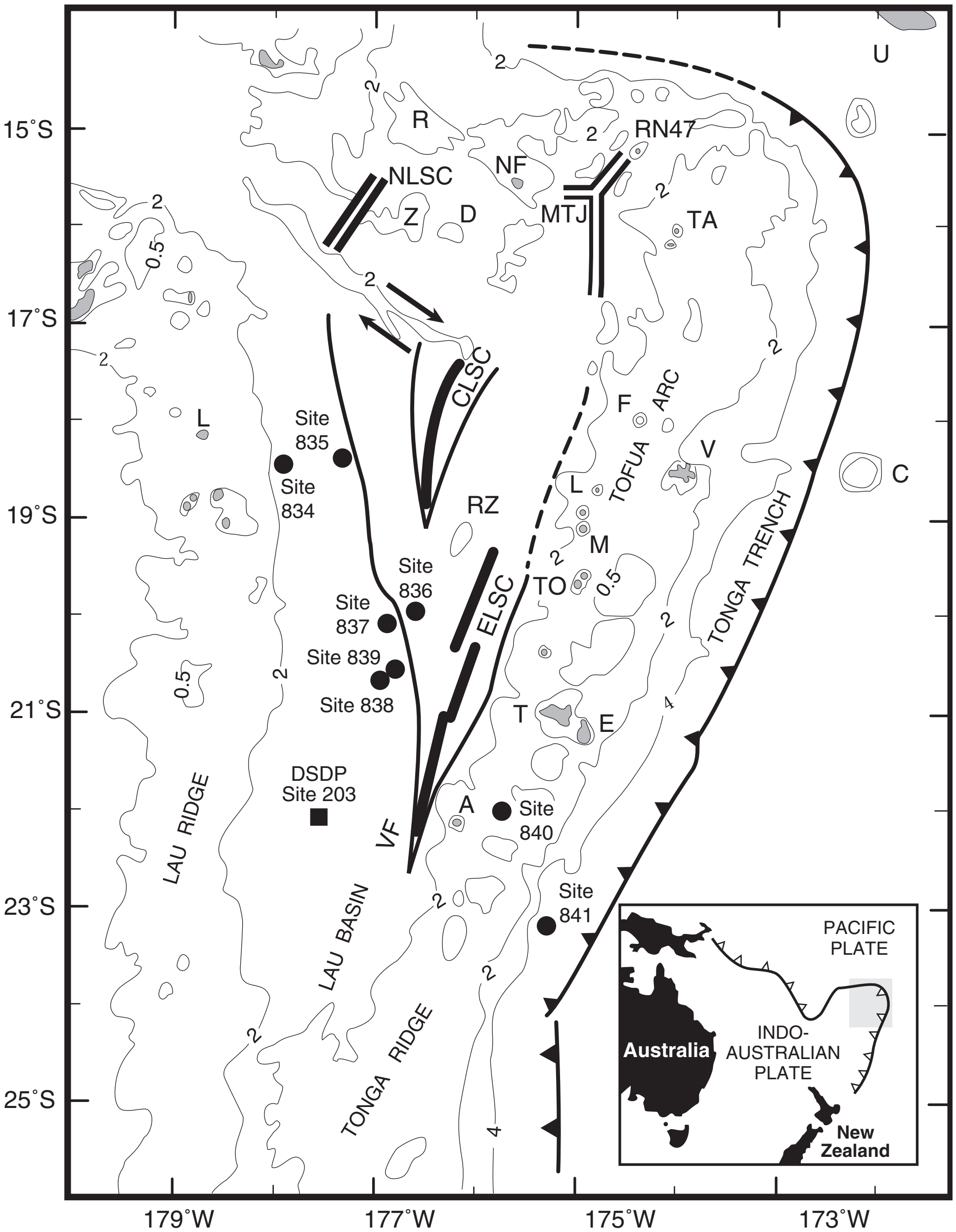

Figure 5. Map of Lau-Tonga arc-trench system, showing features discussed in text. Active spreading centers are: ELSC—East Lau Spreading Center; CLSC—Central Lau Spreading Center; RZ—relay zone between them; NLSC—North Lau Spreading Center; and MTJ—Mangatolu Triple Junction. Sites 834–841 are ODP Leg 135 drill sites; DSDP 203 is from DSDP Leg 21. Seamounts are: D—Donna Seamount, R—Rochambeau Bank, Z—Zephyr Shoal, and NF—island of Niua Fo'ou. Features on Tonga Ridge are: volcanic islands, TA—Tafahi; RN 47—an unnamed seamount; L—Late; F—Fonualei; M—Metis Shoal; TO—Tofua; A—Ata; and raised coralline-capped platforms, V—Vava'u; and T—Tongatapu. Other features are E—'Eua Island; U—Upolu, Samoa; and C—Capricorn Seamount.

mineralogy and chemical composition of MORB. The volcaniclastics are overlain by hemipelagic clay-nannofossil oozes stained by ferro-manganese oxy-hydroxide (Parson et al., 1994; Rothwell et al., 1994). In the central part of the Lau Basin, the igneous basement has only a discontinuous cover of pelagic and hemi-pelagic sediments, many of which in their basal section are heavily stained with ferro-manganese oxyhydroxides from hydrothermal circulation at the axial ridge (Hodkinson and Cronan, 1994). Arc-derived volcaniclastic rocks from the Lau Ridge and the Tofua Arc partially blanket both sides of the basin and partly fill the rift basins (Clift et al., 1998).

Mariana Trough—Form and Geologic Setting

The Mariana Trough (Fig. 4) separates the West Mariana Ridge remnant arc from the active Mariana volcanic arc. The age of initial opening is uncertain, but the best estimate for the initiation of rifting is ca. 10 Ma (Fryer, 1992). The crescent shaped Mariana Trough narrows northward to ~25° N, where its extension may be pinned by thickened crust of the Magellan seamounts that lie outboard of the Mariana Trench. The ~200 km wide south end of the Mariana Trough is unconstrained, and basin depths descend to the ~10 km-deep, west-trending, part of the Mariana Trench.

TABLE 1. LAU BASIN

Location:	ELSC	ELSC	ELSC	ELSC	VALU FA	CLSC	CLSC	CLSC	CLSC	CLSC
Site:	ODP 834	R-5	R-11	R-7	VF-1	ST-64	R-26	R-25	P-33	R-13
Sample:	3H7	RNDB	RNDB	RNDB	Lee	STO	RNDB	RNDB	PPTU	RNDB
	54-55	5-3 GL	11-1 GL	7-2 GL	1-5	64-3 GL	26-2 GL	25-1 GL	33-1 GL	13-2 GL
Mg#:	69.3	62.9	61.8	35.6	39.9	66.3	63.3	59.5	45.5	25.2
					Wt%					
SiO_2	47.13	51.44	50.62	52.58	56.69	48.58	50.25	50.96	50.45	60.41
TiO_2	0.69	0.90	1.02	2.11	1.38	0.70	0.94	1.05	1.93	1.62
Al_2O_3	19.18	15.44	15.18	13.43	14.57	16.85	15.86	14.65	14.11	12.93
FeO*	8.08	9.39	9.54	14.76	11.51	9.54	9.79	10.6	13.29	12.16
MnO	0.16	0.19	0.18	0.21	0.22	0.19	0.19	0.20	0.25	0.22
MgO	8.88	8.02	7.51	3.99	3.73	8.93	8.25	7.60	5.41	1.99
CaO	13.03	12.87	12.83	8.55	7.65	12.77	13.06	12.76	10.17	6.15
Na_2O	1.90	1.91	1.90	3.24	3.36	2.42	2.05	2.26	2.85	3.91
K_2O	0.01	0.03	0.13	0.14	0.52	0.01	0.07	0.03	0.08	0.42
P_2O_5	0.14	0.06	0.12	0.20	0.16	0.03	0.08	0.07	0.13	0.32
Total	99.20	100.25	99.03	99.21	99.79	100.02	100.54	100.18	98.67	100.13
					Trace elements (ppm)					
Cr		337	342		4	430	325	303	20	
Ni	111	110	103	8	25	189	149	116	37	7
Co		43	43			70	43		44	
V	216	307	338	502	350	202	323	332	455	147
Zr	47	53	57	127	81	33	51	45	128	441
Y	19	27	26	55	36	20	2929	30	59	138
Nb	2	1	1	1		2	2	1	1	
Hf		1.16	1.55		2.3			1.89	3.57	
Ta		<.05	0.08					0.04	0.16	
Rb	0.1	1	6	4	8.6	0.5	0.3	0.1	1	10
Ba	17	4	8	27	96	6.2	4	2	2	75
Sr	140	69	111	95	168	62	63	49	114	86

Locations: ELSC—East Lau Spreaading Center; CLSC—Central Lau Spreading Center

(*continued*)

The axial ridge of the Mariana Trough is asymmetric, offset to the east of the center of the basin; however, the axis lies slightly to the west of the deepest part of the inclined seismic zone, which here is nearly vertical (Isacks and Barazangi, 1977). Fryer (1992) recognized an early stage of rifting of small basins, followed by a later stage of seafloor spreading, similar to the early stages of evolution of the Lau Basin. Fryer (1995) interpreted the fabric of the northern Mariana Trough as representing an early stage of rifting, whereas the central region represents a more mature stage typical of slow spreading. The crustal fabric is dominated by north-south-trending abyssal hills. The western side of the basin, next to the remnant arc, is complexly faulted with numerous rift basins that record the earliest stage of opening (Fryer, 1992; Martinez et al., 1995). Cross-chains of arc volcanic rocks cut though the modern volcanic arc and transect the axial ridge.

A significant difference between the Lau Basin and the Mariana Trough is that the axial ridge of the slowly spreading Mariana Trough is a relatively simple feature, apparently propagating northward and having only minor offsets of the axis, whereas the fast spreading Lau Basin has six separate spreading ridges, three of which meet at a triple junction, and contains evidence for sequential development of propagating rifts since inception of rifting at ~6.5 Ma. Both sides of the Mariana Trough are blanketed with arc-derived clastic rocks and, as in the Lau Basin, the sediments lie on nearly coeval seafloor basalts. The occurrence of arc-derived clastic sediments, deposited on back-

TABLE 1. LAU BASIN (*continued*)

Location:	RZ	WB	WB	WB	CB	NLSC	MTJ	MTJ	Donna	Rochambeau
Site:	R-31	834B	834B	835	839A	P-19	R-17	R-19	Seamount	Bank
Sample:	RNDB	29R1	56R1	6R1	24x1	PPTU	RNDB	RNDB	123	PPTU
	31-4GL	/07-12	/01-15	/32-37	/00-06	19-1 GL	17-2 GL	19-1 GL	89-4 GL	23-2
Mg#:	68.3	64.6	5-Feb	65	67.1	69.1	27.2	64.4	53.5	54.5
					Wt%					
SiO_2	49.25	50.17	52.20	50.35	53.03	49.17	59.98	50.89	50.53	48.88
TiO_2	0.79	1.42	2.11	1.09	0.65	0.82	1.37	0.94	1.26	1.89
Al_2O_3	17.36	16.33	15.71	16.47	15.06	17.15	15.16	15.93	14.30	14.94
FeO*	8.65	8.67	12.41	8.46	9.04	8.72	10.50	8.53	11.97	11.35
MnO	0.17	0.20	0.20	0.16	0.17	0.18	0.23	0.17	0.25	0.19
MgO	9.10	7.73	3.40	7.67	9.00	9.51	1.91	7.52	6.73	6.63
CaO	12.82	11.87	8.58	12.90	11.66	12.75	5.62	13.10	11.72	11.79
Na_2O	2.20	3.08	3.87	1.82	1.35	2.11	3.89	2.14	2.88	3.10
K_2O	0.03	0.09	0.73	0.28	0.23	0.03	0.97	0.21	0.08	0.26
P_2O_5	0.03	0.11	0.21	0.11	0.07	0.04	0.42	0.11	0.12	0.17
Total	100.40	99.67	99.42	99.31	100.26	100.48	100.05	99.54	99.84	99.20
					Trace elements (ppm)					
Cr	334		0	179	618	741	39	309	150	238
Ni	163	102	12	78	123	312	6	90	45	72
Co	44					56	15	41	50	47
V	212	227				212	51	226	334	355
Zr	68	102	146	44	32	44	238	64	77	127
Y	24	29	40	22	16	20	78	24	32	31
Nb	3	3	1	1	1	1	9	1	3	7
Hf	1.14					1.15	6.09	1.59		3.27
Ta	<.05					0.045	0.65	0.133		0.375
Rb	1	3	13	5	5	0.3	27	7	16	5
Ba	3	22	60	42	78	1	147	27	11.1	40
Sr	93	176	195	130	148	79	135	133	90	215

Locations: WB—Western Basin; CB—Central Basin; NLSC—North Lau; MTJ—Mangatolu Triple Junction.

arc crust of essentially the same age, is an important feature of arc-backarc systems and important in recognizing an SSZ origin for some ophiolites. In the case of the Lau Basin, the Lau Ridge "remnant arc" remained active for ~4 my after initial rifting that formed the Lau Basin, and overlapped in time the volcanism on the Tofua (Tonga) Arc.

Petrology and Chemistry of Backarc Basins

Introduction

The primary rock type forming the crust of both the Lau and Mariana Trough backarc basins is tholeiitic basalt that is similar in most respects to N-MORB (Table 1). For detailed discussions of data supporting this claim, see Hart et al. (1972); Hawkins (1974, 1976, 1977, 1994, 1995a, b); Hawkins et al. (1984, 1990); Fryer et al. (1981); Sinton and Fryer (1987); Volpe et al. (1987, 1988, 1990); Stern et al. (1990); Falloon et al. (1992); Pearce et al. (1995); and Gribble et al. (1996, 1998). As discussed below, both basins have a range in rock composition that, in addition to N-MORB, includes IAT, backarc basin basalt (BABB, cf. Fryer et al., 1981), and enriched basalts that include OIB, as well as fractionated rocks of each of these series.

In both basins, the freshest samples occur on the axial ridge on presumed "zero age" crust and on seamounts close to the ridge axes. Axial ridge samples collected with ALVIN in the Mariana Trough were picked from fresh accumulations of pillows having essentially no sediment cover (Hawkins et al., 1990). Deeper crustal levels sampled at DSDP and ODP sites include older and more altered material, and some off-axis dredge sites at depths >4000 m also have altered rocks. Nearly all dredged rocks are wedge-shaped fragments of pillow basalts, many of which have rinds of fresh vitrophyre; minor amounts of microdiabase have been dredged, but these are more common in drill cores, and represent dikes, sills, or interiors of large pillows.

Gabbro is rare from back-arc basins, but one Lau Basin site (ANT 233, for details see Hawkins, 1976) yielded plagioclase rich, clinopyroxene gabbro with primary brown hornblende. DSDP Site 453, on the western edge of the Mariana Trough, recovered plagioclase-rich, two-pyroxene gabbro and mylonitized gabbro; Natland (1981) interpreted these rocks as likely to be from the remnant arc. Stern et al. (1996) and Ohara et al. (2002) recovered probable mid- or lower crust and upper mantle rocks at an amagmatic rift basin in the northern Mariana Trough. Ultramafic rocks are even less common; Ohara et al. (2002) recovered serpentinized harzburgite from this same Mariana Trough rift basin. It is mainly olivine (Fo 90.5), with orthopyroxene (En 90). Chrome spinel has Cr# 25–40, where Cr# = 100 × (Cr/Cr+Al) and the rocks have up to 2% modal clinopyroxene. These rocks are depleted mantle residue, but are less depleted than peridotites from forearcs, discussed below, which commonly have spinel with higher Cr#. The deep crustal rocks from this rift basin include anorthosite, diorite, tonalite, and granitoids. The latter have up to 79.7% SiO_2, and may be plutonic equivalents of rhyolitic glasses found in the Lau Basin and discussed below. Arai (1991) proposed that upper mantle rocks of the Circum-Izu Massif peridotite of central Japan represent backarc basin mantle of the Shikoku Basin. It, too, is mainly harzburgite with lesser amounts of dunite and clinopyroxene-poor lherzolite. Olivine in harzburgite is Fo 90–92 and in dunite is as high as Fo 94; the Cr# for chrome spinel ranges from 20 to 70, but is mostly 40 to 60. Some rocks have minor plagioclase ranging in composition from bytownite to anorthite. The spinel is unusual in having hydrous inclusions of pargasite or phlogopite + orthopyroxene. The peridotites have low-Ti, like many depleted peridotites but are unusual in having primary amphibole and mica. These minerals are never found in "abyssal" peridotites and have not been reported in depleted peridotites of trenches. Arai concludes that the peridotite composition is consistent with an SSZ origin.

The crystallization sequence commonly found in backarc basin rocks, as determined from vitrophyres and crystalline rocks, is: olivine + chrome spinel-plagioclase-clinopyroxene-iron-titanium oxides, just as we find for MORB. Orthopyroxene is not found in basalts. We can infer that backarc basin basal cumulates are likely to be clinopyroxene-bearing gabbros, or troctolite (i.e., plagioclase and olivine), rather than norites (olivine and orthopyroxene) or wehrlites (olivine and clinopyroxene), which are typical of arc series.

Backarc basin magma trace element chemistry (Hawkins et al., 1990) shows a small range from arc-like basalt (IAT) to N-MORB, with variable depletion in HFSE relative to N-MORB (e.g., Ti, Zr, Y, Hf, heavy rare earth elements, and Nb, and Ta in particular, Woodhead et al., 1993). These depletions in elements considered to be incompatible indicate magma derivation from a highly depleted mantle source. The LILE (e.g., K, Rb, Cs, Ba, Sr, and, to a lesser extent, La, Ce) show varied enrichments relative to N-MORB stemming from addition of low ionic potential elements mobile in aqueous fluids (cf. Tatsumi and Eggins, 1995). The presumed source of the LILE and fluid phase is altered ocean crust and subducted sediments.

Thus, in backarc basins we have evidence for samples petrographically and chemically equivalent to presumed oceanic seismic layers 2a and 2b, and a few samples representative of layer 3, and depleted upper mantle. These rocks are petrographically comparable to their putative "equivalents" in ophiolites. In addition, metalliferous sediments and hydrothermal vents with polymetallic sulfide deposits have been found in both basins (Hawkins et al., 1990; Hawkins and Helu, 1986; von Stackelberg et al., 1988).

Lau Basin. The early eruptions of backarc magmas in the Lau Basin at ~6.5 Ma were in rift basins seaward of the (still active) Lau Ridge. They were erupted in what then was the forearc of the Lau Ridge. At ODP Site 834, we drilled through 112.5 m of sediments with clayey nannofossil ooze overlying pyroclastic airfall ashes and epiclastic turbidites, and then 323 m into basalt having 13 texturally different units having aphyric to diabasic textures. These comprise a series of basalt flows or sills with rare interbedded layers of silts and clays presumed to have come from the Lau Ridge volcanic arc. Table 1 has representative data for Site 834 and other features discussed below. At Site 835, on crust ~3.4 m.y. old and in a setting like that of Site 834, we recovered similar sediments and basalt. The distribution of rock types in the Lau Basin suggests that the early eruptions of basalt at ODP Site 834 and 835 are MORB in most respects, but have a distinctive arc-like depletion in HFSE, especially Nb and Ta, as seen in the abundances of trace elements normalized to values for N-MORB (Fig. 6A). Site 835 is even more depleted in HFSE than are the older rocks of Site 834. Somewhat younger samples at Sites 837 and 838 (Fig. 6B) are even more "arc-like" than the oldest samples. In contrast, the propagating rift that formed the ELSC (Fig. 6C) has trace element values more like true N-MORB, suggesting that MORB-source mantle had mixed with older SSZ mantle sources. Sample VF-1 (Table 1) is a differentiated rock (andesite) from the propagating rift tip of ELSC (Valu Fa). The rocks from CLSC and NLSC (Fig. 6D) have a strong MORB pattern. Basalts from Sites 834 and 835 have Pb and Sr isotope ratios that are indicative of a "Pacific" mantle MORB source, in contrast to other Lau Basin rocks, which have ratios typical of "Indian" MORB-source mantle (Hergt and Hawkesworth, 1994). The data suggest that convective influx of relatively fertile (Indian mantle-type) MORB-source has mixed with and swamped the original strongly depleted SSZ "Pacific" mantle (cf. Fig. 2). This retention of an arc signature in early melts is not necessarily a general evolutionary path for backarc basins,

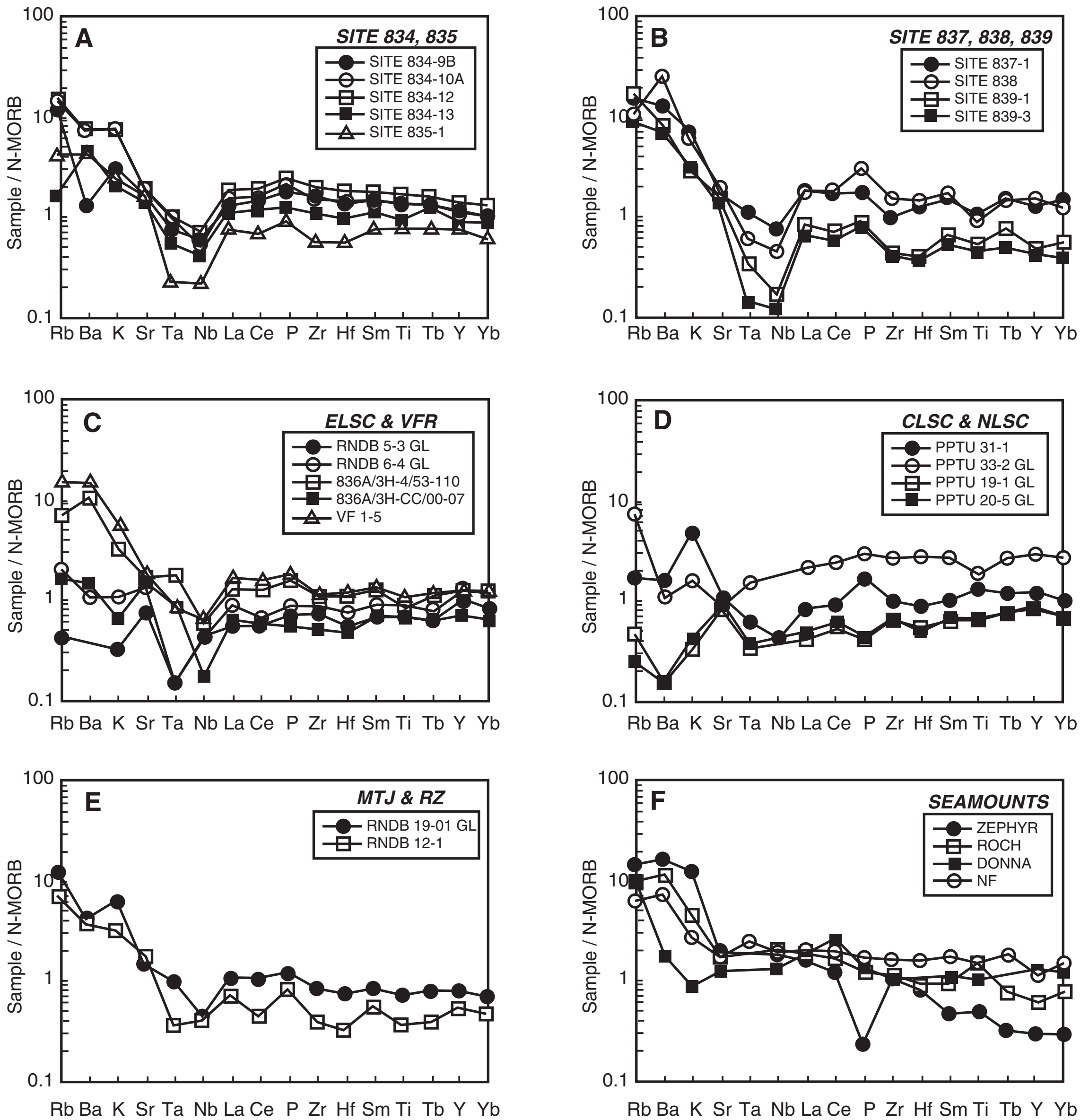

Figure 6. Trace elements for Lau Basin rocks normalized to N-MORB values (Sun and McDonough, 1989). Locations of ODP sites and spreading ridges are in Figure 5. SSZ magmas are characterized by variable LILE enrichments and HFSE (especially Nb and Ta) depletions relative to N-MORB. A: Data from ODP Sites 834 and 835 are from crust formed in rift basins near Lau Ridge remnant arc early in opening of the basin (~6.5–5 Ma). B: Sites 837, 838, 839 are from younger crust on west side of ELSC. C: Data from ELSC and its propagating rift tip (VF-1). Most ELSC data are close to N-MORB abundances, but Nb and Ta depletions are retained. D: CLSC, samples PPTU 31–1 and 32–2, and NLSC, samples PPTU 19–1 and 20–5 have MORB-like abundances with only minor variations in Rb, K, and Ba. E: RNDB 12–1 is from the relay zone between CLSC and ELSC and shows arc-like depletions in HFSE and enriched LILE. Mangatolu Triple Junction, sample RNDB 19–1, is from within 80 km of Tofua arc and shows pattern similar to relay zone. F: Data for enriched MORB of Niuafo'ou Island (NF) and Rochambeau Bank (ROCH); Donna Seamount sample is altered MORB crust, whereas Zephyr Shoal sample is a hybrid of silicic glass and mafic phenocrysts.

as shown by the presence of N-MORB-like basalt in the nascent Sumisu Rift of the Izu system (Fryer et al., 1992).

The Mangatolu Triple Junction (RNDB 19–01) and Relay Zone (RNDB 12–1) basalts (Fig. 6E) show the SSZ pattern. That both are from within ~50 km of the Tofua arc must reflect an arc source contribution. Seamount data (Fig. 6F) for Rochambeau Bank (ROCH) and the island of Niua Fo'ou (NF) are relatively enriched in LILE, but have MORB-like HFSE; these probably reflect an E-MORB contribution. Donna Seamount is a large uplifted block of moderately altered MORB-like seafloor, whereas Zephyr Shoal is a hybrid of rhyolitic glass and mafic phenocrysts.

The important point to be made from these data is that in backarc basins, some seafloor mimics mid-ocean ridge crust in those elements (HFSE) least likely to be modified by low-grade metamorphism or alteration, whereas in other parts of the same backarc basin, the HFSE may have a more arc-like signature. These should be important criteria for determining the provenance of ophiolites.

Mariana Trough. Mariana Trough basalts also are broadly similar to N-MORB in terms of mineralogy, major elements, HFSE, and isotope ratios for Sr, Nd, Pb, He, and O (Fryer et al., 1981; Sinton and Fryer, 1987; Hawkins et al., 1990; Volpe et al., 1987, 1990; Gribble et al., 1996, 1998). Element ratios for both HFSE and LILE resemble MORB, but the LILE and volatiles, especially H_2O, are variable, with many being higher than for MORB. Typical values for H_2O in N-MORB are close to 0.2%, whereas for the Mariana Trough, Stolper and Newman (1994) determined a range for H_2O from 0.49% to 2.1%; the CO_2 range for the same samples is from <10 ppm to171 ppm. The data display a negative correlation between CO_2 and H_2O and a negative correlation between depth of samples and H_2O. Garcia et al. (1979) determined an average H_2O value of 1.02% and 0.21% for CO_2 for a different set of samples.

Some chondrite-normalized REE patterns differ from MORB in having flat to slightly enriched light rare earth elements (LREE) patterns (e.g., La/Sm_{ef} ~1; Hawkins et al., 1990). Samples were divided into N-MORB type and Mriana Trough Basalts (MTB) type (Hawkins et al., 1990) on the basis of volatile content, LILE, and ratios of oxides of Al, Ca, Ti, and Fe relative to MgO (cf. BABB of Fryer et al., 1981; Sinton and Fryer, 1987). For 177 vitrophyric samples having Mg# of 60 to 70, the two types are approximately equally represented. Some representative data are in Table 2, where glass analyses were averaged for different

TABLE 2. MARIANA TROUGH

Sample:	MARA	MARA	MARA	MARA	MARA	MARA	MARA	ALV
Type:	N	N	N	M	M	M	"ARC"	1846-9
Mean:	n=50	n=42	n=26	n=43	n=33	n=10	n=10	
	Mg# >65	Mg# 60–65	Mg <60	Mg# >65	Mg# 60–65	Mg <60	Mg# 63–68	Mg# 67
				Wt%				
SiO_2	50.63	51.38	52.67	50.79	51.52	54.51	50.15	49.67
TiO_2	1.13	1.41	1.23	1.01	1.21	1.75	0.58	0.57
Al_2O_3	17.21	16.34	16.51	17.41	16.71	16.05	17.37	17.45
FeO*	7.64	8.58	8.57	7.28	7.98	9.53	7.48	7.03
MnO	0.15	0.18	0.18	0.15	0.16	0.19	0.16	0.16
MgO	7.72	6.94	5.51	7.37	6.59	3.92	6.83	6.95
CaO	11.65	11.11	10.26	11.86	11.19	8.2	13.06	13.46
Na_2O	2.79	3.11	3.03	2.67	2.92	3.21	2.13	2.07
K_2O	0.21	0.25	0.49	0.32	0.36	0.73	0.76	0.71
P_2O_5	0.13	0.17	0.19	0.14	0.17	0.31	0.13	0.13
Total	99.26	99.47	98.64	99	98.81	98.4	98.65	98.2
				Trace elements (ppm)				
Cr	310	230	57	250	260	247		198
Ni	135	110	30	115	95	56		115
Co	35	35	65	30	30	57		69
V	220	250	271	220	250	292		195
Zr	90	100	81	75	95	139		44
Nb	1	2	1	1	2	2		1
Y	26	33	28	24	32	39		15
Ta	0.18	0.19		0.16	0.22			0.054
Th	0.31	0.24		0.4	0.36			0.53
Sr	195	170	222	205	220	193		314
Rb	2.6	1.5	9.3	4.7	4.7	5.4		25
Ba	25	25	69	45	50	38		49
La	4.8	4.6	7.14	5.1	5.1	6.79		8.2
Nd	9.45	8.1	11.33	9.2	10.9	13.42		10.48
Sm	3.1	2.7	3.22	2.8	3.5	4.02		2.26

Note: Blank—no data; Type N—N-MORB; Type M—MTB (BABB).

ranges of Mg# and are separated into N-MORB types (N) and MTB (M). Table 2 also has examples of arc-like lavas from the axial ridge. Element ratios plotted on Harker diagrams (cf. Hawkins, 1995b) show that Mariana Trough axial ridge data from near 18° N define fields different from N-MORB. Mean value for Al_2O_3 is 17% versus an upper limit of 16% for N-MORB, CaO is higher (mean value is 12%), thereby retaining the N-MORB ratio of 0.5–0.65 for CaO/Al_2O_3. The FeO* (mean value of 7%) is higher than for N-MORB at comparable MgO content.

Our ALVIN samples collected along the axial ridge have Mg# ranging from >65 to 50 over distances of 1 to 2 km. Data for these samples are in Table 2. Axial ridge seamount samples labeled "ARC" and ALV 1846–9 on Table 2 have island arc tholeiite chemistry; they are surrounded by crust with MORB chemistry having Mg# = 62. The most fractionated axial ridge sample we found is a basaltic andesite having 56% SiO_2, 4.3% MgO, and Mg# = 36, suggesting that shallow-level fractionation was important in the Mariana Trough, as in other ocean ridges.

Mariana Trough mantle source history may differ from that of the Lau Basin. In the Mariana Trough, we found that both N-MORB type lavas and MTB, a variety of transitional to arc-like lavas, erupted synchronously at closely spaced sites along the axial ridge (Table 2). Volpe et al. (1987, 1990) proposed mixing of sources. At its northernmost part, 22° N, the axial ridge lavas display arc-like chemical and isotopic signatures (Stern et al., 1990).

On backarc magmas. The HFSE abundances and element ratios are the most useful chemical signatures likely to be preserved, even if rocks are altered or have experienced low-grade metamorphism. The cause of the HFSE depletion is a matter of debate, e.g., theoretically, a titanate residual in the mantle helps explain this, but the implicit Ti concentrations for the mantle make this unlikely. The simplest explanation is that the SSZ mantle has been depleted in HFSE relative to MORB. These depletions are imparted to SSZ melts and appear to be diagnostic for the setting. Several classification schemes and discriminant diagrams are useful. Pearce and Cann (1973), Pearce and Norry (1979), and Pearce et al. (1984) used multi-element ratios such as Ti, Zr, Y, and Cr (and Sr for unaltered rocks) to define fields for mafic magma from various sources, including "island arc" and MORB. Shervais (1982) used Ti vs. V to define fields for arc (Ti/V <20), ocean ridge (Ti/V 20–50) and ocean island (Ti/V >50). Cr-Y variations (Fig. 7) prove useful for separating IAT magmas from N-MORB and from boninites. Pearce et al. (1981, 1984) proposed that abundances of a compatible element, such as Cr or Ni, relative to high field strength (high ionic potential) incompatible elements, such as Y or Zr, would indicate effects of fractionation or extent of melting. Cr and Y are "not significantly affected by the processes that cause heterogeneity in the convecting upper mantle." Cr would decrease with melt fractionation because it is accommodated in olivine and pyroxene, whereas Y is excluded and would increase in residual melt; the converse would apply for fractional melting. Data for MORB and island arc lavas show that, for a given Cr content, Y and other HFSE are relatively depleted in island arc lavas. MORB-source mantle is assumed to have >2500 ppm Cr and ~4 ppm Y; primitive N-MORB has ~340 ppm Cr and 25 ppm Y, with a negative correlation between them for fractionated magmas (e.g., 60 ppm Y at 100 ppm Cr). In contrast, melts from IAT settings commonly have markedly lower Y than N-MORB at comparable Cr content (e.g., 10–20 ppm Y). Boninites have high Cr and very low Y relative to Cr. Inasmuch as Cr and Y are relatively unaffected by weathering and low-*T* metamorphism they are useful in understanding the origin of ophiolites, and in particular, help recognize SSZ ophiolites. Discriminant diagrams show that most backarc basin basalts plot in the field for N-MORB, but some have an SSZ signature. The IAT and boninite data require derivation from a source that has lost a basalt component and thus is depleted in Y relative to bulk earth mantle. Figure 7 shows data for the Lau Basin, Mariana Trough, and their nearby contemporary arcs, compared with fields for boninite, island arcs, and MORB. Also shown are data for the Palau-Kyushu Ridge. Note that for the Lau Basin, the ODP drill data from the older, pre-spreading crust partly overlaps the IAT field, whereas most of the axial ridge dredged samples lie in the MORB field. The Mariana Trough data are all from near-zero age crust and lie in the MORB field. The Palau-Kyushu Ridge data include IAT and MORB, as well as some samples that overlap onto the boninite field. Pearce et al. (1981) and Wallin and Metcalf (1998) used the Cr-Y plot to demonstrate an SSZ origin for the Oman and Trinity ophiolites, respectively.

The ternary plot for Hf/3-Ta-Th (Wood et al., 1981) is another useful example (Fig. 8). In Figure 8A, we see that Site 834 data lie in the N-MORB field along with much younger Site 838, whereas Site 835 data lie in the SSZ field and are displaced toward Th, an effect of sediment admixture. The field for the Tofua Arc and RN 47, a primitive Tofua arc seamount, are shown for comparison. Figure 8B shows that Site 839 data, from the central part of the basin, lies on the Tofua field, whereas Site 836, adjacent to the ELSC, plots in the N-MORB field. It is interesting that ODP Site 834 data, some of the oldest Lau Basin crust, plot in the N-MORB field, as do Site 836 data, young rocks drilled close to the ELSC. Dredged samples from the axial ridges (Fig. 8C) plot mainly in the N-MORB field, whereas the MTJ samples trend into the SSZ field. The MTJ is situated within 50 km of the Tofua arc and likely is influenced by the subduction component added to the mantle wedge. Two seamounts show an E-MORB component, but the other ODP drill sites lie in the SSZ field, together with samples from the Tofua Arc and RN 47.

Mariana Trough ALVIN samples and dredged basalts lie partly in the N-MORB field and partly in the SSZ field on Figure 8D. ODP drill data for Site 454 from the west flank of the axial ridge has an SSZ signature, whereas Site 456 on the east flank plots in the N-MORB field. The same diagram show two Mariana forearc sites, DSDP Sites 458 and 459, which plot in the SSZ field.

In summary, backarc basin crust, especially crust of "mature" basins such as the Lau Basin, is dominated by MORB-like

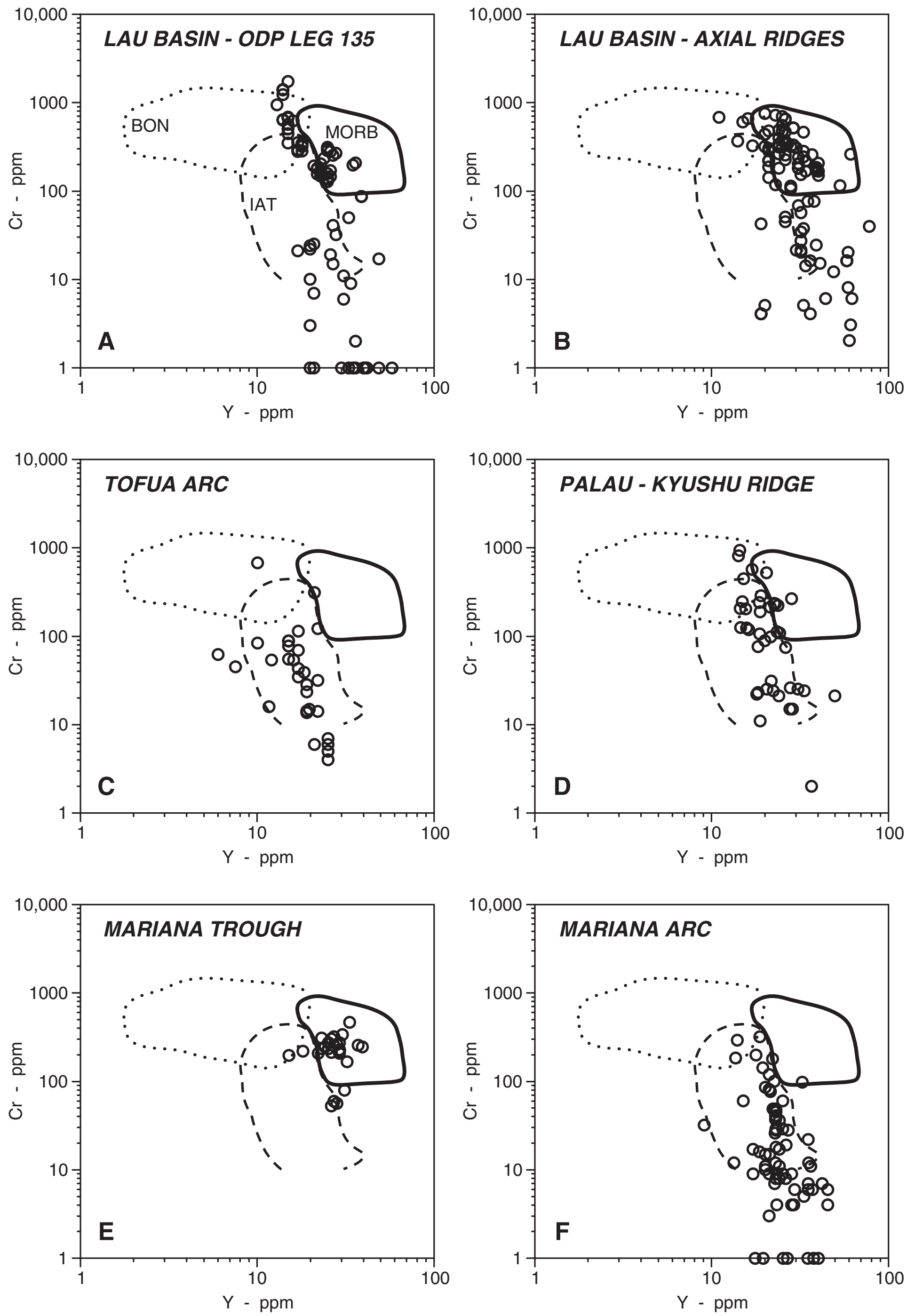

Figure 7. Cr-Y plot after Pearce et al. (1981), showing fields for BON—boninite, IAT—island arc tholeiite, and MORB—mid-oceanic ridge. A: ODP Leg 135, Lau Basin–off axis samples (data from Hawkins and Allan, 1994); B: Lau Basin axial ridge (data from Hawkins and Melchior, 1985; Sunkel, 1990; Falloon et al., 1992; Hawkins, 1995 a, b); C: Tofua Arc (data from Ewart et al., 1973; Ewart et al., 1977; Ewart and Hawkesworth, 1987; Vallier et al., 1985); D: Palau-Kyushu Ridge (data from Hawkins and Castillo, 1998); E: Mariana Trough axial ridge (data from Hawkins and Melchior, 1985; Hawkins et al., 1990); F: Mariana Arc (data from Woodhead, 1989; Peate and Pearce, 1998).

basalt. Ophiolite remnants of these could easily be misinterpreted as having formed at a mid-ocean ridge spreading center; however, many backarc basin basalts also carry the chemical imprint of the SSZ mantle, as seen, for example, in Figure 8 and in comparing relative abundances of HFSE and LILE. As will be shown in a section to follow, many "classic" ophiolites plot in the SSZ or IAT fields on the Hf/3-Ta-Th diagram.

GEOLOGY OF FOREARCS

Introduction

Clearly, the term "forearc" has meaning only if there is a nearby volcanic arc to serve as a reference point. In most modern arc-trench systems, the forearc includes ocean crust

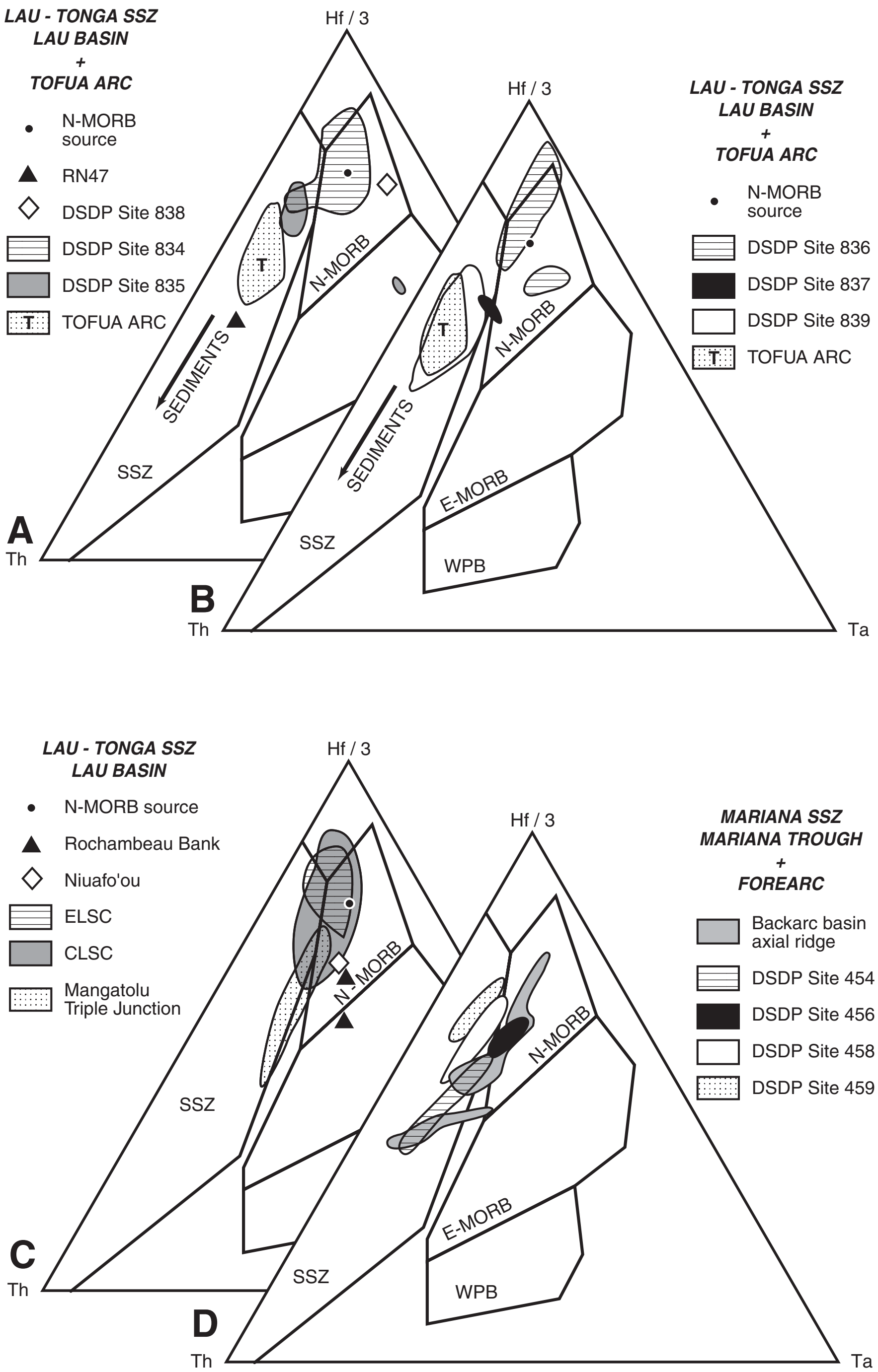

Figure 8. Plot of Hf/3-Ta-Th after Wood et al. (1981), showing fields for N-MORB, E-MORB, WPB (within plate basalts), and SSZ magmas. Arrow labeled "sediment" shows effect of adding Th from sediment. Data for arc and backarc samples are discussed in text.

extending outboard from the volcanic arc as much as 100 km or more to the break in seafloor slope down into the trench. For purposes of this discussion, I will also use it for the seafloor that will become part of the upper oceanic plate upon the initiation of a subduction system. In a simplified view of SSZ geometry, forearcs may be interpreted as trapped fragments of aged, cold, amagmatic, oceanic lithosphere on which island arcs are built. Our studies of Western Pacific Basin SSZ systems show that they are far more complex and interesting. The amagmatic inference is supported by their (present) low heat flow and their position in the thermal setting envisioned for subduction systems (i.e., the cooling effect of penetration of "cold" oceanic lithosphere into the ambient mantle thermal regime; cf. Stein and Stein, 1996). "Isotherms in the slab extend downward such that the maximum depth reached by an isotherm is proportional to the product of the vertical descent rate…and the square of plate thickness." However, this strong downward deflection of isotherms may not always be the case throughout SSZ history. At the onset of subduction, oceanic crust above the nascent subduction zone may be extended and magmas erupted into the extension zone. Hawkins et al. (1984) proposed that subduction might begin on a former, serpentinized, fracture zone trace because of a change in relative plate motion. We suggested that "differences in crustal age, density, and thickness, plus changes in relative plate motion cause[s] the failure along the fracture zone," and that "[mantle] counterflow heats the serpentinized peridotite, causes an uparch of the lithosphere, and promotes dilation of…the forearc," "boninite[s] …are the first melts to leak out." We called this the boninite stage of an "immature arc." Stern and Bloomer (1992), in their more refined model, showed how this helps explain generation of both boninite and high-Mg arc tholeiite magmas that form in the earliest stages of SSZ magmatism, referring to this as an "infant arc." They also proposed that an initial phase of extension, or transtension, would result in foundering of a more dense plate that eventually becomes the subducted plate. R.N. Taylor and Nesbit (1995) presented evidence for the "broad region of volcanism generated in an extensional regime" in the Eocene-Oligocene Izu-Bonin arc-forearc. They visualized the nascent forearc as consisting of a "series of small rift basins, punctuated and bordered by coherent volcanic edifices."

As discussed below, boninites require extensive melting of depleted mantle, which implies high thermal gradients. Recognition that boninite occurs interlayered with arc tholeiite supports the idea that forearcs are potential sites for the origin of island arcs and are a likely source for at least some ophiolites (Hawkins and Evans, 1983; Hawkins et al., 1984; Stern and Bloomer, 1992; Bloomer et al., 1995).

The geography of forearcs differs; for example, the Tonga forearc is only ~80 km wide, whereas the Mariana forearc is upward of 200 km wide. The early magmatic, tectonic, and sedimentary history of an SSZ system is recorded in forearc sediments and igneous rocks, as well as in successive overprints of magmatic activity. The Tonga and Mariana SSZ systems have evolved by trench rollback and sequential development of arc and backarc systems through Cenozoic time. This history is represented by fragments of pre-existing lithosphere that was rifted in the extensional and magmatic episodes that generated arc and backarc systems. Some of this older forearc crust could be ancient seafloor that formed at a mid-ocean ridge, but in the Tonga and Mariana SSZ systems, the composite crust comprises remnants of older arc and forearc terranes. In both arc systems, parts of the forearc include uplifted platforms covered with Miocene and younger reef limestones.

The Tonga forearc preserves mid- to late-Eocene age fragments of arc crust that are similar to rocks of Viti Levu, Fiji. These may have formed adjacent to either the "ancestral" Tonga Trench (prior to opening of the Lau Basin) or the now inactive Vitiaz Trench. The Mariana forearc includes a frontal arc, comprising islands (Guam, Saipan) with arc volcanic rocks as old as Eocene and similar rock series that are exposed on the outer arc high, near the trench. These probably are fragments of the Palau-Kyushu Ridge that formed in response to subduction in the "ancestral" Mariana Trench, when it lay several hundred kilometers west of its present location (cf. Hawkins and Castillo, 1998).

Guam and Saipan—Insight to Forearc Geology

The broad forearc of the Mariana arc-trench system has been well explored; I will use it as a "type" example. The frontal arc (Guam, Saipan, Rota, Tinian) is separated from an outer arc high that has rocks of similar age and composition. The forearc has a complicated structure that shows evidence of extension in the development of half-grabens. Seismic data show that sedimentary layers are cut by dipping reflectors that are interpreted as traces of normal faults (Mrzowski and Hayes, 1980). Fryer (1992) suggested three possible explanations for the origin of the Mariana forearc. It could have been a continuous broad zone of Eocene volcanism later overprinted by forearc volcanism; the frontal arc and outer high might be separate features; or the forearc terrane might have originated by extension and rifting that dispersed parts of a formerly contiguous Eocene arc. The latter seems to be the most likely explanation.

Mariana forearc crust is well exposed on Guam and Saipan. The oldest rocks on Guam are the late-middle Eocene Facpi Formation comprising "largely interbedded boninite series pillow lavas, pillow breccias and dikes"; these are capped with arc-tholeiitic lavas. The Facpi Formation is overlain by "interbedded volcanic breccia, tuffaceous sandstone, lava flows and sills" of the late Eocene to Oligocene Alutom Formation (Reagan and Meijer, 1984). These range in composition from boninitic to arc tholeiitic and calc-alkaline. Miocene limestones and minor flows and pyroclastics overlie the Alutom Formation. Saipan also has a volcanic core overlain by platform carbonates (Cloud et al., 1958). The oldest rocks are probable late Eocene dacitic flows and pyroclastic rocks. These are overlain by andesitic flows and pyroclastic rock, which in turn are overlain by late Eocene limestone. Probable Oligocene andesitic flows are interbedded with fossiliferous carbonates. Miocene and Pliocene(?), Pleistocene, and Holocene bioclastic, foraminiferal, and coral-algal limestones, with occasional interbeds of tuffaceous limestone and calcareous tuffs complete the section.

For the student of ophiolites, the important geologic factors are the generation of boninite, arc volcanic rocks, and platform carbonates all in a geologic setting in close proximity to MORB-like crust of the backarc basin, all having formed in a geologically short time span (e.g., late Eocene to early Miocene or ca. 15–20 Ma).

Forearc Sediments

DSDP and ODP drill cores from the forearc complement exposures on Guam and Saipan. The drill hole data show that the sediments are largely calcareous ooze with varying amounts of vitric ash and volcanic debris ranging in age from middle Eocene to Recent. The seismic data that show half grabens, together with evidence from DSDP Leg 60, Sites 459 B, and 460 in the forearc (Latouche et al., 1981), support an interpretation of episodic periods of extension that were accompanied by arc volcanism that shed sediment having varying amounts of vitric ash into developing rift basins. From youngest to oldest these are: middle to upper Eocene reworked volcanic detritus, representing early volcanism on the Palau-Kyushu Ridge; upper Oligocene through lower Miocene, representing volcanism on the West Mariana Ridge; and middle Pliocene to present vitric ash from the modern Mariana Arc. Between early Oligocene and early Miocene time, there is a marked diminution of volcanic debris, suggesting a hiatus or minimum in volcanism. The forearc has moved continually trenchward, carrying the rift basins that serve as depocenters for arc detritus, tracking seaward rollback of the trench axis (cf. Elsasser, 1971; Hamilton, 2002). Note that the feature we call a backarc basin, because of its relative position, is initiated in the forearc of an active arc that eventually becomes extinct. There is a lag time of a few million years before the new arc gets established.

Forearc Seamounts

The Mariana forearc has more than 50 seamounts, which rise as much as 2.5 km above the seaflooor and range from 10 to 30 km in diameter; some are irregular in form, but many are conical. The conical features are serpentinite mud volcanoes (diapirs); others are uplifted blocks of serpentinized peridotite. Together they form an array aligned parallel to the trench-slope break and to the Mariana Trench axis and set back from the slope break ~50 to 120 km (Fryer, 1996). They have been sampled by dredging (Bloomer and Hawkins, 1983), and some were drilled on ODP Leg 125 (Fryer et al., 1992). The mud volcanoes formed from unconsolidated slurries of serpentinite mud and rock fragments. The latter gives insight to forearc basement, the SSZ-MW, and low-grade metamorphism under the forearc. The entrained rocks are mostly serpentinized harzburgite, lesser amounts of serpentinized dunite, and minor amounts of meta-gabbro, meta-basalt, and metasediments. The harzburgite is highly depleted, probably representing the residue from (multi-stage?) >20% melting. The harzburgite has only traces of TiO_2, and both CaO and Al_2O_3 are <0.6%. Olivine and orthopyroxene have Mg# = 92, and there are only traces of diopside. Chrome spinels have Cr# up to 83 (Ishii et al., 1992); these authors concluded that the peridotites are more depleted than abyssal peridiotites from mid-ocean ridges. Metamorphic assemblages are typical of prehnite-pumpellyite or greenschist facies. Particularly notable is the occurrence of blueschist facies minerals in pebbles of aphyric basalt entrained in the mud. The diagnostic minerals are lawsonite, aragonite, blue amphibole (winchite), and Na-pyroxene (chloromelanite with 30%–35% jadeite) and aegerine-augite (with <2% jadeite) (Maekawa et al., 1992). On the basis of thermal modeling and presence of serpentinite debris in Eocene sediments, Fryer (1996) proposed that extensive forearc metamorphism probably occurred within the first several hundred thousand years of initiation of subduction, and continues today.

Forearc Magmas

The main development of forearc magmatism may occur at the onset of subduction to form nascent arc systems (Hawkins et al., 1984; Stern and Bloomer, 1992; Fryer, 1996). Forearc eruptions of "middle and late Eocene supra-subduction zone magmatism formed a vast terrane of boninites and arc tholeiites that is unlike active arc systems but is similar to many ophiolites" (B. Taylor, 1992). In addition to this major eruption of magmas at the onset of subduction, later basaltic flows and intrusions are found in forearc sedimentary units.

Arc-basalt intrusives. There is evidence from ODP drill sites on the Bonin, Mariana, and Tonga forearcs that arc composition magmas have intruded volcaniclastic sediments; "these intrusives appear to be a common feature of Western Pacific intra-oceanic forearcs" (R. Taylor et al., 1994). There is strong petrologic evidence indicating that these are derived from the volcanic arc rather than forearc mantle (e.g., at ODP Site 841 in the Tonga forearc we recovered basaltic dikes and sills intrusive into lower and middle Miocene sediments, Table 3). Compositions ranged from basalt to andesitic (50–54% SiO_2); they are enriched in K, Rb, Ba, and Sr, but depleted in HFSE, relative to MORB. They resemble Lau Ridge volcanic rocks and show no geochemical evidence to suggest that they were derived by melting of the highly depleted mantle inferred to underlie the forearc. The dikes were likely propagated from the Lau Ridge (Bloomer et al., 1994), which required transport under forearc crust for ~100 km. Marlowe et al. (1992) described similar sills in the Mariana forearc for which R. Taylor et al. (1995) concluded that they are "sills… most likely to have been fed by dikes which propagated from the arc though basement rocks at deeper crustal levels. They do not resemble backarc basalts…." ODP drilling in the Bonin forearc encountered similar intrusives (cf. B. Taylor, 1992). If these intrusive bodies are derived from the volcanic arc, the long transit of relatively unfractionated melts under the "cold" forearc poses problems in view of the thermal setting.

Boninites. Boninites are one of the most distinctive rock types found in the SSZ environment. They never have been found on or near mid-ocean ridges or intra-plate settings, but they are commonly found in close geological association with many ophiolites and arc series rocks. The close geological association of boninites with ophiolites offers unequivocal evidence that the ophiolite had an SSZ origin. Boninites are found in some "classic" ophiolite examples, including Troodos, Pindos, California Coast Range, and Zambales (cf. Pearce et al., 1984; Thy and Xenophontos, 1991; Jones and Robertson, 1991; Shervais and Kimbrough, 1985; Hawkins and Evans, 1983).

Boninites are olivine - chrome spinel - pyroxene rocks in which plagioclase is conspicuously absent. Clinoenstatite is highly distinctive; both clinopyroxene and Mg-rich orthopyroxene are common. Commonly, boninites are vitrophyric with the

TABLE 3. MARIANA FOREARC BONINITES, TONGA FOREARC INTRUSIVES

Sample:	MARA 50-14	MARA 28-18	MARA 28-1	MARA 28-9	MARA 28-43	MARA 51-8	MARA 27-3	ODP 841 Unit 1 Ave
				Wt%				
SiO_2	53.60	57.12	59.46	61.90	67.45	55.22	55.33	51.52
TiO_2	0.18	0.18	0.16	0.17	0.19	0.40	0.74	1.05
Al_2O_3	10.77	10.18	11.16	11.66	13.20	14.45	16.31	16.33
FeO*	8.07	7.70	7.13	5.97	6.27	6.11	9.68	11.40
MnO	0.15	0.16	0.13	0.11	0.09	0.11	0.16	0.24
MgO	15.53	15.25	11.26	8.92	3.55	7.76	4.26	4.67
CaO	6.98	4.36	4.97	4.13	3.97	8.44	8.56	9.59
Na_2O	1.00	1.96	2.19	3.48	3.78	2.60	2.25	3.09
K_2O	0.66	0.98	0.64	0.70	1.37	0.76	0.35	0.28
P_2O_5	0.05	0.02	0.04	0.05	0.06	0.09	0.12	0.16
H_2O	2.60		2.92	1.90	1.40	3.26	2.49	
Total	99.59	97.91	100.06	98.99	101.33	99.20	100.25	98.33
Type:	OL LowCa 3	OL LowCa 1	B LowCa 1	A	Bd	T	At	At
				Trace elements (ppm)				
Cr	1152	1310	722	477	77	3313	29	20
Ni	384	337	2345	150	24	134	14	20
V	161	119	126	100	85	160	321	376
Zr	32	42	50	63	76	60	64	67
Y	7	6	5	3	7	15	30	25
Nb	4	4	6			5	4	1
Rb	10	18	10	10	7	5	7	2.7
Ba	17		36	44	42	17	45	157
Sr	71	93	154	220	283	196	140	234
				Rare earth elements (ppm)				
Ce	2.9		4.01	6.3	7.02	6.24	7.46	
Nd	1.59		2.18	3.22	3.97	4.39	6.6	
Sm	0.46		0.58	0.79	1	1.31	2.28	
Eu	0.15		0.2	0.25	0.3	0.47	0.85	
Gd	0.58		0.71		0.98	1.68	3.26	
Dy	0.68		0.69	0.73	0.86	1.75	3.33	
Er	0.49		0.46	0.48	0.48	1.21	2.6	
Yb				0.45		1.24	2.63	

Note: Blank—no data; A—Boninitic andesite; At—Arc tholeiite; B—Boninite; Bd—Boninitic dacite; OL—Olivine boninite; T—Transitional; LowCa 3—Low Ca Type 3; LowCa 1—Low Ca Type 1.

glass having a broadly andesitic to dacitic composition (Meijer, 1980; Bloomer and Hawkins, 1987). Holocrystalline varieties are also found. The chemistry of boninites is distinctive—rocks considered boninitic "have >53% SiO_2 and Mg# >60, or are demonstrably derived from parental magmas meeting these compositional requirements" (Crawford et al., 1989). R. Taylor et al. (1994) expanded on this by adding that TiO_2 is <0.6%, 7% < MgO < 25%, and orthopyroxene forms phenocrysts and microphenocrysts. These authors propose that "presence of clinoenstatite, or absence of plagioclase, is diagnostic of boninites sensu stricto" (R. Taylor et al., 1994). A complete discussion of boninite petrology is beyond the scope of this paper. I will summarize some of their distinctive features and tabulate data for representative samples in Table 3. See *Boninites and related rocks,* Crawford (1989); Hickey-Vargas (1989); Pearce et al. (1992); and R. Taylor et al. (1994) for more thorough presentations.

Boninites were first described by Kikuchi (1890), from their type locality on Chichijma in the Bonin arc. The type boninites are distinctive in having phenocrysts of olivine, chrome spinel, clinoenstatite, and orthopyroxene enclosed in silica-rich glass. Olivine boninite from the Mariana Trench has glass with 59% SiO_2, and dacite boninite glass has 75% SiO_2 (Bloomer and Hawkins, 1987). Other boninite varieties have both low- and high-Ca pyroxene, but plagioclase in all varieties is restricted to the groundmass of more highly evolved melts. The combination of olivine, chrome spinel, and orthopyroxene, as phenocrysts, the general absence of plagioclase, and the high-silica glass matrix are distinctive features that characterize boninites.

Olivine typically ranges from Fo 86 to 92, chrome spinel has Cr# ranging from 68 to 75, and the clinoenstatite is En 90–96 with <5% Wo. Clinopyroxene, when present, usually is diopside. The distinctive mineralogy is matched with distinctive chemistry. The range in SiO_2 for boninites is 53% to >58%, whereas MgO ranges from 14% to 18%. Fe, Al, and Ca are low, compared to N-MORB. The ratio CaO/Al_2O_3 is variable (e.g., 0.6–0.9) because of the low Al_2O_3 content (commonly <12%) and varied CaO. This probably serves to inhibit crystallization of plagioclase. The HFSEs are depleted relative to MORB, the transition metals are high, as would be expected with the high Mg, and bulk rock Cr# ranges from 70 to 98. The LILE are variable but enriched relative to MORB. The REE are greatly depleted relative to MORB, and chondrite normalized patterns have either negative slope or a U-shaped pattern. Chondrite normalized REE patterns for Bonin and Mariana Trench samples, and for a Zambales Range boninite discussed in a separate section, are shown in Figure 9. Boninites have variable enrichment in volatiles, alkali metals, and alkaline earth trace elements, as well as in water. Using combined deuterium and oxygen isotope ratios to discriminate between primary and secondary water, Dobson and O'Neill (1987) determined that primary boninite and bronzite andesite melts from Chichi-jima had from 1.6 to 2.4 wt % water. This water is attributed to dehydration fluids derived from subducted oceanic crust.

Boninites may be divided into low-Ca and high-Ca classes; low-Ca boninites are further subdivided into three types on the basis of their SiO_2, CaO, FeO, alkalies, and the CaO/Al_2O_3 ratio. High-Ca boninites do not develop clinoenstatite. Table 3 gives data for Type 1 and Type 3 low-Ca boninites from the Mariana Trench and for a rock transitional between boninite and arc tholeiite. Low-*P* crystal fractionation forms a boninite series (Meijer, 1980; Bloomer and Hawkins, 1987) comprising olivine boninite, boninite, boninitic bronzite andesite, and boninitic hypersthene dacite. Some important criteria for boninites are high Cr and Ni, which is consistent with high Mg, very low HFSE, especially in relation to their high SiO_2 and low Al_2O_3 contents. These relationships are retained through the boninite series, and distinguish it from calc-alkaline series rocks. The boninites of the Hunter fracture zone region (Sigurdsson et al., 1993) are layered with low-K rhyolites having high Na and Si. Similar low-K rhyolites of the late Eocene Sankakuyama Formation on Saipan comprise the oldest arc-related volcanic rocks in the Marianas. Meijer (1983) attributed these to extreme crystal fractionation of high-Mg andesites of the boninite series.

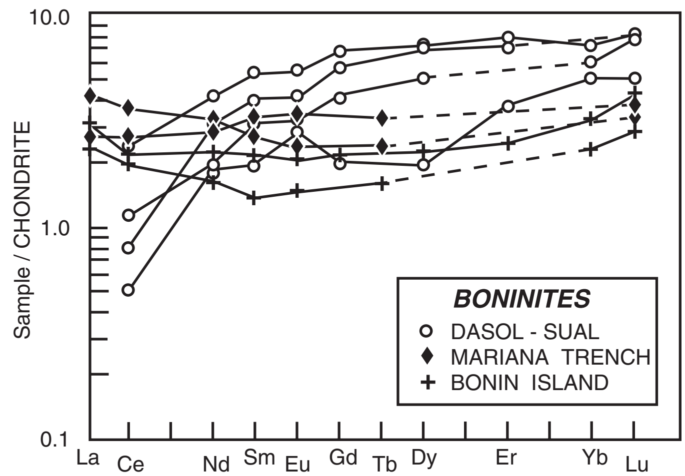

Figure 9. Chondrite normalized REE for boninites of Dasol–Sual series of Zambales Range Luzon (Hawkins and Florendo, 1997, personal commun); Mariana Trench (Bloomer and Hawkins, 1983); and type boninite (Crawford, 1989).

Meijer pointed out their chemical resemblance to some trondhjemites, plagiogranites, and quartz keratophyres that are common constituents of ophiolite assemblages.

Meta-boninites might not be readily recognized as such but would have abundant serpentine, talc, or tremolite, with relict high Cr# spinel; anorthite substitute minerals would likely be rare owing to low Al_2O_3 content of boninites.

The distinctive mineralogy and chemistry (Table 3 and Fig. 9) point to an origin in highly depleted mantle harzburgite, e.g., mantle from which MORB or IAT basalt has been extracted. Note the distinctive field for boninites in the Cr–Y diagram (Fig. 7), where high Cr indicates extensive melting and low Y indicates a depleted source. This melt loss could have been at a mid-ocean ridge or in a backarc basin. Sr and Nd isotope data point to a source with high $^{143}Nd/^{144}Nd$ and low $^{87}Sr/^{86}Sr$, i.e., like MORB-source mantle. This depleted source has been re-enriched in LILE, LREE, Zr, Hf, water, Si, and Na, presumably by subduction fluids. The last three constituents probably help to promote melting of this refractory source material.

The origin of boninites is still not completely resolved, although there seems to be a consensus on several points. A depleted peridotite source such as harzburgite or dunite is required in order to explain the low HFSE and REE and the low Ca and Al. Moderately extensive melting (e.g., up to 25%, Pearce et al., 1992) is required to derive the high Mg and transition metals, and the source must have been enriched in aqueous fluids carrying LILE, LREE, Si, and alkalies in order to facilitate melting and to give the LILE signatures. Addition of Zr may be required to explain high Zr/Ti and Zr/Sm. Melting must have been at moderately low-*P* but high-*T* (e.g., 1250° C; Pearce et al., 1992). The unusual conditions required are most likely to have developed at convergent margins that are not in thermal equilibrium, such as during the initial stages of subduction (e.g., Hickey and Frey, 1982; Bloomer and Hawkins, 1983; Kobayashi, 1983; Crawford et al., 1989; Hawkins et al., 1984; Stern et al., 1991). Influx of high-*T* mantle is essential to provide the heat; thus, conventional models involving cold/cold subduction as modeled for most subduction zones (Minear and Toksoz, 1970; Oxburgh and Turcotte, 1970; Hasebe et al., 1970) are inadequate. Tomographic imagery for the mantle wedge indicates that "slow" velocities are under the arcs, whereas velocity structures are too "fast" near the subduction zone to indicate high temperatures (van der Hilst, 1995). Thus, subduction of high-temperature lithosphere or mantle upwelling, rather than "cold" subduction, is required. Pearce et al. (1992) summarizes several of the more plausible geologic settings for boninite genesis. They include: "partial melting of serpentinized harzburgite during initiation of subduction on a former transform fault," as suggested by Hawkins et al. (1984); "partial melting owing to subduction of young hot crust" (e.g., subduction of a ridge lying parallel to the subduction zone), as suggested by Tatsumi (1982) and Crawford et al. (1989); and "contact melting of hydrous mantle wedge by injection of a hot mantle diapir during arc rifting," as suggested by Crawford et al. (1989). Stern and Bloomer (1992) proposed upwelling of hot mantle into a rift formed by transtension at the initiation of a subduction zone. Yet another possibility is propagation of a ridge into the forearc from the *backarc* regions, as suggested by Hawkins and Castillo (1998). An example of the latter is found on the Hunter fracture zone where a propagating ridge from the North Fiji Basin encounters depleted mantle above an incipient oblique subduction zone (Sigurdsson et al., 1993). Boninite is layered with primitive high-Mg arc tholeiite and low-K rhyolite. The Hunter fracture zone setting may be a unique place, where the earliest stage of subduction zone initiation is found.

The forearc of the north end of the Lau Basin has a similar, but less well understood, occurrence of boninites (Falloon et al., 1989). The authors concluded that these lavas were formed as primary melts of upper-mantle peridotite. The boninite was collected from the upper slopes of the westward trending part of the Tonga Trench. One limb of the active Mangatolu Triple Junction in the Lau (backarc) Basin penetrates into the forearc near these boninites; this may have been important as a heat source. Depleted peridotite of the mantle wedge was the melt source.

The various models for boninite genesis involve extensive low-*P* melting (input of heat) into depleted mantle of the forearc region, this requires upwelling of hot mantle into rifted oceanic crust. Boninites are found in forearcs. Apparently they are in situ; it is likely that this is where they formed. The common occurrence of arc tholeiite interlayered with boninite argues that they represent the nascent stages of arc genesis.

GEOLOGY OF TRENCHES

The 10- to 10.9-km deep Tonga and Mariana Trenches are far removed from major continental drainage systems or heavily glaciated regions and, thus, receive an insignificant amount of sediment, except for the thin layer of siliceous and carbonate pelagic material and the volcaniclastic material from within-plate volcanic edifices, that forms oceanic seismic layer 1. A small amount of airborne and suspended material from the continents and nearby volcanic arcs, as well as a very small amount of meteoritic material, is also added. As a consequence, there is no accreted wedge of sediments, such as is found in the Sunda, eastern Aleutian, and Middle America trenches. The Tonga and Mariana trenches resemble other Western Pacific Basin trenches, such as those of Yap, Palau, Philippine, New Hebrides and northern Kermadec. Rock dredges collected over short depth increments have given us a qualitative understanding of rock types and their probable layering on trench walls. Intuitively, we might expect to find the classic ophiolite sequence neatly displayed in cross-section. Such is not the case; although all parts of the ophiolite sequence are found, they have been shuffled by deformation (Fisher and Engel, 1969; Dietrich et al., 1978; Sharaskin et al., 1980; Bloomer and Hawkins, 1983; Bloomer, 1983; Bloomer and Fisher, 1987; Hawkins and Castillo, 1998). In both the Tonga and Mariana Trenches, and indeed in all Western Pacific Basin trenches, samples from the inner trench wall include large amounts of highly depleted, variably serpentinized peridotite, presumably representing sub-arc mantle. We have the best idea of the range in rock types and their relative "stratigraphy" from the Mariana Trench (Bloomer and Hawkins, 1983; Bloomer, 1983), and we use it as our model for an inner trench wall, as in Figure 10, which is based on dredge sampling at ~18° N. Note that Figure 10 is a composite sketch based on several dredge collections over short depth increment. The inferred "petrologic layering" is based on the first appearance with depth of a rock type (rocks do not roll upslope) or the unique composition of a single dredge increment. The same rock types and similar relative "layering" are found in the Tonga Trench (e.g., Fisher and Engel, 1969; Bloomer and Fisher, 1987; Hawkins et al., 2000). It should be emphasized that we have no data for the nature of contacts between rock types, and nothing to indicate that an "ophiolite" layering or any other kind of primary layering is preserved. In fact, the abundance of serpentinized peridotite found at all depths on trench walls points to the highly disrupted and interlayered distribution of rock types. The presumed layers are shown dipping inward, but this is an inference based on the likelihood that rocks have been imbricated because of relative overthrusting of the hanging wall of the subduction zone. The diagram has great vertical exaggeration in order to show the layers; true slopes angles are shown on the sketch.

It is highly significant that crustal rock assemblages exposed on the inner trench wall carry the chemical signature of SSZ magmatism and thus are likely to be from the *forearc*. DSDP and ODP drilling of the Mariana forearc support this inference. Minor scraps of MORB-composition seafloor or seamounts have been dredged, but for the most part trench wall rocks *are not* fragments of mid-ocean ridge-derived lithosphere. They do, however, represent all parts of classic ophiolite assemblages.

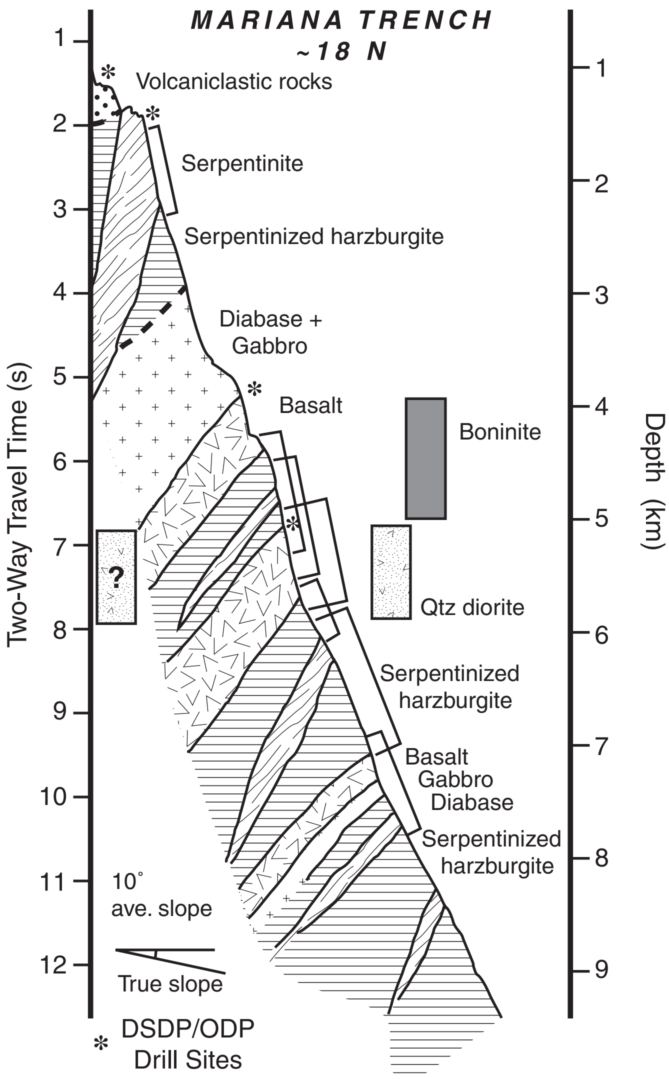

Figure 10. Schematic section based on the author's fieldwork, showing distribution of rock types and their probable depth range of exposure on the inner wall of the Mariana Trench at 18° N (cf. Bloomer and Hawkins, 1983). Boninite and tonalite were collected from depth ranges shown, but their relation to other types is not known. Inward dip of layers is inferred and their lateral extent is not known. Note that the slope of the trench wall has been greatly exaggerated to show the rocks. True slope angle is ~10°.

Our data for the Mariana Trench wall indicates that it mainly exposes ultramafic rocks between depths of 2 and 8.8 km; these are largely serpentinized, with lizardite as the main serpentine mineral and with lesser amounts of antigorite. The protoliths, as indicated by relict grains are (in decreasing order of abundance) harzburgite, dunite, and lherzolite. Most are tectonites, but some are cumulates. Many show effects of deformation, such as flattened olivine and kink-bands in olivine, indicating high-*T* deformation. At the southern end of the trench, southwest of Guam, we collected many samples having cataclastic textures; some are recrystallized to talc-tremolite schists.

The following summary of rock types is based on a more extensive discussion given in Bloomer and Hawkins (1983). Harzburgite is the main non-cumulate rock type found, and consists of olivine, Fo 91–92; orthopyroxene with En 90–92 and Wo 0.2–2; and chrome spinel with Cr# 62–84. Lherzolite is rare, and has Fo 87–91; orthopyroxene with En 85–89 and Wo 1–3; clinopyroxene with En 50–52 and Wo 42–45; and chrome spinel with Cr# 43–58. The clinopyroxene is diopside or diopsidic augite. Lherzolite commonly has small patches of plagioclase and clinopyroxene, suggesting either that they have been impregnated with melt or that they retain trapped melt. Cumulate textured ultramafic rocks, orthopyroxenites, and websterites have Fo 76–85; orthopyroxene En 67–85 and Wo 1–2; and clinopyroxene En 43–48 and Wo 45–46. One cumulate textured harzburgite has olivine poikilitically enclosed in orthopyroxene; calcic plagioclase; and primary, intercumulus, brown pargasitic amphibole (Mg# 74).

Massive and cumulate textured gabbros constitute an important part of the Mariana collection and were collected at eight dredge sites. They are most abundant between 3 and 4 km depth and between 7 and 7.5 km depth. The massive gabbros have hypidiomorphic textures, consisting mainly of plagioclase and clinopyroxene; noritic gabbro is less common and comprises plagioclase-clinopyroxene-orthopyroxene-Fe-Ti oxide. Both types grade texturally into micro-gabbro and diabase. Cumulate textured anorthositic norite, clinopyroxene norite, and orthopyroxene gabbro have heteradcumulate texture comprising plagioclase-clinopyroxene-orthopyroxene-Fe-Ti oxide. All of the gabbros have minor amounts of secondary minerals (e.g., uralite, chlorite, talc, serpentine, calcite). Many gabbro samples display metamorphic textures reflecting moderately high-*T* shearing and recrystallization. Some are blasto-mylonitic hornblende gabbro gneisses.

The crystallization sequence of most gabbros was olivine-chrome spinel-orthopyroxene-clinopyroxene-plagioclase; less common was olivine-plagioclase-clinopyroxene. The dominance of orthopyroxene in most samples indicates derivation from silica-saturated melts. Bulk rock chemistry of the cumulate gabbros shows very low concentrations of K_2O, P_2O_5, Rb, Ba, Zr, and generally low TiO_2 (Bloomer and Hawkins, 1983). We interpreted the cumulate gabbros as probable derivatives of boninite series magmas. This is consistent with the interpretation that most of the Mariana Trench collection of mafic and ultramafic rocks formed in the forearc and may have formed contemporary with the boninites.

Tonalite and quartz diorite are rare, but significant rock types collected from the Mariana Trench in the depth range of 5 to 6 km (Bloomer and Hawkins, 1983), as shown schematically on Figure 10. These rock types have been collected from the Tonga Trench wall from depths between 5.9 and 5.2 km (Hawkins et al., 2000). Suyehiro et al. (1996) proposed, on the basis of geophysical data, that tonalite, like the tonalite of the Tanzawa (Japan) arc, may underlie the Izu-Ogasawara ridge. The occurrence of tonalite-quartz diorite on intra-oceanic trench walls is significant in terms of arc evolution, but we have no idea of its abundance or distribution and attach no significance to the similar depth of occurrence in the Mariana and Tonga Trenches.

Rocks representing higher crustal level include variously altered basalt, diabase, andesite, dacite, and *boninite*. The only evidence for accreted material is scraps of alkalic basalt, chert, hyaloclastite, shallow water limestone, and upper Cretaceous radiolarian silicic sediment. It is likely that these were swept from a seamount riding in on the incoming plate. Highest samples on the trench slope are volcaniclastic siltstones containing Miocene discoasters and glass shards. The glass shards have refractive indices that are consistent with 54%–56% SiO_2 (Bloomer and Hawkins, 1983). Lee and Stern (1995) presented a comprehensive study of the 40 m.y. record of tephra glass from DSDP and ODP drill sites for the region from the West Philippine Sea to the Mariana forearc. Their data are mainly for low and medium-K basaltic to rhyolitic compositions, but they noted a pronounced minimum in the 65% to 66% SiO_2 range. Johnson et al. (1991) reported samples of MORB and OIB, together with arc-like material, dredged from the forearc. They interpreted the MORB and OIB as oceanic plate material accreted to the forearc.

INTRA-OCEANIC VOLCANIC ARCS

Introduction

The Plio-Pleistocene to Recent island arc volcanoes erupted on the Tonga-Kermadec and Mariana Ridges can serve as type examples of subduction related intra-oceanic volcanism. Inasmuch as there is a voluminous literature on island arc magmas (cf. Barker, 1979; Perfit et al., 1980; Gill, 1981), I will summarize some of their distinctive features and associations that may be useful in recognizing them as components of SSZ arc-ophiolite terranes. One of the keys is the close geologic association of island arc volcanics with MORB-like crust of backarc basins and depleted rocks of forearcs. A mature arc system, with 20–30 km thickness of plutonic and volcanic rocks, which had evolved to a calc-alkaline series, would not be interpreted as an ophiolite. An objective of this discussion is to show how deep crust remnants of such a complex, or remnants of a nascent arc that was terminated in its earliest stages, could mimic the petrology, geochemistry, and layer thickness much of what we term an ophiolite.

In early stages of arc volcanism (the "infant arc" stage of Stern and Bloomer [1992]), eruption of high-Mg melts—boninite, picrite, and high-Mg basalts—forms pillows, dikes, and plutons. Their chemistry requires high thermal gradients and extensive melting of the mantle sources (Hawkins et al., 1984; Stern and Bloomer, 1992; Eggins, 1993). Depth of melt separation may be <50 km. Boninites form in the forearc concurrently with Mg-rich, low-K arc tholeiite, as shown by their interbedded occurrence in ophiolites and their in situ presence in forearcs. As arcs evolve, they become more enriched in K and other LILE, and depleted in Mg by fractionation. Thus, arcs may evolve from an early boninite plus arc tholeiite stage, to a basalt-dominated arc tholeiitic stage (Hawkins et al., 1984; Stern and Bloomer, 1992), to the "classic" calc-alkaline volcanic series (cf. Gill, 1981). The earliest stages of "arc volcanism" obviously involve eruption of magmas onto the seafloor, either in rift systems (cf. Hawkins and Evans, 1983; Hawkins et al., 1984; Stern and Bloomer, 1992) or as point sources. Several tectonic factors (e.g., a change in relative plate motions from transtension to transpression) may be involved in the early termination of a nascent arc system, such that it never develops a chain of robust emergent edifices that evolve to calc-alkaline series lavas or shed large aprons of volcaniclastic material. Remnants of nascent arc crust would resemble an ophiolite. A strong argument can be made that nascent arc crust is the likely protolith for many ophiolites (cf. Stern and Bloomer, 1992).

Active volcanic island arcs are separated from extinct remnant arcs by backarc basins. It is debatable as to whether or not the *active volcanic arc* was rifted from the remnant arc at the initiation of backarc opening, or whether a wave of ephemeral arc volcanoes marched seaward across the forearc until it was established in a new fixed position. In my opinion, the latter seems to be most likely. The evidence from the Lau Basin demonstrates that the Lau Ridge (now a remnant arc) remained active well into the opening history of the basin. Local intrabasinal arc volcanoes were active as the basin opened; these were less voluminous than the basalt eruptions that formed *new* Lau Basin seafloor and filled rift basins.

Tofua arc volcanic debris first appears on the Tonga Ridge (forearc) at ca. 3 Ma (Clift et al., 1994, 1998), long after rifting of the basin began and at about the same time as last volcanism on the Lau Ridge. If the ancestral Lau Ridge arc had been split, there should be an ~6–7 m.y. record of Tofua Arc volcanism on the forearc. The arc is constructed along a trend parallel to the Tonga Ridge, ~50 km west of its inboard edge and on the outboard edge of the Lau Basin. In most places, it is separated from the Tonga Ridge by small rift basins. The arc may overlie part of the Tonga Ridge crust, but there is no evidence for this; my prejudice is that arc magmas penetrated along deep zones of weakness that correspond to the extensional faults that bound the backarc basin on each side. This observation may be of use interpreting arc-ophiolite assemblages. In the case of the Tofua Arc, the arc edifice and its underlying plutonic section may lie directly on depleted mantle rather than on older crust. Hawkins and Evans (1983) reached a similar conclusion for the island arc part (Acoje block) of the Zambales ophiolite.

The Tonga Ridge comprises basement formed of older arc volcanic, plutonic, and arc-derived clastic rocks that are overlain by six seismic sequences inferred to range in age from late Eocene to late Miocene (Austin et al., 1989). These comprise volcaniclastic turbidites and hemipelagic carbonates. They are overlain by post–Late Miocene marls, reefal deposits, and shelf limestones. Between Vava'u and 'Eua are uplifted and westward-tilted blocks of forearc volcaniclastics, marls, and carbonates capped by shoals, reefs, and coralline islands. The oldest known rocks of the Tonga Ridge rocks are mid-to late-Eocene (46–40 Ma) arc volcanic and plutonic material exposed on ‘Eua Island (Ewart and Bryan, 1972). Gabbro dredged at the north end of the Tonga Ridge gave an Ar/Ar radiometric age of 50±9 Ma, as cited in Clift et al. (1998). ODP Site 841, drilled on the forearc near 23° S, recovered low-K rhyolite, rhyolite tuff, breccias, welded tuffs, and lapilli tuffs of a rhyolite caldera. The pyroclastic rocks are interbedded with mid- to late-Eocene reef material that includes the foraminifer *Discocyclina.* A K-Ar age determination for rhyolite glass gave an age of 44±2 Ma (McDougall, 1994). The forearc igneous basement probably was rifted from the late Eocene arc series of Viti Levu, Fiji Islands (cf. Whelan et al., 1985; Gill, 1970, 1987). This igneous basement also includes Oligocene lavas (33–31 Ma) and early Miocene dikes (19–17 Ma) that probably represent Lau Ridge volcanism (Hawkins and Falvey, 1985; Duncan et al., 1985). The Oligocene and Miocene arc magmas intrude, or are emplaced on, late-Eocene arc material of the Tonga Ridge.

The present Mariana Arc probably formed after the ca. 10 Ma initiation of rifting and the subsequent episode of spreading that formed Mariana Trough crust (e.g., by mid-Pliocene; Latouche et al., 1981; Fryer, 1996). The Mariana Arc is constructed in part on thinned forearc crust (e.g., like crust exposed on Guam and Saipan), and in part on backarc basin crust (Stern et al., 1989). The Mariana Trough has cross-chains of arc volcanism that persisted during basin rifting (Fryer, 1996), and at least one arc composition volcano sits on “zero age” crust of the axial ridge (Table 2, this paper; Hawkins et al., 1990). The data for both the Mariana and Tonga systems suggest that as trenches and their hinge lines retreat seaward because of rollback, a continual igneous carpet of arc, backarc, and forearc magmas is laid down. This is seen in the fabric of seafloor features of the northwestern Pacific basin (Fig. 4).

Island Arc Petrology

The petrologic peculiarities of boninites have been discussed and an argument made that their most likely site of origin is in the forearc. In the earliest stages of arc evolution, as seen on Guam and on the Mariana forearc, we find boninite interlayered with arc tholeiite. An important implication of this association is that eruption of high-Mg island arc lavas may begin anew in the forearc and, as discussed above, that in the course of SSZ evolution, “new” arcs need not have split off from an ancestral “remnant”arc.

*Origin of High-*Al *Basalts*

Occurrences of low-Ti, high-Mg basalts and picrites (e.g., MgO >8%, Mg# >70, Ni >200 ppm) are rare in island arc systems, but important for understanding the petrogenesis of arc magmas. An explanation for this rarity is that high-Mg magmas are relatively dense (e.g., >2.7 gm cm^{-3}); this traps them under less dense crust (e.g., 2.6–2.65 gm cm^{-3}), which effectively acts as a filter (Stolper and Walker, 1980; Gust and Perfit, 1987). Stolper and Walker (1980) showed that there is a liquid density minimum in fractionation of picritic melts. Melt density is controlled by molar ratio (Fe/(Fe + Mg)); minimum melt density is achieved at a molar ratio of ~0.3 to 0.4. Fractional crystallization of olivine obviously decreases Mg, whereas it increases Si, Al, Ca, and Fe. Plagioclase and clinopyroxene form next as liquidus phase, but elevated water content (as envisioned for arc magma systems) would suppress plagioclase crystallization, causing further increase in Al and Ca. Thus, high-Mg basalts could be parental magmas for low-K, low-Ti, high-Al basalts common in many immature island arcs. The probable parental role for picrite can be demonstrated by fractionation models using high-Mg olivine and clinopyroxene phenocrysts, and putative derivative melts. For example, Kay and Kay (1985) modeled the evolution of high-Al arc tholeiite of Adak Island, Aleutian Arc from primitive tholeiitic melts, using the mineralogy of olivine clinopyroxenite xenoliths as the presumed cumulate assemblage. One can visualize a sequence of parental picritic magmas being stripped of high-Mg olivine to form high-Al basalt, followed by crystallization of both olivine and clinopyroxene; this reduces Mg and the Ca/Al ratio, and leads to basaltic andesite compositions. The ultramafic cumulates would comprise dunite, harzburgite, wehrlite, websterite, and clinopyroxenite. As discussed below, these are common in cumulate horizons of what are interpreted as “roots” of island arcs.

Arc picrites are known from the Solomon Islands (Ramsay et al., 1984); Aleutians (Gust and Perfit, 1987); and Vanuatu (Eggins, 1993). Seamount RN-47 from the northern Tofua Arc (Fig. 5 and Table 4) also has high-Mg chemistry. Petrographic evidence and melting experiments demonstrate that the liquidus phases for these high-Mg lavas were olivine and spinel, closely followed by clinopyroxene. Crystal cumulates would form wehrlite and olivine clinopyroxenite, or websterite in melts more saturated in Si. Complementary fractionated melts would be high-Al basalt. These ultramafic rock types form layered cumulates of some ophiolite terranes. Some examples are the Acoje block of the Zambales ophiolite, Luzon (Hawkins and Evans, 1983), and the Tonsina-Nelchina segment of the Talkeetna arc, Alaska (DeBari and Coleman, 1989; DeBari and Sleep, 1991). The Talkeetna cumulates form thin layers of wehrlite, websterite, and olivine clinopyroxenite beneath a 5- to 6-km-thick series of layered two-pyroxene gabbro and gabbro-norite. DeBari and her colleagues interpreted the complex as the remains of a deep (30–35 km) magma chamber that was the roots of a mature Jurassic island arc. Upper levels of the complex include plutonic rocks ranging from gabbro to tonalite; these are overlain by basaltic andesite and silicic ignimbrite.

Early in their histories the Tofua and Mariana arcs erupted low-K island arc tholeiite series lavas, primarily basaltic and basaltic andesite in composition (e.g., Ewart and Bryan, 1972; Woodhead, 1988; Table 4). Andesite and silica-rich varieties become more abundant as arcs mature and evolve into the K-enriched calc-alkaline series (cf. Gill, 1981). Many Mariana arc

TABLE 4. TOFUA AND MARIANA ARC

Sample:	RNDB 47-19	TAFAH1 116	LATE L-20	LATE L -5	LATE L-13	TOFUA N32	ATA 8-12	ASCUNCION AS-3	AGRIGAN 9-1A	AGRIGAN 7-1	SARIGAN SA-52	SARIGAN SA-53	S. HYOSHI ave	N. HYOSHI ave
								Wt%						
SiO_2	49.12	52.52	53.45	53.86	57.57	56.91	51.90	53.09	48.70	60.10	52.59	57.06	50.68	45.79
TiO_2	0.85	0.36	0.51	0.58	0.80	0.77	0.66	0.84	0.49	0.81	0.74	0.89	0.98	1.01
Al_2O_3	16.39	16.85	15.86	16.95	14.14	14.77	16.30	17.88	20.00	16.00	16.50	17.74	18.74	18.52
FeO*	8.48	9.68	9.83	9.62	11.86	10.94	9.72	9.42	8.31	8.32	9.22	7.48	8.97	10.82
MnO	0.19	0.18	0.19	0.16	0.21	0.19	0.19	0.17	0.16	0.22	0.18	0.20	0.18	0.18
MgO	7.42	6.04	6.06	4.85	3.29	3.81	6.72	3.63	6.19	1.87	6.76	3.87	3.33	6.43
CaO	13.12	11.95	11.38	11.23	8.42	8.85	12.10	10.01	13.32	5.81	10.75	8.68	8.79	11.56
Na_2O	2.17	0.95	1.65	1.74	2.42	2.22	1.83	2.75	1.57	4.25	3.13	3.11	3.48	2.32
K_2O	0.53	0.15	0.38	0.49	0.70	0.58	0.41	0.54	0.37	2.06	0.53	0.89	2.07	1.40
P_2O_5	0.21	0.04	0.08	0.07	0.12	0.11	0.12	0.15	0.12	0.35		0.15	0.25	0.29
Total	98.48	98.72	99.39	99.55	99.53	99.15	99.95	98.48	99.23	99.79	100.4	100.07	97.47	98.32
								Trace elements (ppm)						
Cr	544	62	78	55	6	15	89	13	28	22			11	49
Ni	154	23	26	25	9	10	28		26	13	33		17	30
Co	50	41	33	29	31	36	39	30	28					
V		315	295	320	410	392	550						260	398
Sc	36	45	40	40	41	41	48						23	35
Zr	41	8	25	24	38	43	40		45		64		88	31
Y		6	15	15	25	20	15				23		25	21
Nb	<1	0.47				0.4								
Hf	1.33	0.38				1.65	0.859	1.7						
Ta	<0.1	0.03				0.04		1.7						
Rb	14	2	6	6	9	8	8	8.58	8	39	10		48	26
Ba	34	47.9	96	125	165	173	133	178	86	333	271		672	714
Sr	120	108	215	230	225	207	214	293	305	314	350		674	1134
B			10	10	18									
								Rare earth elements (ppm)						
La	3.95	1.05	2.7	1.4	3	3.1	2.91	4.65						
Ce	8.46	2.28	6.6	3.4	7.6	8.48	7.74	10.94						
Nd	6	1.98				8.15	5.42							
Sm	2.35	0.75				2.73	1.78	2.77						
Eu	0.825	0.31				0.95	0.692	0.09						
Gd		1.12				3.47	2.33							
Tb	0.387	0.22				0.62	0.43	0.65						
Yb	1.85	1.03				2.45	1.68	2.66						
Lu	0.242	0.17				0.4	0.263	0.39						
Source:	JWH	EH	EB	EB	EB	EB	V	DB	S	S	MR	MR	B	B

Note: B—Bloomer et al., 1989; DB—Dixon and Batiza, 1979; EB—Ewart et al. 1973; EH—Ewart and Hawkesworth, 1987; JWH—J. Hawkins, this paper; MR—Meijer and Reagan,1981; S—Stern, 1979; V—Vallier et al., 1985; blank—no data.

volcanoes are arc tholeiitic, but some have evolved to medium-K tholeiitic and low-K calc-alkaline lavas that include andesite (e.g., Asuncion, Agrigan, and Sarigan, Stern, 1979; Meijer and Reagan, 1981; Table 4). The northernmost volcanoes of the Mariana Arc, Northern Seamount Province, are unusual in that they are LILE-enriched shoshonites (Bloomer et al. 1989; Lin et al., 1989). Quaternary air-fall tephra collected from the backarc basin record melts that are more silicic than are found as flows in the arc volcanoes. Straub (1995) described a range in tephra chemistry from basaltic to dacitic (48%–71% SiO_2; 0.8%–3.2% K_2O). These silica-rich sediments could serve as protoliths for "cherty argilites" common in many arc-ophiolite assemblages.

Tofua and Mariana Arcs

Many Tofua arc volcanoes are basalt or basaltic-andesite (e.g., Tafahi, Late, Tofua, Ata; Table 4), yet there is evidence for bimodalism represented by low-K dacite and rhyodacite (Btryan et al., 1972; Ewart et al., 1973; Hawkins, 1985). Tofua arc volcanoes (e.g., Fonualei Island and Metis Shoal, Fig. 5 and Table 5) have erupted low-K rhyolite pumice and low-K, high-Ca dacites, respectively. Bryan (1979) noted that Tongan dacites are among the most calcic in the world. Metis Shoal has phenocrysts of An 84 and hypersthene, and xenocrysts of bronzite and olivine, enclosed in high-silica glass. Fonualei Island dacite has labradorite, hypersthene, and augite phenocrysts. Both lack modal alkali feldspar or quartz, even though they have up to 20% normative quartz. These low-K, high-Ca dacites are not unique to Tonga; they are found in the Mariana and Bonin arcs, as well in as several others.

Zephyr Shoal, a Lau Basin seamount (Fig. 5), has andesitic bulk rock chemistry (Table 5). It has vitrophyric texture consisting of a web of low-K rhyolite glass (73% SiO_2) that encloses abundant phenocrysts of An 65, hypersthene, and augite. Clearly high-silica melts have been important, even in the juvenile stages of arc genesis, and may erupt in backarc basins, as well, as shown by Lau Basin sediments that have glass shards ranging in composition from basaltic to rhyolitic, with up to 76% SiO_2 (Clift and Dixon, 1994).

The plutonic equivalents of low-K dacites and rhyolites are the trondhjemite-tonalite plutons known from several arc systems (Table 5). The mid- to late-Miocene gabbro to tonalite plutons of the Coto plutonic suite of Viti Levu, Fiji are an example (Gill and Stork, 1979). The Miocene Tanzawa gabbro-tonalite plutonic complex may underlie much of the Izu arc (Kawate and Arima, 1998). Suyehiro et al. (1996), using seismic velocity data, proposed that as much as 30% of arc crust might be composed of rocks similar to the Tanzawa tonalite. The depth to the Moho under the Tonga Ridge has not been determined, but Crawford et al. (2003) suggest that it may be 20–25 km. They recognized a layer having relatively low P-wave velocities (6.0–6.5 km sec^{-1}) that begins at 4–5 km below seafloor and is as much as 5 km thick. They proposed that these velocities are consistent with the layer being tonalite, as was interpreted for similar observations on the ~22 km thick Izu-Ogasawara ridge (Suyehiro et al., 1996). Granitoids are rarely exposed in young arcs (e.g., post-Pliocene), but are found in some Aleutian volcanoes, and are more common in those older than late Miocene (cf. Byers, 1961; Vallier et al., 1994). Minor amounts of tonalite and quartz diorite have been dredged from the Tonga Trench inner wall near 15° S (Table 5; Hawkins et al., 2000), as well as from the Mariana Trench near 18° N (Bloomer and Hawkins, 1983). A possible explanation for the origin of the tonalitic-dacitic rocks is fractionation of silicic melts formed by dehydration partial melting of hydrous (amphibolitic) basalt, as suggested by Kawate and Arima (1998).

EXAMPLES OF SSZ OPHIOLITES

The objective of this presentation has been to summarize the geology of SSZ with special emphasis on the petrology and geochemistry of the igneous rocks. A primary goal was to show the broad petrologic similarity of backarc basin crust to ocean crust formed at mid-ocean ridges. There are many similarities, but there also are subtle differences in trace element chemistry that serve to distinguish between them as possible sources of ophiolites. Of these, the HFSE are the most important in that they will survive effects of alteration and low-T metamorphism that may be imposed during terrane accretion and deformation. Another objective was to show that backarc basin crust forms nearby, and concurrently with, other rock units such as boninites, arc volcanic and volcaniclastic rocks, granitoids, as well as pelagic sediments. Thus, there is a rock association—*geology*—that helps to recognize SSZ ophiolites. Using similar lines of reasoning, others have concluded that some ophiolites, including many "classic" examples first thought to be from mid-ocean ridges, probably originated in an SSZ setting. Dewey and Bird (1971) based on their study of Newfoundland ophiolites, were among the first to recognize that "The ophiolite suite is probably generated at oceanic ridges, and by diffuse spreading in marginal basins behind and within island arc complexes."

Table 6 is a partial list of ophiolites that have abundant evidence for their assignment to an SSZ origin. Data for several of these ophiolites are plotted on the Hf/3-Ta-Th diagram (Fig. 11). They form a virtually complete overlay with data for SSZ systems shown in Figure 8. Trace element patterns for different segments of the Zambales Range, Luzon ophiolite closely resemble patterns for forearc, arc, and backarc rocks from the Mariana and Tonga SSZ systems as shown in Figures 9, 12, 13. Estimates of crustal layer thickness for ophiolite sections are compared to oceanic crust in Figures 14 and 15. There are large differences in layer thickness especially for those interpreted as being from an arc ophiolite. In several instances (e.g., Zambales, Luzon and Bay of Islands, Newfoundland), the ophiolite probably is a composite of arc, and backarc crust. For the Zambales, we recognized a probable forearc (boninite) unit as well. Part of the reasoning behind these interpretations was the ophiolite geochemistry, but the primary factors were the *geology* of the ophiolites and their associated rocks.

TABLE 5. SUPRA-SUBDUCTION ZONE SILICA-RICH ROCKS

Sample	FON F-30	FON F-4	METIS ROCK	METIS GL	S.P. LEE DR-4	ZEPHYR SHOAL-rk	ZEPHYR SHOAL-gl	TONGA TR. PPT 2-15	TONGA TR. PPT 2-5	TONGA TR. PPT 2-14	FIJI COLO	FIJI COLO	TANZAWA KU-1	TANZAWA FJ-1
Wt%														
SiO_2	60.33	65.09	63.66	73.6	73.15	63.26	73.16	59.30	67.23	58.57	64.40	72.40	60.72	71.13
TiO_2	0.65	0.56	0.39	0.52	0.44	0.53	0.71	0.83	0.25	0.76	0.51	0.43	0.61	0.30
Al_2O_3	14.6	14.39	12.42	12.29	14.85	14.56	13.54	15.59	16.08	15.64	16.90	14.10	17.45	15.55
FeO*	9.83	7.76	6.44	3.89	3.25	5.33	3.29	7.85	4.59	6.78	5.10	3.10	6.58	2.83
MnO	0.21	0.17	0.12	0.06	0.16	0.09	0.07	0.07	0.06	0.09	0.14	0.07	0.16	0.12
MgO	2.59	1.32	5.10	1.07	0.72	3.42	0.71	4.81	2.01	7.40	2.20	0.90	3.09	0.93
CaO	7.48	5.90	6.97	3.61	3.30	5.16	2.31	7.67	6.55	6.86	6.20	3.20	7.18	3.58
Na_2O	2.69	3.00	2.58	3.17	2.85	4.03	4.40	3.62	2.85	3.80	3.20	4.60	3.54	4.46
K_2O	0.85	1.13	0.90	1.47	0.50	0.98	1.63	0.17	0.37	0.16	0.67	1.00	0.57	0.99
P_2O_5	0.17	0.17	0.07	0.07	0.06	0.05	0.17	0.09	0.02	0.03	0.10	0.10	0.09	0.10
Total	99.40	99.49	98.65	99.75	99.28	97.41	99.99	100.00	100.01	100.09	99.42	99.90	99.99	99.99
Trace elements (ppm)														
Rb	10	15	14	21				5	7	5	8	9	7	16.1
Ba	195	315	360	610				9	124	21	171	160	181	317
Sr	295	305	140	130				104	163	119	349	210	254	247
Zr	40	46	46	68				67	40	48	82	125	87	99
Y	20	25	21	25				27	11	27	22	35	24	17
Hf			1.22					1.65		1.21				
Ta			0.04					0.132		0.159				
Nb			0.44					4	2	3			1.8	.0.9
Cr		5	230	7				12	27	4	7	4	20	9
Ni		8	53	4				22	16	33	<5	<5	7	2
Co	21	13	25	10				26		22	12	5	19	
Sc	34	25	29	12				29	24	31	12	9		
V	175	105	175	130				233	165	312	66	32	163	29
B	12	15	53	105										
Rare earth elements (ppm)														
La	5.50	3.50	3.24					2.04		1.59			4.6	2
Ce	11.00	10.50	7.46					5.63		4.59			13.7	11.6
Nd			5.58					3.7		3.1			8.5	4.1
Sm			2.25					1.95		1.99			2.94	1.43
Eu			0.84					0.692		0.619			0.9	0.45
Tb			0.39					0.444		0.434				
Yb			1.64					2.34		2.26			2.25	0.85
Lu			0.24					0.291		0.332			0.39	0.2
Source:	EBG	EBG	EBG	EBG	JH	JWH	JWH	JWH	JWH	JWH	GS	GS	KA	KA

Note: EBG—Ewart, Bryan, and Gill,1973; JH—J. Hawkins, 1985; JWH—J. Hawkins, this paper; GS—Gill and Stork, 1979; KA—Kawate and Arima, 1998; blank = no data.

TABLE 6. SUPRA-SUBDUCTION ZONE OPHIOLITES

Location	References
Troodos# *	Miyashiro, 1973; Pearce, 1975; Pearce, 1980; Schmincke et al., 1983; Robinson et al, 1983; Pearce et al., 1987; Flower and Levine, 1987
Oman*	Pearce, 1980; Pearce et al., 1981; Alabaster et al., 1982; Pearce et al., 1984
Vourinos*	Noiret et al., 1981; Pearce et al., 1984
Pindos# *	Pearce et al., 1987; Jones and Robertson, 1991; Capedri et al., 1982
Bay of Islands	Coish and Church, 1979; Elthon, 1991; Jenner et al., 1991
Josephine#	Harper,1984; Alexander and Harper, 1992
Trinity	Wallin and Metcalf, 1998; Metcalf et al., 2000
Coast Range#	Menzies et al., 1977; Shervais and Kimbrough, 1985
Zambales# *	Hawkins, 1980; Hawkins and Evans, 1983; Hawkins and Florendo, 1994
Marum	Davies and Smith, 1971

#Ophiolites having boninite as part of the assemblage.
*Ophiolites having island arc tholeiites as part of the assemblage.

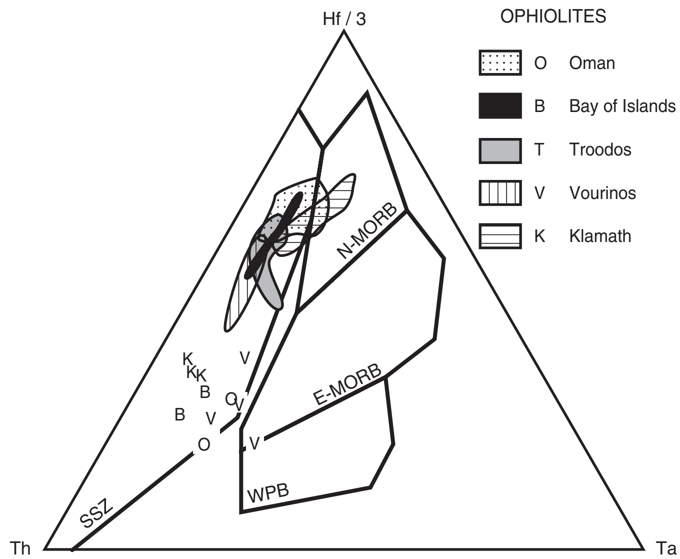

Figure 11. Plot of Hf/3-Ta-Th (after Wood et al., 1981), showing fields for N-MORB, E-MORB, WPB (within plate basalts) and SSZ magmas. Addition of sediment as fluid or melt will shift magma compositions toward Th apex of diagram. Data for ophiolites are from references cited in text. Letters are for isolated datum lying outside fields shown for main groups of data.

OPHIOLITE PROTOLITH—BACKARC OR FOREARC?

In the preceding discussion, I have made the case that forearc, backarc, and nascent arc crust all could be progenitors for the *igneous* part of ophiolites. Is there a way to distinguish between these potential sources? Mature arcs, especially those that have matured beyond the low-K arc tholeiitic stage, are unlikely to be called an ophiolite, although basal cumulates might be confused if not tied to upper level petrology. It is important to recognize that arc assemblages of varied stage in their evolution are to be expected in close association with SSZ ophiolites.

In their nascent stages, at least some arcs are initiated in rift basins (cf. R. Taylor and Nesbitt, 1995). If this nascent stage occurs on crust formed at a mid-ocean ridge (i.e., at the initiation of subduction as described elsewhere in this chapter), boninite, high-Mg basalts, island arc tholeiites, and low-K rhyolites would constitute a distinctive assemblage that is highly unlikely to form in a backarc basin, i.e., these have not been found in backarc basins, to date. The occurrence of this assemblage on the Hunter fracture zone (as described above) is in the forearc of what probably is an incipient subduction zone. This may be an "actualistic" example of a forearc ophiolite in genesis. Guam and the Mariana forearc display examples of parts of a forearc ophiolite, but we do not see them as an integral unit. They are dispersed across the forearc and exposed on walls of the Mariana and Palau trenches. Figure 10 shows all of these constituents.

Forearc ophiolites would differ from those derived from backarc basins in that the latter would resemble mid-ocean ridge crust in terms of petrology and chemistry, as well as in having basalts transitional to arc tholeiite. This compositional range distinguishes them from MORB formed at a true mid-ocean ridge. The lack of boninite and general rarity of high-Mg basalts would separate them from forearc assemblages. The close spatial and temporal association of backarc basin rocks with arc rocks and volcanic detritus would be another important factor. This may not be unique, but a true nascent forearc ophiolite might not have evolved into an emergent volcanic arc. Even if it had, there should be upward continuity without intervening layers of pelagic sediments as found in backarc basins (cf. Hawkins and Evans, 1983).

It is obvious that crustal contraction would tectonically juxtapose arc, backarc, and forearc crust. Rather than hoping to identify a particular ophiolite as having a forearc or backarc origin, it may be more important to recognize it as having an SSZ origin and then attempt to separate out units having petrologic features distinctive for the original setting. I would be surprised to find an SSZ ophiolite associated with an arc terrane that came only from a forearc or a backarc setting, but I would let the geology lead me to the interpretation.

ON THE EMPLACEMENT OF OPHIOLIT1ES

Classic ideas of ophiolite emplacement by obduction have faced the problem of how oceanic lithosphere, which was

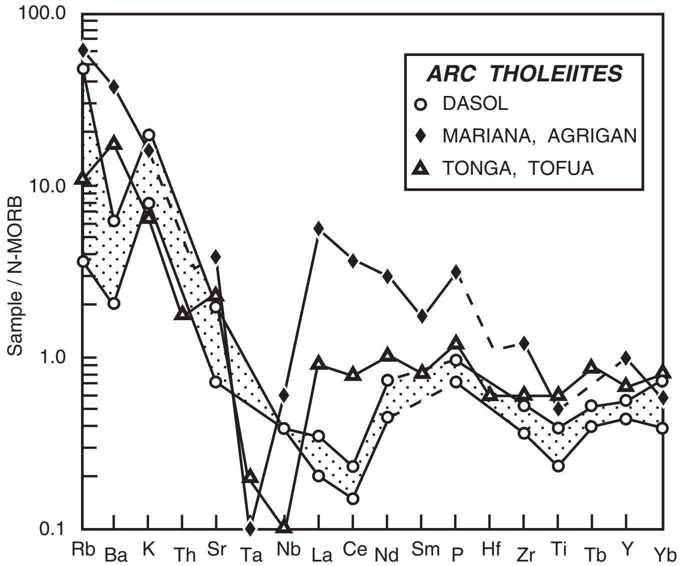

Figure 12. Trace elements for arc tholeiitic rocks from Agrigan Island, Mariana Arc; Tofua Island, Tofua Arc; and Dasol series of Zambales Range, Luzon (Hawkins and Evans, 1983). Data are normalized to N-MORB values (Sun and McDonough, 1989).

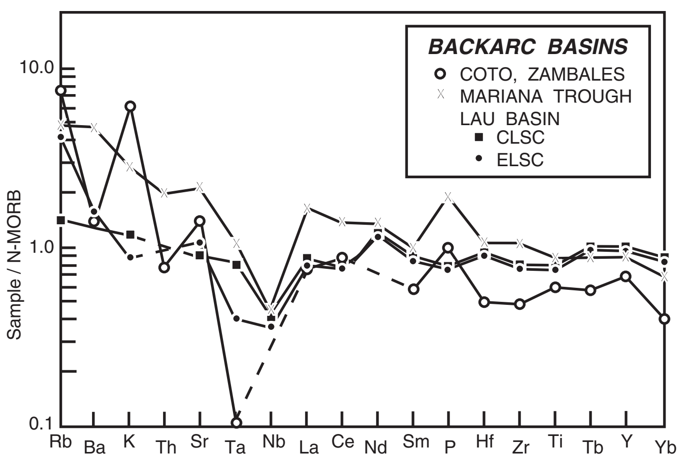

Figure 13. Trace elements for Mariana Trough, Lau backarc basin tholeiitic rocks, and probable backarc basin crust in Zambales Range, Luzon (Hawkins and Evans, 1983). Data are normalized to N-MORB values (Sun and McDonough, 1989).

destined to return to the mantle in a subduction zone, could be diverted and make its way onto continental margins. Coleman (2000) discussed five different mechanisms pertinent to California ophiolites, one of which concerns SSZ ophiolites. He observes that "obduction of ophiolites probably is a rare event when compared to the stranding of oceanic-crust slabs in subduction zones along convergent margins." I agree; this probably could explain the origin of dismembered fragments of ophiolite material, but it seems hard to visualize this process for explaining large, intact ophiolites.

Clearly, some unquestionably oceanic crust is presently being thrust onto other ocean crust (on Macquarie Ridge) in the case of the incipient Puysegur trench; however, many of the characteristic constituents of arc-ophiolite assemblages, such as those in the North American Cordillera, are missing (e.g., boninites and arc plutonic-volcanic series).

In the case of young intra-continental ocean basins such as the Red Sea and Gulf of California, it seems likely that a change in stresses from extension to crustal contraction would lead to emplacement of true ocean crust onto continental crust. Some of the Tethyan ophiolites may have formed this way. These ophiolites would be lacking in the associated arc and forearc rocks common to many ophiolites, but they too would represent ocean crust.

The purpose of this essay is to discuss geologic tracers pointing to an SSZ origin for many ophiolites. The emphasis has been on the *geologic association* of rocks that are formed in this setting and that are common to many ophiolites. If we consider the spatial relations shown in Figures 1 and 2, it should be easy to see how a change from crustal extension to contraction—say, by the blocking of subduction by an oceanic plateau or a microcontinent such as Fiji—could lead to the imbrication of the different rock series. The weakest part of the system would be the axial ridges of the backarc basin. Low-angle thrusts would emplace young, hot, ocean crust over colder ocean crust and set up conditions for inverted metamorphic gradients on the sole of the thrust zone. If the backarc crust were thrust onto a nascent arc, normal, prograde downward metamorphism would prevail. The backarc crust could be thrust onto either older remnant arc crust or younger active arc crust.

There are strong arguments for ophiolites being derived from nascent arcs formed in what might be the forearc to an older arc, or at the inception of subduction (cf. Stern and Bloomer, 1992). Several models for the inception of subduction propose an initial phase of transtension on a fracture zone to set the stage for foundering of older lithosphere. If this were to be followed by transpression before subduction got under way, it could imbricate the young crust and transport it onto thicker crust, e.g., a remnant arc. There are many possible permutations. The key point is that ophiolites have been emplaced onto older crust, many ophiolites carry the SSZ petrologic signature, and their geology indicates a temporal and genetic association with island arc systems. Geometrically, it seems not to be a problem to bring these geologic components together if they are initially arranged as shown in Figures 1 and 2.

SUMMARY

The geology of SSZ systems, as exemplified by Western Pacific Basin backarc-arc-trench systems such as the Lau-Tonga and Mariana zones of plate convergence, exhibits a spectrum of rock types, including boninites (formed in forearcs as initial stages

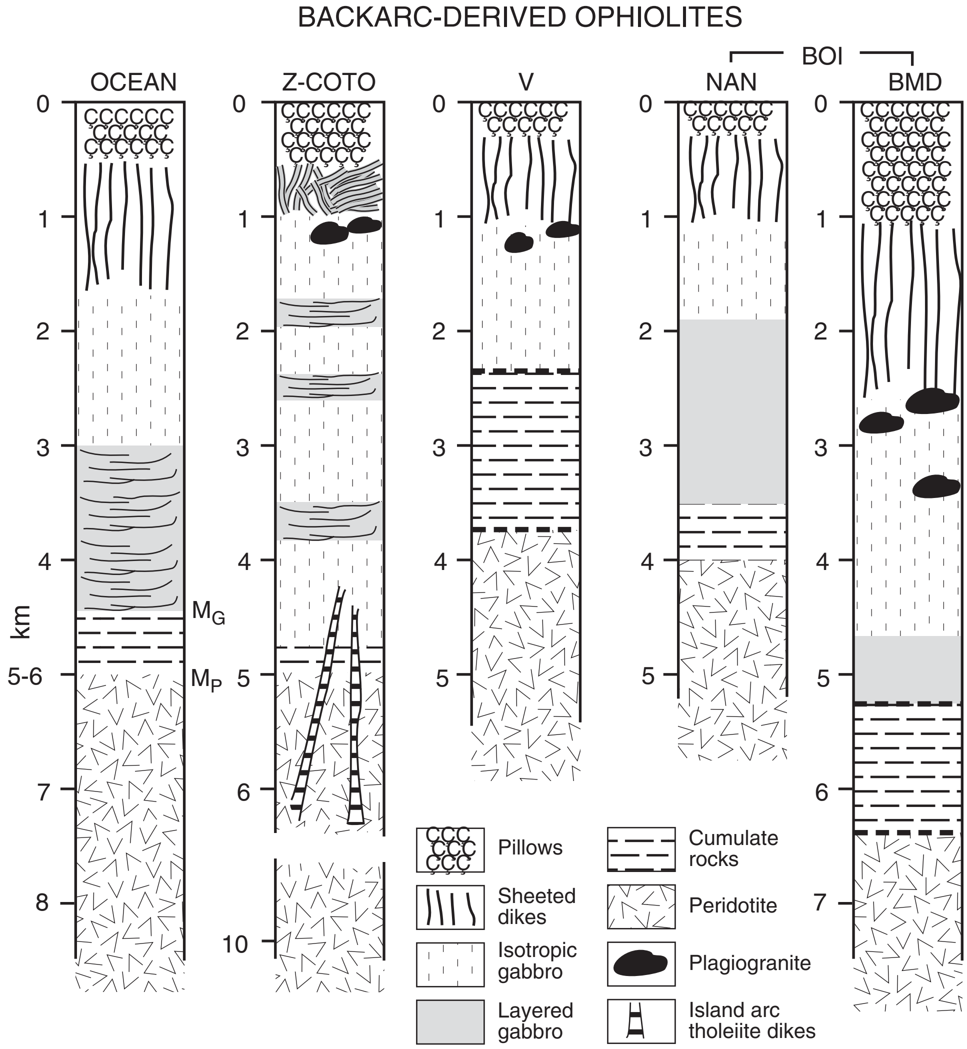

Figure 14. Comparison of oceanic crust with probable backarc basin ophiolite crust for: Z-COTO—Coto, Zambales Range, Luzon; V—Vourinos, Greece; BMD—Blow-me-down Mountain; BOI—Bay of Islands, Newfoundland. References are in Table 6.

of arc volcanism); mafic ocean crust (formed in early stages of arc volcanism and in backarc basins); arc volcanic series ranging from picritic to rhyolitic, and including arc tholeiitic, high-K calc-alkaline; and shoshonitic magma series. Deeper-level arc crust comprises various types of layered and isotropic gabbro, as well as granitoid plutons (e.g., tonalite and quartz diorite). There is a wide range of sediment types, from arc-derived tephra and volcaniclastic turbidites to silicic and carbonate oozes. Crustal units are underlain by mafic cumulates and by mantle rocks comprising depleted harzburgite, dunite, and their serpentinized equivalents. Ultramafic cumulates from beneath island arcs may be dominated by pyroxene and may lack plagioclase. All of the crustal units may be coeval, although some may overlie fragments of an older substrate that comprises similar SSZ material. Arc-derived clastics are deposited on new seafloor of essentially the same age. In the early rift stage of backarc basins, as well as in the forearc, basalt flows, sills, and dikes are interspersed with arc-derived clastic and hemipelagic sediments. Carbonate nannofossil ooze, as well as metalliferous sediments and hydrothermal vent material, are deposited on and among basalt pillows.

The geology of SSZ plate margins illustrates a dynamic system comprising converging plates, mantle upwelling, generation of new oceanic crust and sialic proto-continental crust. On the 100–200 km scale across strike, SSZ systems are regions of lithosphere extension. On a broad regional scale (100s of kms), they are regions of lithosphere contraction; this leads to their eventual shortening and incorporation into tectonic belts at a continental, micro-continental, or oceanic plateau backstop. Details of rock chemistry and isotope systematics, coupled with the *geology* known for SSZ, are criteria that help distinguish the ophiolite assemblage from N-MORB crust formed at “bluewater” true mid-ocean ridges. The ophiolite conundrum, the puzzle, is solved once you understand the nature of SSZ systems. Yes, some ophiolites may be derived from “bluewater” spreading systems and emplaced on continents by obduction, although unusual tectonic settings must be involved. Consider-

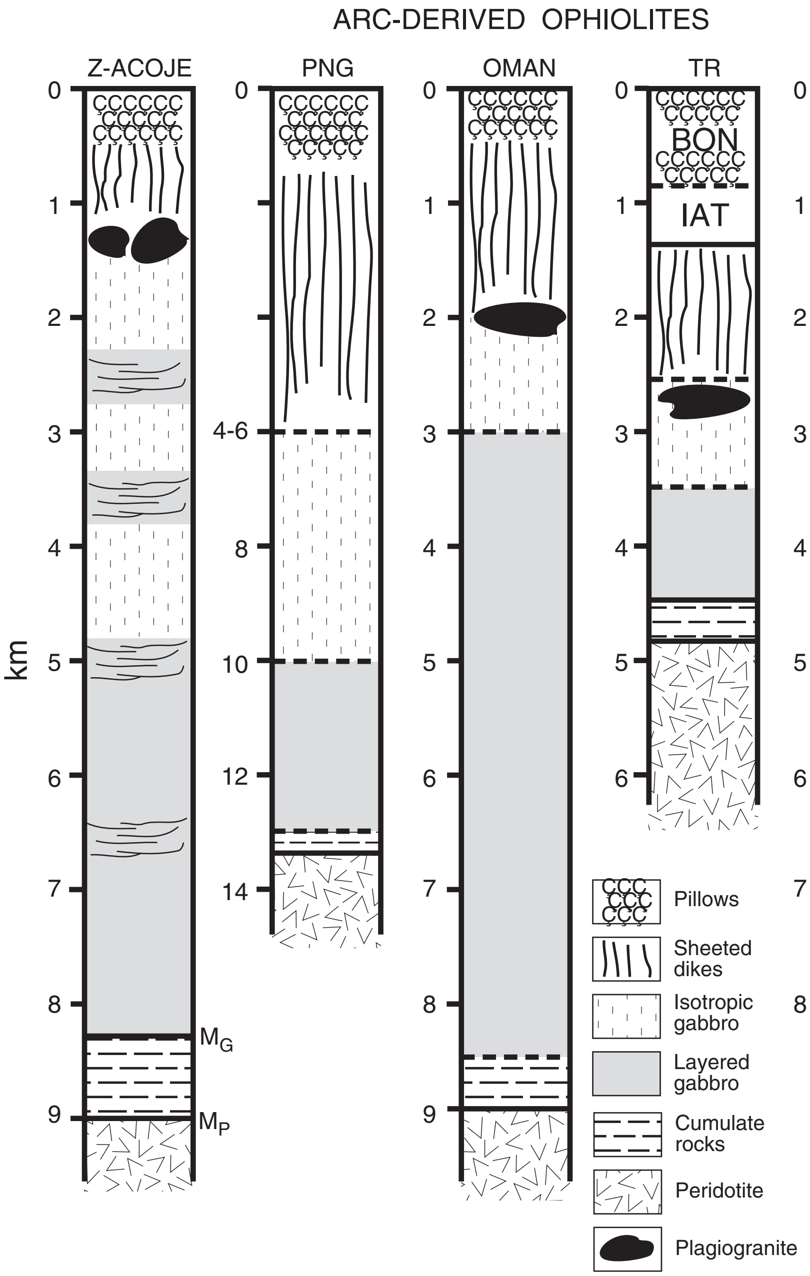

Figure 15. Comparison of oceanic crust with probable nascent arc ophiolite crust for: Z-ACOJE—Acoje, Zambales Range, Luzon; PNG—Papua, New Guinea; Oman; TR—Troodos; and NAS—Southern North Arm; BOI—Bay of Islands, Newfoundland. References are in Table 6.

ation of the geology and chemistry of ophiolites, their common rock associations, and the geology of SSZs leads me to reiterate an old claim of mine that many, perhaps most, ophiolites come from intra-oceanic plate convergence systems.

ACKNOWLEDGMENTS

This synthesis has benefitted from discussions, over many years, with co-workers and students. In particular I thank Bob Stern, Sherman Bloomer, Patty Fryer, Jim Gill, Brian Taylor, Jim Natland, and Bob Fisher for sharing their ideas and insights. The original manuscript has been improved by helpful comments and suggestions provided by Robert Stern, Warren Hamilton, and Yildirim Dilek. Special thanks to Jo Griffith for her help with illustrations. The author's research has been supported by research awards from the National Science Foundation, OCE, and the University of California—San Diego, USA.

REFERENCES CITED

Alabaster, T., Pearce, J.A., and Malpas, J., 1982, The volcanic stratigraphy and petrogenesis of the Oman ophiolite complex: Contributions to Mineralogy and Petrology, v. 81, p. 168–183.

Allwart, A.O., 1988, Working at cross-purposes; Holmes and Vening Meinesz on convection: Eos (Transactions, American Geophysical Union), v. 69, p. 899–906.

Arai, S., 1991, The circum-Izu massif peridotite, central Japan, as back -arc mantle fragments of the Izu-Bonin arc system, *in* Peters, Tj., Nicolas, A., and Coleman, R.G., eds., Ophiolite Genesis and Evolution of the Oceanic Lithosphere: Dordrecht, Kluwer Academic Publishers, p. 801–816.

Alexander, R.J., and Harper, G.D., 1992, The Josephine ophiolite: an ancient analogue for slow- to intermediate-spreading oceanic ridges, *in* Parson, L.M., Murton, B.J., and Browning, P., eds., Ophiolites and their Modern Oceanic Analogues: London, Geological Society of London Special Publication 60, p. 3–38.

Austin, J., A., Taylor, F.W., and Cagle, C.D., 1989, Seismic stratigraphy of the central Tonga Ridge: Marine and Petroleum Geology, v. 6, p. 71–92.

Bailey, E.H., Blake, M.C., and Jones, D.L., 1970, On-land Mesozoic oceanic crust in California coast ranges: Geological Survey Research, 1970, U.S., Geological Survey, Prof. Paper 700-C, p. C-70–C-81.

Baker, P.E., 1968, Comparative volcanology and petrology of the Atlantic island arcs: Bulletin of Volcanology, v. 32, p. 189–206.

Barazangi, M., and Isacks, B., 1971, Lateral variations of seismic-wave attenuation in upper mantle above the inclined earthquake zone of the Tonga Island Arc: Journal of Geophysical Research, v. 76, p. 8493–8516.

Barker, F., ed., 1979, Trondhjemites, Dacites, and Related Rocks: Elsevier, New York, 659 p.

Benioff, H., 1949, Seismic evidence for the fault origin of oceanic deeps: Geological Society of America Bulletin, v. 60, p. 1337–1356.

Ben Othman, D., White, W.M., and Patchett, J., 1989, The geochemistry of marine sediments, island arc magma genesis, and crust-mantle recycling: Earth and Planetary Science Letters, v. 94, p. 1–21.

Bernal, J.D., 1961, Continental and oceanic differentiation: Nature, v. 192, p. 123–124.

Bevis, M., Taylor, F.W., Schutz, B.E., Recy, J., Isacks, B.L., Hely, S., Singh, R., Kendrick, E., Stowell, J., Taylor, B., and Calmant, S., 1995, Geodetic observations of very rapid convergence and back-arc extension at the Tonga arc: Nature, v. 374, p. 249–251.

Bloomer, S.H., 1983, Distribution and origin of igneous rocks from the landward slope of the Mariana Trench: implications for its structure and evolution: Journal of Geophysical Research, v. 88, p. 7411–7428.

Bloomer, S.H., and Fisher, R.L., 1987, Petrology and geochemistry of igneous rocks from the Tonga Trench-A non-accreting plate boundary: Journal of Geology, v. 95, p. 469–495.

Bloomer, S.H., and Hawkins, J.W., 1983, Gabbroic and ultramafic rocks from the Mariana Trench, *in* Hayes, D.E., eds., The Tectonic Evolution of Southeast Asian Seas and Islands, pt. 2: Washington, D.C., American Geophysical Union Memoir 27, p. 294–317.

Bloomer, S.H., and Hawkins, J.W., 1987, Petrology and geochemistry of boninite series volcanic rocks from the Mariana Trench: Contributions to Mineralogy and Petrology, v. 97, p. 361–377.

Bloomer, S.H., Stern, R.J., Fisk, E., and Geshwind, C.H., 1989, Shoshonitic volcanism in the northern Mariana Arc 1. Mineralogic and major and trace element characteristics: Journal of Geophysical Research, v. 94, p. 4469–4496.

Bloomer, S.H., Ewart, A., Hergt, J.M., and Bryan, W.B., 1994, Geochemistry and origin of igneous rocks from the outer Tonga forearc (site 841), *in* Hawkins, J., Parson, L., Allan, J., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, p. 625–646.

Bloomer, S.H., Taylor, B., MacLeod, C.J., Stern, R.J., Fryer, P., Hawkins, J.W., and Johnson, L., 1995, Early arc volcanism and the ophiolite problem: A perspective from drilling in the Western Pacific, *in* Taylor, B., and Natland, J., eds., Active Margins and Marginal Basins of the Western Pacific: Washington, D.C., American Geophysical Union Monograph 88, p. 1–30.

Bostrom, K., and Peterson, M.N.A., 1966, Precipitates from hydrothermal exhalation on the East Pacific Rise: Economic Geology, v. 61, p. 1258–1265.

Bowin, C.O., Nalwalk, A.J., and Hersey, J.B., 1966, Serpentinized peridotite dredged from the north walls of the Puerto Rico Trench: Geological Society of America Bulletin, v. 77, p. 257–270.

Brocher, T.M., Karson, J.A., and Collins, J. A, 1985, Seismic stratigraphy of the oceanic Moho based on ophiolite models: Geology, v. 13, p. 62–65.

Bryan, W.B., 1979, Low K_2O dacite from the Tonga-Kermadec arc: Petrography, chemistry and petrogenesis, *in* Barker, F., ed., Trondhjemites, Dacites, and Related Rocks: Amsterdam, Elsevier, p. 581 –600.

Bryan, W.B., Stice, G.D., and Ewart, A.E., 1972, Geology, petrography, and geochemistry of the volcanic islands of Tonga: Journal of Geophysical Research, v. 77, p. 1566–1585.

Byers, F.M., Jr., 1961, Petrology of three volcanic suites, Umnak and Bogoslof Islands, Alaska: Geological Society of America Bulletin, v. 72, p. 93–128.

Capedri, S., Venturelli, G., and Toscani, L., 1982, Petrology of an ophiolitic cumulate sequence from Pindos, Greece: Geological Journal, v. 17, p. 223–242.

Clift, P.D., and Dixon, J.E., 1994, Variations in arc volcanism and sedimentation related to rifting of the Lau Basin*, in* Hawkins, J.W., Parson, L.M., Allan, J.F., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, p. 23–50.

Clift, P.D., MacLeod, C.J.,Tappin D.R., Wright, D.J., and Bloomer, S.H., 1998, Tectonic controls on sedimentation and diagenesis in the Tonga Trench and forearc, southwest Pacific: Geological Society of America Bulletin, v. 110, p. 483–496.

Cloud, P.E., Schmidt, R.G., and Burke, H.W., 1958, Geology of Saipan Island: U.S. Geological Survey Professional Paper 280-A, p. 1–111.

Coats, R.R., 1962, Magma type and crustal structure in the Aleutian Arc, *in* Macdonald, G.A., and Kuno, H., eds., The Crust of the Pacific Basin: American Geophysical Union, Geophysical Monograph 6, p. 92–109.

Coish, R.A., and Church, W.R., 1979, Igneous geochemistry of mafic rocks in the Betts Cove ophiolite, Newfoundland: Contributions to Mineralogy and Petrology, v. 70, p. 29–39.

Coney, P.J., Jones, D.A., and Monger, J.W.H., 1980, Cordilleran suspect terranes: Nature, v. 288, p. 329–333.

Coleman, R.G., 1977, Ophiolites, Ancient Oceanic Lithosphere?: New York, Springer-Verlag, 229 p.

Coleman, R.G., 2000, Prospecting for ophiolites along the California continental margin, *in* Dilek, Y., Moores, E.M., Elthon, D., and Nicolas, A., eds., Ophiolite and Oceanic Crust: New insights from field studies and the Ocean Drilling Program: Boulder, Colorado, Geological Society of America Special Paper 349, p. 351–364.

Crawford, A, J., ed., 1989, Boninites: London, Unwin Hyman, 465 p.

Crawford, A.J., Falloon, T.J., and Green, D.H., 1989, Classification, petrogenesis and tectonic setting of boninites, *in* Crawford, A.J., ed., Boninite: London, Unwin Hyman, p. 2–49.

Crawford, W.C., Hildebrand, J.A., Dorman, L.M., Webb, S.C., and Wiems, D.A., 2003, Tonga Ridge and Lau Basin crustal structure from seismic refraction data: Journal of Geophysical Research, v. 108, EPM 6, DOI 10.10291/2001JB00 1435.

Davies, H.L., and Smith, E.I., 1971, Geology of Eastern Papua: Geological Society of America Bulletin, v. 82, p. 3299–3312.

Davies, J.H., and Stevenson, D.J., 1992, Physical model of source region of subduction zone volcanics: Journal of Geophysical Research, v. 97, p. 2037–2070.

DeBari, S.M., and Coleman, R.G., 1989, Examination of the deep levels of an island arc: Evidence from the Tonsina ultramafic-mafic assemblage, Tonsina, Alaska: Journal of Geophysical Research, v. 94, p. 4373–4391.

DeBari, S.M., and Sleep, N.H., 1991, High-Mg, low-Al bulk composition of the Talkeetna island arc, Alaska: Implications for primary magmas and the nature of arc crust: Geological Society of America Bulletin, v. 103, p. 37–47.

Defant, M.J., and Drummond, M.S., 1990, Derivation of some modern arc magmas by melting of young subducted lithosphere: Nature, v. 347, p. 662–665.

Dewey, J.F., and Bird, J.M., 1971, Origin and emplacement of the ophiolite suite: Appalachian ophiolites of Newfoundland: Journal of Geophysical Research, v. 76, p. 3179–3206.

Dickinson, W.R., and Hatherton, T.R., 1967, Andesitic volcanism and seismicity around the Pacific: Science, v. 157, p. 801–803.

Dietrich, V.J., Emmerman, R., Oberhansli, R., and Puchelt, H., 1978, Geochemistry of basaltic and gabbroic rocks from the west Mariana Basin and the Mariana Trench: Earth and Planetary Science Letters, v. 38, p. 127–144.

Dietz, R.S., 1961, Continent and ocean basin evolution by spreading of the sea floor: Nature, v. 190, p. 854–857.

Dilek, Y., Moores, E.M., Elthon, D., and Nicolas, A., eds.*,* 2000, Ophiolites and Oceanic Crust: New insights from field studies and the Ocean Drilling Program: Boulder, Colorado, Geological Society of America Special Paper 349, 552 p.

Dixon, T.H., and Batiza, R., 1979, Petrology and chemistry of Recent lavas in the northern Marianas: Implications for the origin of island arc basalts: Contributions to Mineralogy and Petrology, v. 70, p. 167–181.

Dobson, P.F., and O'Neill, J.R., 1987, Stable isotope compositions and water content of boninite series volcanic rocks from Chichi-jima, Bonin islands, Japan: Earth and Planetary Science Letters, v. 82, p. 75–86.

Duncan, R.A., Vallier, T.L., and Falvey, D.A., 1985, Volcanic episodes at 'Eua, Tonga Islands, *in* Scholl, D.W., and Vallier, T.L., 1985, Geology and Offshore Resources of Pacific Island Arcs-Tonga Region: Houston, Texas,

Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, v. 2, p. 281–290.
Eggins, S.M., 1993, Origin and differentiation of picritic arc magmas, Ambae (Aoba) Vanuatu: Contributions to Mineralogy and Petrology, v. 114, p. 79–100.
Ellam, R.M., and Hawkesworth, C.J., 1988, Elemental and isotopic variations in subduction related basalts: Evidence for a three component model: Contributions to Mineralogy and Petrology, v. 98, p. 72–80.
Elsasser, W.M., 1971, Sea-floor spreading as thermal convection: Journal of Geophysical Research, v. 76, p.1101–1112.
Elthon, D., 1991, Geochemical evidence for formation of the Bay of Islands ophiolite above a subduction zone: Nature, v. 354, p. 140–143.
Ewart, A., and Bryan, W.B., 1972, Petrography and geochemistry of the igneous rocks from "Eua, Tongan Islands: Geological Society of America Bulletin, v. 83, p. 3281–3298.
Ewart, A., and Hawkesworth, C.J., 1987, The Pleistocene-Recent Tonga-Kermadec arc lavas: Interpretation of new isotopic and rare earth data in terms of a depleted mantle source model: Journal of Petrology, v. 28, p. 495–530.
Ewart, A., Bryan, W.B., and Gill, J.B., 1973, Mineralogy and geochemistry of the younger Tongan Islands, S.W., Pacific: Journal of Petrology, v. 14, p. 429–466.
Ewart, A., Brothers, R.N., and Mateen, A., 1977, An outline of the geology and geochemistry, and the possible petrogenetic evolution of the volcanic rocks of the Tonga-Kermadec-New Zealand island arc: Journal of Volcanology and Geothermal Research, v. 2, p. 205–250.
Falloon, T.J., Green, D.H., and McCullough, M.T., 1989, Petrogenesis of high-Mg and associated lavas from the northern Tonga Trench, *in* Crawford, A.J., ed., Boninite: London, Unwin Hyman, p. 357–395.
Falloon, T.J., Malahoff, A., Zonenschain, L.P., and Bogdanov, Y., 1992, Petrology and geochemistry of back-arc basin basalts from Lau Basin spreading ridges at 15, 18, 19º S: Mineralogy and Petrology, v. 47, p. 1–36.
Fisher, R.L., and Engel, C.G., 1969, Ultramafic and basaltic rocks dredged from the nearshore flank of the Tonga Trench: Geological Society of America Bulletin, v. 80, p. 1373 –1378.
Fisher, R.L., and Hess, H.H., 1963, Trenches *in* Hill, M.N., ed., The Sea, v. 3: New York, Wiley Interscience, p. 411–436.
Flower, M.F.J., and Levine, H.M., 1987, Petrogenesis of a tholeiite-boninite sequence from Ayios Mamas, Troodos ophiolite: evidence for splitting of a volcanic arc: Contributions to Mineralogy and Petrology, v. 97, p. 509–524.
Fryer, P., 1992, A synthesis of Leg 125 drilling of serpentinite seamounts on the Mariana and Izu-Bonin forearcs, *in* Fryer, P., Pearce, J.A., Stokking, L.B., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, p. 593–614.
Fryer, P., 1995, Geology of the Mariana Trough, *in* Taylor B., ed., Backarc Basins: Tectonics and Magmatism: New York, Plenum Press, p. 237–279.
Fryer, P., 1996, Evolution of the Mariana convergent plate margin system: Reviews of Geophysics, v. 34, p. 89–125.
Fryer, P.J., Sinton, J.M., and Philpotts, J.A., 1981, Basaltic glasses from the Mariana Trough, *in* Hussong, D.M., Uyeda, S., et al., Initial Reports of the Deep Sea Drilling Project, v. 60: Washington, D.C., U.S. Government Printing Office, p. 601–610.
Fryer, P., Pearce, J.A., Stokking, L.B., et al., eds., 1992, Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, 716 p.
Gans, P. B, Mahood, G.A., and Schermer, E., 1989, Synextensional magmatism in the Basin and Range province: A case study in the eastern Great Basin: Boulder, Colorado, Geological Society of America Special Paper 233, p. 1–53.
Garcia, M.G., Liu, N.W.K., and Muenow, D.W., 1979, Volatiles in submarine volcanic rocks from the Mariana Island arc and trough: Geochimica et Comochimica Acta, v. 43, p. 305–312.
Gill, J.B., 1970, Geochemistry of Viti Levu, Fiji, and its evolution as an island arc: Contributions to Mineralogy and Petrology, v. 27, p. 179–203.
Gill, J.B., 1981, Orogenic andesites and plate tectonics: New York, Springer-Verlag, 390 p.
Gill, J.B., 1987, Early geochemical evolution of an oceanic island arc and back-arc: Fiji and the South Fiji Basin: Journal of Geology, v. 95, p. 589–615.
Gill, J.B., and Stork, A.L., 1979, Miocene low-K dacites and and trondhjemites of Fiji, *in* Barker, F., ed., Trondhjemites, Dacites, and Related Rocks: New York, Elsevier, p. 629–649.
Green, T.H., and Ringwood, A.E., 1968, Genesis of the calc-alkaline igneous rock suite: Contributions to Mineralogy and Petrology, v. 8, p.105–162.
Gribble, R.F., Stern, R.J., Bloomer, S.H., Stuben, D.H., O'Hearn, T., and Newman, S., 1996, MORB mantle and subduction components interact to generate basalts in the southern Mariana Trough back-arc basin: Geochimica et Cosmochimica Acta, v. 60, p. 2153–2166.
Gribble, R.F., Stern, R.J., Bloomer, S.H., Newman, S, and O'Hearn, T., 1998, Chemical and isotopic composition of lavas from the northern Mariana Trough: Implications for magmagenesis in back-arc basins: Journal of Petrology, v. 39, p. 122–154.
Gust, D.A., and Perfit, M.R., 1987, Phase relations of a high-Mg basalt from the Aleutian Island Arc: Implications for primary island arc basalts and high-Al basalts: Contributions to Mineralogy and Petrology, v. 97, p. 7–18.
Hamilton, W.B., 1988, Plate tectonics and island arcs: Geological Society of America Bulletin, v. 100, p. 1503–1527.
Hamilton, W.B., 1995, Subduction systems and magmatism, *in* Smellie, J.L., ed., Volcanism Associated with Extension at Consuming Plate Margins: Geological Society Special Publication 81, p. 3–28.
Hamilton, W.B., 2002, The closed upper-mantle circulation of plate tectonics, *in* Stein S., and Freymuller, J.T., eds., Plate Boundary Zones: American Geophysical Union, Geodynamics Series, v. 30, p. 359–410.
Harper, G.D., 1984, The Josephine ophiolite, northwestern California: Geological Society of America Bulletin, v. 95, p.1009–1026.
Hart, S.R., and Zindler, A., 1986, In search of a bulk-earth composition: Chemical Geology, v. 57, p. 247–267.
Hart, S.R., Glassley, W.E., and Karig, D.E., 1972, Basalts and seafloor spreading behind the Mariana Island Arc: Earth and Planetary Science Letters, v. 15, p. 12 –18.
Hasebe, K., Fujii, N., and Uyeda, S., 1970, Thermal processes under island arcs: Tectonophysics, v. 10, p. 335–355.
Hawkins, J.W., 1974, Geology of the Lau Basin a marginal sea behind the Tonga Arc, *in* Burk, C., and Drake, C., eds., Geology of Continental Margins: New York, Springer, p. 505–520.
Hawkins, J.W., 1976, Petrology and geochemistry of basaltic rocks of the Lau Basin: Earth and Planetary Science Letters, v. 28, p. 283–298.
Hawkins, J.W., 1977, Petrological and geochemical characteristics of marginal basin basalts, *in* Talwani, M., and Pitman, W., eds., Island Arcs, Deep Sea Trenches and Back-Arc Basins: Washington, D.C., American Geophysical Union, Maurice Ewing Series 1, p. 355–365.
Hawkins, J.W., 1980, Petrology of back-arc basins and island arcs: Their possible role in the origin of ophiolites, *in* Panayiotou, A., ed., Ophiolites: Proceedings, International Ophiolite Symposium, 1979: Nicosia, Cyprus, Geological Survey Department, p. 244–255.
Hawkins, J.W., 1985, Low-K rhyolitic pumice from the Tonga Ridge, *in* Scholl, D.W., and Vallier, T.L., 1985, Geology and Offshore Resources of Pacific Island Arcs-Tonga Region: Houston, Texas, Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, v. 2, p. 171–178.
Hawkins, J.W., 1994, Petrologic synthesis, ODP Leg 135-Lau Basin transect, *in* Hawkins, J.W., Parson, L.M., Allan, J.F., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, p. 879–908.
Hawkins, J.W., 1995a, The geology of the Lau Basin *in* Taylor B., ed., Backarc Basins: Tectonics and Magmatism: New York, Plenum Press, p. 63–138.
Hawkins, J.W., 1995b, Evolution of the Lau Basin- Insights from ODP Leg 135, *in* Taylor, B., and Natland, J., eds., Active Margins and Marginal Basins of the Western Pacific: American Geophysical Union Monograph 88, p. 126–174.
Hawkins, J.W., and Allan, J.F., 1994, Petrologic evolution of Lau Basin Sites 834 through 839, *in* Hawkins, J.W., Parson, L.M., Allan, J.F., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, p. 427–470
Hawkins, J.W., and Castillo, P.R., 1998, Early history of the Izu-Bonin-Mariana arc system: Evidence from Belau and the Palau Trench: The Island Arc, v. 7, p. 559–578.
Hawkins, J.W., and Evans, C.A., 1983, Geology of the Zambales Range, Luzon, Philippines: ophiolite derived from an island arc-backarc basin pair, *in* Hayes, D.E., ed., The Tectonic Evolution of Southeast Asian Seas and Islands (pt 2): Washington, D.C., American Geophysical Union Memoir 27, p. 124–138.
Hawkins, J.W., and Falvey, D.A., 1985, Petrology of andesite dikes and flows from 'Eua, Tonga, *in* Scholl, D.W., and Vallier, T.L., 1985, Geology and Offshore Resources of Pacific Island Arcs-Tonga Region: Houston, Texas, Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, v. 2, p. 269–280.
Hawkins, J.W., and Florendo, F., 1994, Supra-subduction zone magmatism: Implications for the origin of Philippine ophiolites: Acta Geologica Taiwanica, v. 30, p. 63–171.

Hawkins, J.W., and Helu, S., 1986, Polymetallic sulfide deposit from"black smoker" chimney, Lau Basin: Eos, Transactions of American Geophysical Union, v. 67, p. 378.

Hawkins, J.W., and Melchior, J.T., 1985, Petrology of Mariana Trough and Lau Basin basalts: Journal of Geophysical Research, v. 90, p. 11,431–11,468.

Hawkins, J.W., Sclater, J., and Hohnhaus, G., 1970, Petrologic and geophysical characteristics of the Lau Basin Ridge: A spreading center behind the Tonga Arc: Geological Society of America Abstracts with Programs, v. 2, no. 7, p. 571.

Hawkins, J.W., Bloomer, S.H., Evans, C.A., and Melchior, J.T., 1984, Evolution of intra-oceanic arc-trench systems: Tectonophysics, v. 102, p. 175–205.

Hawkins, J.W., Lonsdale, P.F., Macdougall, J.D., and Volpe, A.M., 1990, Petrology of the axial ridge of the Mariana Trough backarc spreading center: Earth and Planetary Science Letters, v. 100, p. 226–250.

Hawkins, J.W., Parson, L.M., Allan, J.F., et al., eds., 1994, Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, 984 p.

Hawkins, J.W., Castillo, P., and Lonsdale, P., 2000, Geology of the Tonga Arc-Forearc-Trench system: Eos (Transactions, American Geophysical Union), v. 81, p. F1119.

Hayes, D.E., ed., 1980, The Tectonic Evolution of Southeast Asian Seas and Islands: Washington, D.C., American Geophysical Union Memoir 23, 326 p.

Hayes, D.E., ed., 1983, The Tectonic Evolution of Southeast Asian Seas and Islands (pt 2): Washington, D.C., American Geophysical Union Memoir 27, 396 p.

Hergt, J.M., and Hawkesworth, C.J., 1994, Pb-, Sr-, and Nd-isotopic evolution of the Lau Basin: implications for mantle dynamics during backarc opening, *in* Hawkins, J., Parson, L., Allan, J., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, p. 505–518.

Hess, H.H., 1948, Major structural features of the western North Pacific: An interpretation of H.O., 5485, bathymetric chart, Korea to New Guinea: Geological Society of America Bulletin, v. 59, p. 417–466.

Hess, H.H., 1962, History of ocean basins, *in* Engel, A.E.J., James, H.L., and Leonard, B.F., eds., Petrologic Studies: A Volume in honor of A.F., Buddington: New York, Geological Society of America, p. 599–620.

Hickey, R.L., and Frey, F.A., 1982, Rare earth element geochemistry of Mariana forearc volcanics, *in* Hussong, D.M., Uyeda, S., et al., Initial reports of the Deep Sea Drilling Project, v. 60: Washington, D.C., U.S. Government Printing Office, p. 735–742.

Hickey-Vargas, R., 1989, Boninites and tholeiites from DSDP Site 458, Mariana forearc, *in* Crawford, A.J., ed., Boninite: London, Unwin Hyman, p. 340–357.

Hodkinson, R.A., and Cronan, D.S., 1994, Variability in the hydrothermal component of the sedimentary sequence in the Lau backarc basin ODP Leg 135 Sites 834A, 835A, 835A, 837A, 838A and 839A, *in* Hawkins, J., Parson, L., Allan, J., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, p. 75–86.

Holmes, A., 1931, Radioactivity and earth movements: Transactions, Geological Society of Glasgow, v. 18, p. 559–606.

Holmes, A., 1965, Principles of Geology, second edition: New York, Ronald Press, 1288 p.

Hopson, C.A., Mattinson, J.M., and Pessagno, E.A., 1981, Coast Range ophiolite, western California, *in* Ernst, W.G., ed., The Geotectonic Development of California, Rubey Volume I: Englewood Cliffs, New Jersey, Prentice Hall, p. 418 –510.

Hussong, D.M., and Uyeda, S., editors, 1981, Initial reports of the Deep Sea Drilling Project Leg 60: Washington, U. S Government Printing Office, 929 p.

Ida, Y., 1983, Convection in the mantle wedge above the slab and the tectonic processes in subduction zones: Journal of Geophysical Research, v. 88, p. 7449–7456.

Ikeda, Y., and Yuasa, M., 1989, Volcanism in nascent back-arc basins behind the Shichito Ridge and adjacent areas in the Izu-Ogasawara arc, northwest Pacific: evidence for mixing between E-type MORB and island arc magmas at the initiation of back-arc rifting: Contributions to Mineralogy and Petrology, v. 101, p. 377–393.

Isacks, B.L., and Barazangi, M., 1977, Geometry of Benioff zones: Lateral segmentation and downwards bending of the subducted lithosphere, *in* Talwani, M., and Pitman, W., eds., Island Arcs, Deep Sea Trenches and Back-Arc Basins: Washington, D.C., American Geophysical Union, Maurice Ewing Series 1, p. 99–114.

Isacks, B.L., Oliver, J., and Sykes, L.R., 1968, Seismology and the new global tectonics: Journal of Geophysical Research, v. 73, p. 5855–5899.

Ishii, T., Robinson, P.T., Maekawa, H., and Fiske, R., 1992, Petrological studies of peridotites from diapiric serpentinite seamounts in the Izu-Ogasawara-Mariana forearcs, *in* Fryer, P., Pearce, J.A., Stokking, L.B., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, p. 445–486.

Jenner, G.A., Dunning, G.R., Malpas, J., Brown, M., and Brace, T., 1991, Bay of Islands and Little Port complexes, revisited: age, geochemical and isotopic evidence confirms suprasubduction-zone origin: Canadian Journal of Earth Science, v. 28, p. 1635–1652.

Johnson, L.E., Fryer, P., Taylor, B., Silk, M., Jones, D.L., Sliter, W.V., Itaya, T., and Ishii, T., 1991, New evidence for crustal accretion in the outer Mariana fore arc: Cretaceous radiolarian cherts and mid-ocean ridge basalt-like lavas: Geology, v. 19, p. 811–814.

Jones, G., and Robertson, A.H.F., 1991, Tectono-stratigraphy and evolution of the Mesozoic Pindos ophiolite and related units, northwestern Greece: Journal of the Geological Society of London, v. 148, p. 267–288.

Karig, D.E., 1970, Ridges and basins of the Tonga-Kermadec island arc system: Journal of Geophysical Research, v. 75, p. 239–254.

Kawate, S., and Arima, M., 1998, Petrogenesis of the Tanzawa plutonic complex, central Japan: Exposed felsic middle crust of the Izu-Bonin-Mariana arc: The Island Arc, v. 7, p. 342–358.

Kay, S.M., and Kay, R.W., 1985, Role of crystal cumulates and the oceanic crust in the formation of the lower crust of the Aleutian arc: Geology, v. 13, p. 461–464.

Keunen, Ph. H., 1950, Marine Geology: New York, Wiley, 568 p.

Kikuchi, Y., 1890, On pyroxene compositions in certain volcanic rocks from Bonin Island: Journal of College of Science, Imperial University of Japan, v. 3, p. 67–89.

Kobayashi, K., 1983, Fore-arc volcanism and cycles of subduction: *in* Shimozuru, D., and Yokoyama, I., eds., Arc Volcanism: Physics and Tectonics: Tokyo, Terra Scientific Publishing Co., p. 153–163.

Kroenke, L., Scott, R., et al., eds., 1980, Initial reports of the Deep Sea Drilling Project Leg 59: Washington, D.C., U. S Government Printing Office, 820 p.

Latouche, C., Maillet, N., and Blanchet, R., 1981, X-ray mineralogy studies, Deep Sea Drilling Project Leg 60, *in* Hussong, D.M., Uyeda, S., et al., eds., Initial reports of the Deep Sea Drilling Project Leg 60, p. 437–454.

Lee, J., and Stern, R.J., 1995, Forty million years of magmatic evolution in the Mariana arc: The glass tephra record: Journal of Geophysical Research, v. 100, p. 17,671–17,687.

Leeman, W.R., 1996, Boron and other fluid-mobile elements in volcanic arc lavas: Implications for subduction processes, *in* Bebout, G.E., Scholl, D.W., Kirby, S.H., and Platt, J.P., eds., Subduction Top to Bottom: Washington, D.C., American Geophysical Union Monograph 96, p. 269 –276.

Lin, P-N., Stern, R.J., and Bloomer, S.H., 1989, Shoshonitic volcanism in the northern Mariana Arc 2. Large-ion lithophile and rare earth element abundances: Evidence for the source of incompatible element enrichments in intra-oceanic arcs: Journal of Geophysical Research, v. 94, p. 4497–4514.

Maekawa, H., Shozui, M., Ishii, T., Saboda, K., and Ogawa, Y., 1992, Metamorphic rocks from the serpentinite seamounts in the Mariana and Izu-Ogasawara forearcs, *in* Fryer, P., Pearce, J.A., Stokking, L.B., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, p. 415–430.

Marlow, M.S., Johnson, L.E., Pearce, J.A., Fryer, P.B., Pickthorn, L.G., and Murton, B.J., 1992, Pleistocene volcanic rocks in the Mariana forearc arc revealed by drilling at Site 781, *in* Fryer, P., Pearce, J.A., Stokking, L.B., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, p. 293–310.

Marshall, P., 1911, Oceania: *in* Handbuch der regionalen Geologie, v. 7, 36 p.

Martinez, F., Fryer, P., Baker, N., and Yamakazi, T., 1995, Evolution of backarc rifting: Mariana Trough, 20°–24° N: Journal of Geophysical Research, v. 100, p. 3807–3827.

Maury, R.C., Sajona, F.G., Pubellier, M., Bellon, H., et Defant, M.J., 1996, Fusion de la croute oceanique dans les zones de subduction/collision recentes: l'exemple de Mindanao (Philippines): Bulletin de la Societe Geologique de France, t. 167, p. 579–595.

McDougall, I., 1994, Dating of rhyolite glass in the Tonga forearc (Hole 841 B), *in* Hawkins, J., Parson, L., Allan, J., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, p. 923–924.

McKenzie, D., and Bickle, M.J.,1988, The volume and composition of melt generated by extension of lithosphere: Journal of Petrology, v. 29, p. 625–679.

Meijer, A., 1976, Pb and Sr isotopic data bearing on the origin of volcanic rocks from the Mariana island arc system: Geological Society of America Bulletin v. 87, p. 1358–1369.

Meijer, A., 1980, Primitive arc volcanism and a boninite series: Examples from western Pacific island arcs, *in* Hayes, D.E., ed., The Tectonic Evolution of Southeast Asian Seas and Islands: Washington, D.C., American Geophysical Union Memoir 23, p. 272–282.

Meijer, A., 1983, The origin of low-K rhyolites from the Mariana frontal arc: Contributions to Mineralogy and Petrology, v. 83, p. 45–51.

Meijer, A., and Reagan, M., 1981, Petrology and geochemistry of the island of Sarigan in the Mariana Arc: Calc-alkaline volcanism in an oceanic setting: Contributions to Mineralogy and Petrology, v. 77, p. 37–354.

Metcalf, R.V., Wallin, E.T., Willse, K.R., and Muller, E.R., 2000, Geology and geochemistry of the ophiolitic Trinity terrane, California: Evidence of middle Paleozoic depleted supra-subduction zone magmatism in a proto-arc setting, *in* Dilek, Y., Moores, E.M., Elthon, D., and Nicolas, A., eds., Ophiolite and Oceanic Crust: New insights from field studies and the Ocean Drilling Program: Boulder, Colorado, Geological Society of America Special Paper 349, p. 403–418.

Menzies, M., Blanchard, D., and Jacobs, J., 1977, Rare earth and trace element chemistry of metabasalts from the Point Sal ophiolite: Earth and Planetary Science Letters, v. 37, p. 203–215.

Minear, J.W., and Toksoz, M.N., 1970, Thermal regime of a downgoing slab and new global tectonics: Journal of Geophysical Research, v. 75, p. 1397–1419.

Miyashiro, A., 1973, The Troodos ophiolite was probably formed in an island arc: Earth and Planetary Science Letters, v. 19, p. 218–224.

Miyashiro, A., 1974, Volcanic rock series in island arcs and active continental margins: American Journal of Science, v. 274, p. 321–355.

Moberly, R., 1972, Origin of lithosphere behind island arcs with reference to the western Pacific: Boulder, Colorado, Geological Society of America Memoir 132, p. 35–55.

Moores, E.M., and Jackson, E.D., 1974, Ophiolites and oceanic crust: Nature, v. 250, p. 136 –138.

Moores, E.M., and Vine, F.J., 1971, The Troodos massif Cyprus as oceanic crust: evaluation and implications: Philosophical Transactions Royal Society of London, ser. A, v. 268, p. 443–466.

Moores, E.M., Kellog, L.H., and Dilek, Y., 2000, Tethyan ophiolites, mantle convection, and tectonic "historical contingency": A resolution of the "ophiolite conundrum," *in* Dilek, Y., Moores, E.M., Elthon, D., and Nicolas, A., eds., Ophiolite and Oceanic Crust: New insights from field studies and the Ocean Drilling Program: Boulder, Colorado, Geological Society of America Special Paper 349, p. 3–12.

Morgan, W.J., 1968, Rises, trenches, great faults, and crustal blocks: Journal of Geophysical Research, v. 75, p. 1397–1419.

Morris, J.D., Leeman, W.P., and Tera, F., 1990, The subducted component in island arc lavas: constraints from Be isotopes and B–Be systematics: Nature, v. 344, p. 31–36.

Mrzowski, C.L., and Hayes, D.E., 1980, A seismic reflection study of faulting in the Mariana fore arc, *in* Hayes, D.E., eds., The Tectonic Evolution of Southeast Asian Seas and Islands: Washington, D.C., American Geophysical Union Memoir 23, p. 223–234.

Natland, J.H., 1981, Petrography and mineral composition of gabbros recovered in DSDP Hole 453 on the western side of the Mariana Trough, *in* Hussong, D.M., Uyeda, S., et al., Initial Reports of the Deep Sea Drilling Project, v. 60, p. 579–599.

Noiret, G., Montigny, R., and Allegre, C., 1981, Is the Vourinos Complex an island arc ophiolite ?: Earth and Planetary Science Letters, v.56, p. 357–386.

Ohara, Y., Stern, R.J., Ishii, T., Yurimoto, H., Yamakazi, T., 2002, Peridotites from the Mariana Trough: first look at the mantle beneath an active back-arc basin: Contributions to Mineralogy and Petrology, v. 143, p. 1–18.

O'Neill, H.C., and Palme, H., 1998, Composition of the silicate earth: Implications for accretion and core formation, *in* Jackson, I., ed., The Earth's Mantle: New York, Cambridge University Press, 566 p.

Oversby, V.M., and Ewart, A., 1972, Lead isotopic compositions of Tonga-Kermadec volcanics and their petrogenetic significance: Contributions to Mineralogy and Petrology, v. 37, p. 181–210.

Oxburgh, E.R., and Turcotte, D.L., 1970, Thermal structure of island arcs: Geological Society of America Bulletin, v. 81, p. 1665–1688.

Packham, G.H., and Falvey, D.A., 1971, An hypothesis for the formation of marginal seas in the western Pacific: Tectonophysics, v. 11, p. 79–109.

Panayiotou, A., ed., 1980, Ophiolites: Proceedings, International Ophiolite Symposium, 1979: Nicosia, Cyprus, Geological Survey Department, 781 p.

Parson, L.M., Rothwell, R.G., and MacLeod, C.J., 1994, Tectonics and sedimentation in the Lau Basin (southwest Pacific), *in* Hawkins, J.W., Parson, L.M., Allan, J.F., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, p. 9–22.

Peacock, S.M., 1990, Fluid processes in subduction zones: Science, v. 248, p. 247–336.

Pearce, J.A., 1975, Basalt geochemistry used to investigate past tectonic environments on Cyprus: Tectonophysics, v. 25, p. 41–67.

Pearce, J.A., 1980, Geochemical evidence for the genesis and eruptive setting of lavas from Tethyan ophiolites, *in* Panayiotou, A., ed., Ophiolites: Proceedings, International Ophiolite Symposium, 1979: Nicosia, Cyprus, Geological Survey Department, p. 261–272.

Pearce, J.A., and Cann, J.R., 1973, Tectonic setting of basic volcanic rocks determined using trace element analyses: Earth and Planetary Science Letters, v.19, p. 290–300.

Pearce, J.A., and Norry, M.J., 1979, Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks: Contributions to Mineralogy and Petrology, v. 37, p. 33–47.

Pearce, J.A., Alabaster, T., Shelton, A.W., and Searle, M.P., 1981, The Oman ophiolite as a Cretaceous arc-basin complex: evidence and implications: Philosophical Transactions Royal Society of London, ser. A, v. 300, p. 299–317.

Pearce, J.A., Lippard, S.J., and Roberts, S., 1984, Characteristics and tectonic significance of supra-subduction zone ophiolites, *in* Kokelaar, B.P., and Howells, M.F., eds., Marginal Basin Geology: Geological Society Special Publication 16, London, Blackwell, p. 77–94.

Pearce, J.A., van der Laan, S.R., Arculus, R.J., Murton, B.J., Ishii, T., Peate, D.W., and Parkinson, I., 1992, Boninite and harzburgite from Leg 125 (Bonin-Mariana forearc): A case study of magmagenesis during the initial stages of subduction, *in* Fryer, P., Pearce, J.A., Stokking, L.B., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 125: College Station, Texas, Ocean Drilling Program, p. 623–659.

Pearce, J.A., Ernewein, M., Bloomer, S.H., Parson, L.M., Murton, B.J., and Johnson, L.E., 1995, Geochemistry of Lau Basin volcanic rocks: influence of ridge segmentation and arc proximity, *in* Smellie, J, ed., Volcanism Associated with Extension at Consuming Plate Margins: London, Geological Society of London Special Publication 81, p. 53–75.

Peate, D.W., and Pearce, J.A., 1998, Causes of spatial compositional variations in Mariana arc lavas: Trace element variations: The Island Arc, v. 7, p. 479–495.

Perfit, M.R., Gust, D.A., Bence, A.E., Arculus, R.J., and Taylor, S.R., 1980, Chemical characteristics of island-arc basalts: Implications for mantle sources: Chemical Geology, v. 30, p. 227–256.

Peters, Tj., Nicolas, A., and Coleman, R.G., eds., 1991, Ophiolite Genesis and Evolution of the Oceanic Lithosphere: Dordrecht, Kluwer Academic Publishers, 903 p.

Poreda, R., and Craig, H., 1993, He and Sr isotopes in the Lau basin mantle: Depleted and primitive mantle components: Earth and Planetary Science Letters, v. 119, p. 319–329.

Price, R.C., Johnson, L.E., and Crawford, A.J., 1990, Basalts of the North Fiji Basin: the generation of back arc basin magmas by mixing of depleted and enriched mantle sources: Contributions to Mineralogy and Petrology, v. 105, p. 106–121.

Raitt, R.W., Fisher, R.L., and Mason, R.G., 1955, Tonga Trench *in* Poldervaart, A. ed. Crust of the Earth: Boulder, Colorado, Geological Society of America Special Paper 62, p. 217 –254.

Ramsay, W.R.H., Crawford, A.J., and Foden, J.F., 1984, Field setting, mineralogy, and genesis of arc picrites, New Georgia, Solomon Islands: Contributions to Mineralogy and Petrology, v. 88, p. 386–402.

Reagan, M.K., and Meijer, A., 1984, Geology and geochemistry of early arc-volcanic rocks from Guam: Geological Society of America Bulletin, v. 95, p. 701 –713.

Rehrig, W.A., 1986, Processes of regional Tertiary extension in the western Cordillera: Insights from the metamorphic core complexes, *in* Mayer, L. ed., Extensional Tectonics of the Southwestern United States: A Perspective on Processes and Kinematics: Boulder, Colorado, Geological Society of America Special Paper 208, p. 97–122.

Ringwood, A.E., 1969, Composition and evolution of the upper mantle, *in* Hart, P.J., ed., The Earth's Crust and Upper Mantle: American Geophysical Union, Geophysical Monograph 13, p. 1–17.

Ringwood, A.E., 1974, The petrological evolution of island arc systems: Journal of Geological Society of London, v. 130, p. 183–204.

Ringwood, A.E., 1975, Composition and Petrology of the Earth's Mantle: New York, McGraw Hill, 604 p.

Robertson, A.H.F., and Hudson, J.D., 1973, Cyprus umbers: Chemical precipitates on a Tethyan ocean ridge: Earth and Planetary Science Letters, v.18, p. 93–101.

Robinson, P.T., Melson, W.G., O'Hearn, T., Schmincke, H.U., 1983, Volcanic glass compositions from the Troodos ophiolite, Cyprus: Geology, v.11, p. 400–404.

Rodolfo, K.S., 1969, Bathymetry and marine geology of the Andaman Basin, and tectonic implications for southeast Asia: Geological Society of America Bulletin, v. 80, p. 1203–1230.

Rothwell, R.G., Bednarz, U., Boe, R., Clift, P.D., Hodkinson, R.A., Ledbetter, J.K., Pratt, C.E., and Soakai, S., 1994, Sedimentation and sedimentary processes in the Lau backarc basin: Results from Leg 135, *in* Hawkins, J.W., Parson, L.M., Allan, J.F., et al., eds., Proceedings, Ocean Drilling Program, Scientific Results, 135: College Station, Texas, Ocean Drilling Program, p. 829–842.

Salisbury, M.H., and Christensen, N.K., 1978, The seismic velocity structure of a traverse through the Bay of Islands ophiolite complex, Newfoundland, an exposure of oceanic crust and upper mantle: Journal of Geophysical Research, v. 83, p. 805–817.

Saunders, A.D., Tarney, J., Marsh, N.G., and Wood, D.A., 1980, Ophiolites as ocean crust or marginal basin crust: A geochemical approach, *in* Panayiotou, A., ed., Ophiolites: Proceedings, International Ophiolite Symposium, 1979: Nicosia, Cyprus, Geological Survey Department, p. 193–204.

Sharaskin A. Ya., Dobretsov, N.L., and Sobolev, N.V., 1980, Marianites: the clinoenstatite bearing pillow-lavas with the ophiolite assemblage of Mariana Trench, *in* Panayiotou, A., ed., Ophiolites: Proceedings, International Ophiolite Symposium, 1979: Nicosia, Cyprus, Geological Survey Department, p. 473–479.

Schmincke, H.U., Rautenschlein M., Robinson, P., and Mehegan, J.M., 1983, Troodos extrusive series of Cyprus: A comparison with oceanic crust: Geology, v. 11, p. 405–409.

Shervais, J.W., 1982, Ti -V plots and the petrogenesis of modern and ophiolitic lavas: Earth and Planetary Science Letters, v. 59, p. 101–118.

Shervais, J.W., and Kimbrough, D.L., 1985, Geochemical evidence for the tectonic setting of the Coast Range ophiolite: A composite island arc-oceanic crust terrane in western California: Geology, v. 13, p. 35–38.

Sigurdsson, I.A., Kamenetsky, V.S., Crawford, A.J., Eggins, S.M., and Zlobin, S.K., 1993, Primitive island arc and oceanic lavas from the Hunter ridge-Hunter fracture zone. Evidence from glass, olivine, and spinel compositions: Mineralogy and Petrology, v. 47, p. 149–169.

Sinton, J., and Fryer, P., 1987, Mariana Trough lavas from 18 N; Implications for the origin of backarc basin basalts: Journal of Geophysical Research, v. 92, p. 12,782–12,802.

Sleep, N., and Toksoz, M.N., 1971, Evolution of marginal basins, Nature, v. 233, p. 548–550.

Smith, G., Wiens, D.A., Fischer, K.M., Dorman, L.M., Webb, S.C., and Hildebrand, J.A., 2001, A complex pattern of mantle flow in the Lau backarc: Science, v. 292, p. 713–716.

Stein, S., and Stein, C.A., 1996, Thermo-mechanical evolution of oceanic lithosphere: Implications for the subduction process and deep earthquakes, *in* Bebout, G., E., Scholl, D.W., Kirby, S.H., and Platt, J.P., eds., Subduction Top to Bottom: American Geophysical Union Monograph 96, p. 1 –18.

Steinmann, G., 1927, Die ophiolitischen Zonen in der mediterranean Kettengebirgen: International Geological Congress, 14th, Madrid, Proceedings, v. 2, p. 638–667.

Stern, R.J., 1979, On the origin of andesite in the northern Mariana Island Arc: Implications from Agrigan: Contributions to Mineralogy and Petrology, v. 68, p. 207–219.

Stern, R.J., 2002, Subduction zones: Reviews of Geophysics, v. 40, 10.1029/2001 RG 000108.

Stern, R.J., and Arima, M., editors,1998, Special Issue; Geophysical and geochemical studies of the Izu-Bonin-Mariana Arc: The Island Arc, v. 7, p. 295–607.

Stern, R.J., and Bloomer, S. H, 1992, Subduction zone infancy: Examples from the Eocene Izu-Bonin-Mariana and Jurassic California arcs: Geological Society of America Bulletin, v.104, p. 1621–1636.

Stern, R.J., Bloomer, S.H., Lin, P-N., and Smoot, N.C., 1989, Submarine arc volcanism in the southern Mariana Arc as an ophiolite analogue: Tectonophysics, v. 168, p. 151–170.

Stern, R.J., Lin P. N, Morris, J.D., Jackson, M.C., Fryer, P., Bloomer, S.H., and Ito. E., 1990, Enriched back-arc basin basalts from the northern Mariana Trough: implications for the magmatic evolution of back-arc basins: Earth and Planetary Science Letters, v. 100, p. 210–225.

Stern, R.J., Morris, J., Bloomer, S.H., and Hawkins, J.W., 1991, The source of the subduction component in convergent margin magmas: Trace element and radiogenic isotope evidence from Eocene boninites, Mariana forearc: Geochimica et Cosmochimica Acta, v. 55, p. 1467–1481.

Stern, R.J., Bloomer, S.H., Martinez, F., Yamazaki, Y., and Harrison, T.M., 1996, The composition of back-arc basin lower crust and upper mantle in the Mariana Trough: A first report: The Island Arc, v. 5., p. 354–372.

Stolper, E., and Newman, S., 1994, The role of water in the petrogenesis of Mariana trough magmas: Earth and Planetary Science Letters, v. 121, p. 293–325.

Stolper, E., and Walker, D., 1980, Melt density and the average composition of basalt: Contributions to Mineralogy and Petrology, v. 74, p. 7–12.

Straub, S.M., 1995, Contrasting compositions of Mariana Trough fallout tephra and Mariana Island arc volcanics: a fractional crystallization link: Bulletin of Volcanology, v. 57, p. 403–421.

Sturm, M.E., Klein E.M., Karsten, J.L., and Karson, J.A., 2000, Evidence for subduction- related contamination of the mantle beneath the southern Chile Ridge: Implications for ambiguous ophiolite compositions, *in* Dilek, Y., Moores, E.M., Elthon, D., and Nicolas, A., eds., Geological Society of America Special Paper 349, p. 13 –20.

Sun, S.S., and McDonough, W.F., 1989, Chemical and isotopic systematics of oceanic basalts: Implications for mantle composition and processes, *in* Magmatism in the Ocean Basins, Saunders, A.D., and Norry, M.J., eds., Geological Society Special Publication, London, v. 42, p. 323–346.

Sunkel, G., 1990, Origin of petrological and geochemical variations of Lau Basin lavas (SW Pacific): Marine Mining, v. 9, p. 205–234.

Suyehiro, K., et al., 1996, Continental crust, crustal underplating, and low-Q upper mantle beneath an oceanic island arc: Science, v. 272, p. 390–392.

Tatsumi, Y., 1982, Origin of high-magnesian andesites in the Setouchi volcanic belt, southwest Japan, II, Melting phase relations at high pressure: Earth and Planetary Science Letters, v. 60, p. 305–317.

Tatsumi, Y., and Eggins, S.E., 1995, Subduction Zone Magmatism: Blackwell Science, Oxford, England, 211 p.

Taylor, B., 1992, Rifting and the volcanic-tectonic evolution of the Izu-Bonin-Mariana Arc, *in* Taylor, B., Fujioka, K., et al., eds., Proceedings ODP Science Results, 126, Ocean Drilling Program: College Station Texas, p. 627–651.

Taylor, B., ed., 1995, Backarc Basins: New York, Plenum Press, 524 p.

Taylor, B., and Natland, J., eds., 1995, Active Margins and Marginal Basins of the Western Pacific: Washington, D.C., American Geophysical Union Monograph 88, 417 p.

Taylor, B, Fujioka, K., et al., editors, 1992, Proceedings, Ocean Drilling Program, Scientific Results, 126: College Station, Texas, Ocean Drilling Program, 709 p.

Taylor, R.N., and Nesbitt, R.W., 1995, Arc volcanism in an extensional regime at the initiation of subduction: a geochemical study of Hahajima, Bonin Islands, Japan, *in* Smellie, J, ed., Volcanism Associated with Extension at Consuming Plate Margins: Geological Society of London Special Publication 81, p. 115–134.

Taylor, R.N., Murton, B.J., and Nesbitt, R.W., 1992, Chemical transects across intra-oceanic arcs: implications for the tectonic setting of ophiolites, *in* Parson, L.M., Murton, B.J., and Browning, P., eds., Ophiolites and their Modern Oceanic Analogues: Geological Society of London Special Publication 60, p. 117–132.

Taylor, R.N., Nesbitt, R.W., Vidal, P., Harmon, R.S., Auvray, B., and Croudace, I., 1994, Mineralogy, chemistry, and genesis of the boninite series volcanics, Chichijima, Bonin Islalnds, Japan: Journal of Petrology, v. 35, p. 577–617.

Tera, F., Brown, L., Morris, J., Sacks, S., Klein J., and Middleton, R., 1986, Sediment incorporation in island arc magmas: Inferences from ^{10}Be: Geochimica et Cosmochimica Acta, v. 50, p. 535–550.

Thy, P., and Xenophontos, C., 1991, Crystallization orders and phase chemistry of glassy lavas from the pillow sequences, Troodos ophiolite, Cyprus: Journal of Petrology, 32, p. 403–424.

Toksoz, M.N., and Bird, P., 1977, Formation and evolution of marginal basins and continental plateaus, *in* Talwani, M., and Pitman, W., eds., Island Arcs, Deep Sea Trenches and Back-Arc Basins: Washington, D.C., American Geophysical Union, Maurice Ewing Series 1, p. 379–394.

Uyeda, S., and Kanamori, H., 1979, Backarc opening and the mode of subduction: Journal of Geophysical Research, v. 84, p.1049–1061.

Vallier, T.L., Stevenson, A.J., and Scholl, D.W., 1985, Petrology of igneous rocks from Ata Island, Kingdom of Tonga, *in* Scholl, D.W., and Vallier, T.L., 1985, Geology and Offshore Resources of Pacific Island Arcs-Tonga Region: Houston, Texas, Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, v. 2, p. 301–306.

Vallier, T.L., Scholl, D.W., Fisher, M.A., Bruns, T.R., Wilson, F.H., von Heune, R., and Stevenson, A.J., 1994, Geologic framework of the Aleutian arc, Alaska, *in* Plafker, G., and Berg, H.C., eds., The Geology of Alaska: Boulder, Colorado, Geological Society of America, The Geology of North America, v. G-1, p. 367–388.

van der Hilst, R., 1995, Complex morphology of subducted lithosphere in the mantle beneath the Tonga trench: Science, v. 374, p. 154–157.

Varne, R., Brown, A., and Falloon, T., 2000, Macquarie Island: Its geology, structural history and the timing and tectonic setting of its N-MORB to E-MORB magmatism, *in* Dilek, Y., Moores, E.M., Elthon, D., and Nicolas, A., eds., Ophiolites and Oceanic Crust: Geological Society of America Special Paper 349, p. 301–320.

Viereck, L.G., Flower, M.F.J., Hertogen, J., Schmincke, H.-U., and Jenner, G.A., 1989, The genesis and significance of N-MORB sub-types: Contributions to Mineralogy and Petrology, v. 102, p. 112–126.

Volpe, A.M., Macdougall, J.D., and Hawkins, J.W., 1987, Mariana Trough basalts (MTB): trace element and Sr-Nd isotopic evidence for mixing between MORB-like and Arc-like melts: Earth and Planetary Science Letters, v. 82, p. 241–254.

Volpe, A.M., Macdougall, J.D., and Hawkins, J.W., 1988, Lau Basin basalt (LBB): trace element and Sr-Nd isotopic evidence for heterogeneity in backarc basin mantle: Earth and Planetary Science Letters, v. 90, p.174–186.

Volpe, A.M., Macdougall, J.D., Lugmair, G., Hawkins, J.W., and Lonsdale, P.F., 1990, Fine-scale isotopic variation in Mariana Trough basalts: evidence for heterogeneity and a recycled component in backarc basin mantle: Earth and Planetary Science Letters, v. 100, p. 251–264.

von Stackleberg, U., and Shipboard Party, 1988, Active hydrothermalism in the Lau backarc basin (SW Pacific): First results from the *Sonne 48* cruise (1987): Marine Minerals, v. 7, p. 431–442.

Wallin, E.T., and Metcalf, R.V., 1998, Supra-subduction zone ophiolite formed in an extensional forearc: Trinity Terrane, Klamath Mountains, California: Journal of Geology, v. 106, p. 591–608.

Weaver, S.D., Saunders, A.D., Pankhurst, R.J., and Tarney, J., 1979, A geochemical study of magmatism associated with the initial stages of back-arc spreading: Contributions to Mineralogy and Petrology, v. 68, p. 151–169.

Wegener, A. 1915, Die Entstehung der Kontinente und Ozeane, Braunschweig, F. Vieweg & Sohn, 242 p.

Wegener, A., 1966, The Origin of Continents and Oceans (translated by J. Biram): New York, Dover Publications, 246 p.

Whelan, P.M., Gill, J.B., Kollman, E., Duncan, R., and Drake, R.E., 1985, Radiometric dating of magmatic stages in Fiji, *in* Scholl, D.W., and Vallier, T.L., 1985, Geology and Offshore Resources of Pacific Island Arcs-Tonga Region: Houston, Texas, Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, v. 2, p. 415–440.

Wood, D.A., Marsh, N.G., Tarney, J., Joron, L.J., Fryer, P., and Treuil, M., 1981, Geochemistry of igneous rocks recovered from a transect across the Mariana trough, arc, fore-arc, and trench, sites 453 through 461, Deep Sea Drilling Project Leg 60, *in* Hussong, D.M., Uyeda, S., et al., eds., Initial Reports of the Deep Sea Drilling Project, v. 60, p. 611–645.

Woodhead, J.D., 1988, The origin of geochemical variations in Mariana lavas: A general model for petrogenesis in intra-oceanic island arcs: Journal of Petrology, v. 29, p. 805–830.

Woodhead, J.D., 1989, Geochemistry of the Mariana arc (western Pacific): Source composition and processes: Chemical Geology, v. 76, p. 1–24.

Woodhead, J.D., Eggins, S., and Gamble, J., 1993, High field strength and transition element systematics in island arc and back-arc basin basalts: evidence for multi-phase melt extraction and a depleted mantle wedge: Earth and Planetary Science Letters, v. 114, p. 491–504.

You, C.F., Castillo, P.R., Gieskes, J.M., and Spivack, A.J., 1996, Trace element behavior in hydrothermal experiments: Implications for fluid processes at shallow depths in subduction zones: Earth and Planetary Science Letters, v. 140, p. 41–52.

Zhao, D., Xu,Y., Wiens, D. A, Dorman, L., Hildebrand, J., and Webb, S., 1997, Depth extent of the Lau back-arc spreading center and its relation to subduction processes: Science, v. 278, p. 254–257.

Manuscript Accepted by the Society May 2, 2003