Subduction zone infancy: Examples from the Eocene Izu-Bonin-Mariana and Jurassic California arcs

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ABSTRACT

A new model for the earliest stages in the evolution of subduction zones is developed from recent geologic studies of the Izu-Bonin-Mariana (IBM) arc system and then applied to Late Jurassic ophiolites of Cailfornia. The model accounts for several key observations about the earliest stages in the evolution of the IBM system: (1) subduction nucleated along an active transform boundary, which separated younger, less-dense lithosphere in the west from older, more-dense lithosphere to the east; (2) initial arc magmatic activity occupied a much broader zone than existed later; (3) initial magmatism extended up to the modern trench, over a region now characterized by subnormal heat flow; (4) early arc magmatism was characterized by depleted (tholeiitic) and ultra-depleted (boninitic) magmas, indicating that melting was more extensive and involved more depleted mantle than is found anywhere else on earth; (5) early arc magmatism was strongly extensional, with crust forming in a manner similar to slow-spreading ridges; and (6) crust production rates were 120 to 180 km^3/km-Ma, several times greater than for mature arc systems. These observations require that the earliest stages of subduction involve rapid retreat of the trench; we infer that this resulted from continuous subsidence of denser lithosphere along the transform fault. This resulted in strong extension and thinning of younger, more buoyant lithosphere to the west. This extension was accompanied by the flow of water from the sinking oceanic lithosphere to the base of the extending lithosphere and the underlying asthenosphere. Addition of water and asthenospheric upwelling led to catastrophic melting, which continued until lithosphere *subsidence* was replaced by lithospheric *subduction*. Application of the subduction-zone infancy model to the Late Jurassic ophiolites of California provides a framework in which to understand the rapid formation of oceanic crust with strong arc affinities between the younger Sierran magmatic arc and the Franciscan subduction complex, provides a mechanism for the formation and subsidence of the Great Valley forearc basin, and explains the limited duration of high-T, high-P metamorphism experienced by Franciscan mélanges.

INTRODUCTION

Tectonic and magmatic processes occurring at convergent margins are important for the evolution of the Earth's crust and mantle. Convergent margins are sites where batholiths form, much ore formation occurs, where continental crust is generated, and where the Earth's crust is distilled, with the residue being recycled back into the mantle. Consequently, an impressive and growing literature focuses on the ancient and modern igneous products of both continental-margin (Andean-type) and intra-oceanic arc systems. In spite of this, there has been little discussion of the implications of the igneous activity that accompanies the earliest stages (infancy) of subduction zone evolution and the implications of changing styles of igneous activity for our understanding of convergent margin evolution. In this paper, we offer some new observations regarding the composition, volume, and intensity of igneous activity associated with the first few million years in the evolution of a subduction zone. We first summarize and further interpret studies of the early history of the Izu-Bonin-Mariana (IBM) arc system. We use these observations to develop a general model for the earliest stages in the evolution of arcs, a stage we term "subduction-zone infancy." This model attempts to explain several features inferred for infant-arc systems, including (1) an unusually broad zone of volcanism, (2) high magma production and eruption rates, (3) a strongly extensional tectonic environment, and (4) progressive migration and focusing of the magmatic front away from the trench. Finally, we use this model to reinterpret the evolution of the Late Jurassic ophiolites of northern California as an example of subduction-zone infancy.

INFANCY OF THE MARIANA-BONIN ARC SYSTEM

The early history of the IBM system (Fig. 1) is best known among intraoceanic arcs; this is because of the youth, lack of subsequent deformation, and thinness of sediment cover of this arc system. This system is interpreted to have formed when an oceanic transform fault was converted into a subduction zone. (We note that competing models exist [for example, Seno and Maruyama, 1984], but this and related models are not preferred here because we believe these are arbitrary and unnecessarily complex.) A long transform fault connected a spreading ridge to the southwest (probably an extension of a ridge between Asia and Australia active during the Mesozoic and Paleogene; Hilde and others, 1977) with the Kula-Pacific ridge to the northeast (Fig. 2A). Following subduction of the Kula-Pacific ridge beneath Japan during the Late Cretaceous (Uyeda and Ben-Avraham, 1972), the transform separated old oceanic lithosphere to the east (Jurassic–Early Cretaceous) from young crust, still being produced at the ridge to the west (Fig. 2B). The transform fault was a zone of weakness that disrupted the entire lithosphere; the difference in age resulted in the sea floor being 2–3 km deeper to the east, providing a crustal environment favorable for the development of a west-dipping subduction zone. The change in motion of the Pacific Plate, from north-northwest to west-northwest, occurred at 43.1 ± 1.4 Ma (Clague and Dalrymple, 1987). The conclusion that the IBM arc system (Fig. 1) formed shortly before, or at, this time is based on paleontologic and K-Ar geochronologic data, from both subaerial exposures (Cloud and others, 1956; Tracey and others, 1964; Kaneoka and others, 1970; Hanzawa, 1974; Meijer and others, 1983; Umino, 1985) and submarine drill core (Ellis, 1981; Kling, 1981; Fryer and others, 1990). It is important to note that there is evi-

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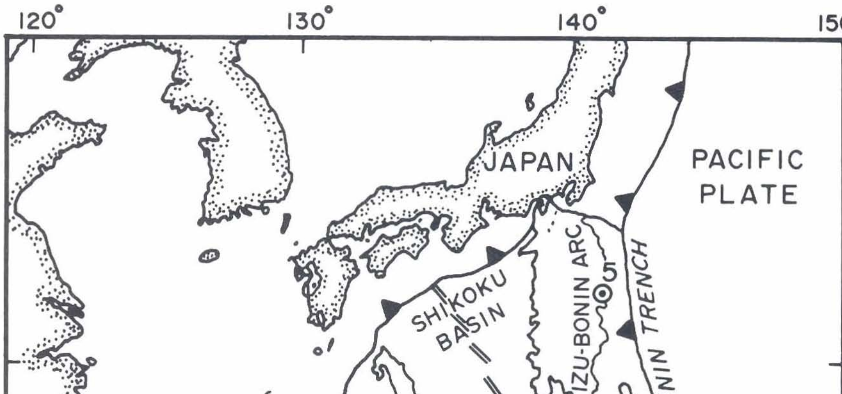

Figure 1. Tectonic and physiographic map of the Philippine Sea, Mariana Arc, and surrounding areas, modified from Taylor and others (1990). Bathymetric contours are at 3,000 m for the Izu-Bonin-Mariana arc and 4,000 m elsewhere. Double lines represent active and extinct spreading ridges. Subduction zones have teeth on overriding plate. Dotted line (A–A′) shows the location of profile shown in Figure 3. Circled dots refer to drill sites discussed in text: 1 = DSDP 459; 2 = DSDP 458; 3 = DSDP 451; 4 = DSDP 448; 5 = ODP 786; 6 = Ogasawara Paleoland. G = Guam; S = Saipan; CBR = Central Basin Ridge; M = Mindanao; L = Luzon; T = Taiwan.

dence for arc volcanism prior to the change in plate motion, as early as 48 Ma (Maruyama and Seno, 1984; Taylor, 1992).

Following generalizations about the origin of oceanic crust beneath fore arcs (Dickinson and Seely, 1979), Casey and Dewey (1984) argued that the Mariana-Bonin fore arc was composed of trapped oceanic crust. They noted that the magnetic anomalies of the West Philippine Basin trend perpendicular to the Mariana Trench and predicted that a similar orientation and age progression would be found in the IBM fore arc. This is refuted by the data available from the fore arc. First, the trends of dikes exposed on Guam are variable (Tracey and others, 1964), inconsistent with the spreading direction identified for the West Philippine Basin (Hilde and Lee, 1984). Second, the oldest crust, everywhere that it has been sampled along 2,200 km of the arc, is middle to late Eocene in age. If the fore arc were composed of trapped oceanic crust, then it should have very different ages along strike. For example Shih (1980) interpreted the magnetic anomalies of the West Philippine Sea to indicate a spreading half rate of 5 cm/yr between 59 and 41 Ma; if the fore-arc crust were produced at such a ridge, an age difference of at least 22×10^6 yr would be expected, much less than the <5–10 Ma differences that are observed. Finally, and most importantly, the composition of igneous rocks from the fore arc are those appropriate to convergent margins: low-Ti arc tholeiite, boninite, andesite, and dacite (Meijer, 1983; Natland and Tarney, 1981; Hickey and Frey, 1982; Reagan and Meijer, 1984; Bloomer and Hawkins, 1987; Ishii, 1985). Gabbroic and ultramafic rocks also have strong arc affinities, including abundant Mg-orthopyroxene and Cr-rich spinel (Bloomer and Hawkins, 1987). The only exceptions are MORB (mid–oceanic-ridge basalt) and OIB (oceanic island basalt) lavas that occur with Cretaceous microfossil assemblages (for example, Bloomer, 1983; Johnson and others, 1991; Taira and Pickering, 1991); to date, no MORB or alkalic gabbro or ultramafic rocks have been reported. Sampled sediments are too old to represent trapped Philippine Sea crust, and Johnson and others (1991) infer that these represent accreted fragments. These arguments lead to the conclusion that the crust beneath the IBM fore arc must be overwhelmingly dominated by crust that formed in an arc setting.

To further examine the evolution of the Mariana arc system during the Tertiary, it is necessary to palinspastically restore the remnant arcs of the Palau-Kyushu Ridge and the West Mariana Ridge (Fig. 1). These were rifted away during opening of the Parece Vela-Shikoku and Mariana Trough back-arc basins, about 30 to 15 m.y. ago (Taylor, 1992). With these reconstructions (Figs. 3A and 3B), we can discuss the extent of igneous activity through the arc system during the Tertiary period (Fig. 4). Two points need to be stressed. First, late Eocene and earliest Oligocene volcanic rocks are the youngest igneous rocks encountered in the region between the frontal arc and the trench. Although Pleistocene lavas were recovered on Ocean Drilling Project (ODP) Leg 125 (Marlow and others, 1992), these are unusual. Second, the magmatic axis of the arc became concentrated along the modern magmatic arc (active arc) sometime during the Oligocene. A similar migration is documented for the Bonin arc (Taylor, 1992), although the migration of the magmatic arc away from the trench continued until the latest Oligocene (Honza and Tamaki, 1985). Retreat of the arc magmatic front is a generally observed phenomenon (Dickinson, 1973), and understanding its cause is critical to understanding arc evolution.

The area underlain by late Eocene arc crust is unusually broad. A minimum breadth of about

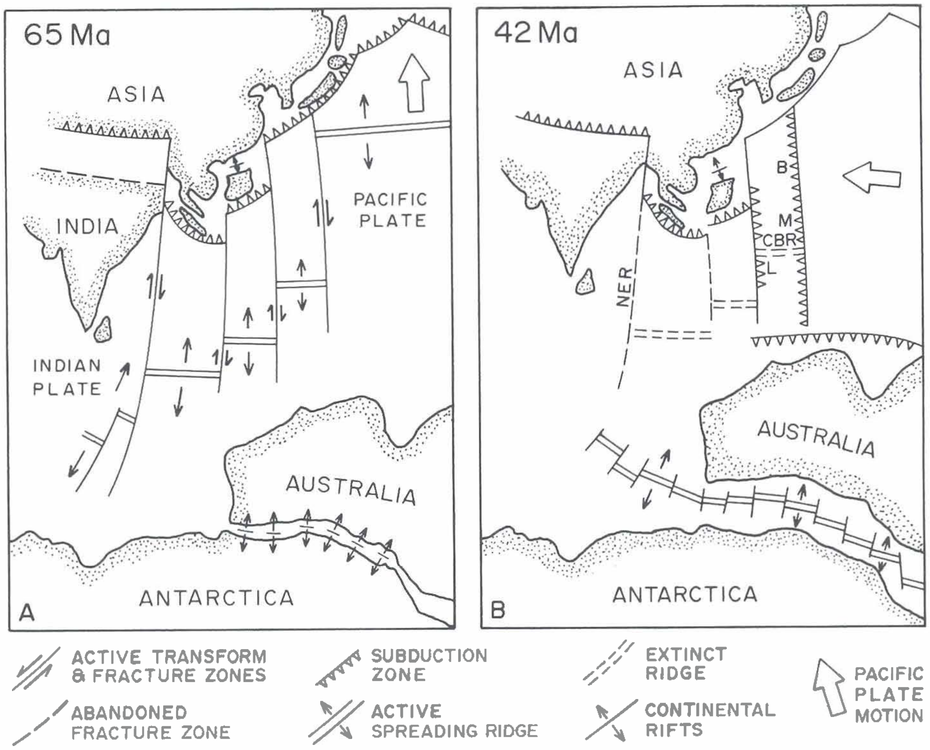

Figure 2. Tectonic cartoon illustrating the transformation of north-south transforms into subduction zones during late Eocene time, ca. 42 Ma. The figure illustrates the formation of the Mariana (M) and Bonin (B) arcs as the transform on the east side of the Philippine Sea became the site of a new subduction zone. Note that Zambales, Luzon (L) is interpreted as the result of a similar transformation on the west side of the Philippine Sea Plate. Modified after Hilde and others (1977). NER = Ninety East Ridge; CBR = Central Basin Ridge.

200 km is inferred for the Izu-Bonin fore arc (Fryer and others, 1990). A similar distance is obtained from the distance between the Mariana Trench and the frontal arc islands of Guam and Saipan; this is a minimum because we do not know how much material has been removed by tectonic erosion (Bloomer, 1983; Johnson and Fryer, 1990; Johnson and others, 1991), nor do we know the exact locus of Eocene volcanism, as some of this crust underlies the Palau-Kyushu Ridge (Ozima and others, 1977). Taylor (1992) argued that this zone was more than 300 km and as much as 450 km wide; this estimate is supported by sedimentary relationships that indicate a volcanic source to the west of the present Mariana frontal arc during the late Eocene (Tracey and others, 1964; Garrison and others, 1975; Karig and Rankin, 1983). Some Oligocene extension has been documented in the northern IBM fore arc (Taylor, 1992); however, closing this still leaves a forearc of >200-km width. In spite of the uncertainties in the width of the Eocene magmatic zone, this was much broader than for modern arc systems. For example, the width of the Mariana arc presently affected by arc volcanism is no more than 30–50 km (Bloomer and others, 1989). Much of this includes volcaniclastic aprons, so that the true width where magmas may transit or erupt may be significantly less, although rare igneous activity away from the arc axis is recorded (Marlow and others, 1992).

The width of the late Eocene volcanic zone indicates that IBM infant arc volcanism was unusually intense for an arc. This is further demonstrated by the abundance of boninite and related cumulates and residues. Along with depleted arc tholeiites, boninitic assemblages of demonstrated, or presumed, late Eocene age characterize the basement of the inner trench slope of the IBM system in most places where it has been sampled, from the southern Marianas (12°N; Bloomer, 1983; Dietrich and others, 1978) and the central Marianas (18°N; Hickey and Frey, 1982; Bloomer, 1983) to the Ogasawara Paleoland (26°N, Ishii, 1985; 26°54′N, Tararin and others, 1987) and the Izu-Bonin fore arc (32°N, Pearce and others, 1992). The generation of boninite involves melting of harzburgite (OL + OPX peridotite), ultradepleted rocks not melted in any other modern tectonic setting. The oceanic lithosphere is expected to be most depleted just beneath the Moho and become less depleted with depth (Scott and Stevenson, 1989). This is consistent with experimental results indicating that melting to form boninite occurs in the uppermost mantle (ca. 1,200 °C and 3–4 kbars, van der Laan and others, 1989) and further implies that melting of less-depleted peridotite at greater depth to form tholeiite would also be likely. The ubiquitous but temporally restricted occurrence of boninite indicates that the earliest stages in the evolution of the IBM arc were characterized by unusually intense and widespread mantle melting.

An estimate of the volume of crust produced during the late Eocene infancy of the Marianas requires some knowledge of crustal thicknesses. We have already estimated a minimum width of 200 to 300 km for the infant arc; determining the thickness of that crust is more difficult. Because there appears to be little but younger sediments over the late Eocene basement (266 m at Deep Sea Drilling Project [DSDP] site 458, 559 m at DSDP site 459B), we infer that all of this crust must be either infant-arc crust or trapped oceanic crust. There is little evidence from frontal arc exposures, DSDP and ODP drilling, and several dredging samplings to support the hypothesis that the fore arc is composed of oceanic crust that was trapped during the transition of the inferred transform fault into a subduction zone. The only fragments of non-arc crust in the outer fore arc occur as inferred thrust slices (Bloomer, 1983; Johnson and Fryer, 1990; Johnson and others, 1991). Because of these arguments, and because there is no evidence to the contrary, we conclude that nearly all of the crust from the trench to the frontal arc is made of infant-arc igneous rocks. Seismic refraction data (Murauchi and others, 1968) indicate that the IBM fore-arc crust at 23°N is 10–12 km thick. More recent refraction studies (LaTraille and Hussong, 1980) indicate that the crust of the fore arc attains velocities of 6.1 to 6.5 km/sec at depths of about 7 km and so must be thicker than this. Sager (1980) used these results to constrain gravity modeling. His models indicate that the fore-arc crust thickens from about 5 km at the trench to more than 15 km beneath the frontal arc. Seismic refraction studies and gravity modeling of the Izu-Bonin fore arc indicates that the crust is 12–17 km thick (Honza and Tamaki, 1985; Horine and other 1990).

The Zambales ophiolite of Luzon is a good approximation to the structure and composition of the deep IBM arc crust (Pearce and others, 1992); this is a "supra-subduction zone (SSZ) ophiolite" (as defined by Pearce and others,

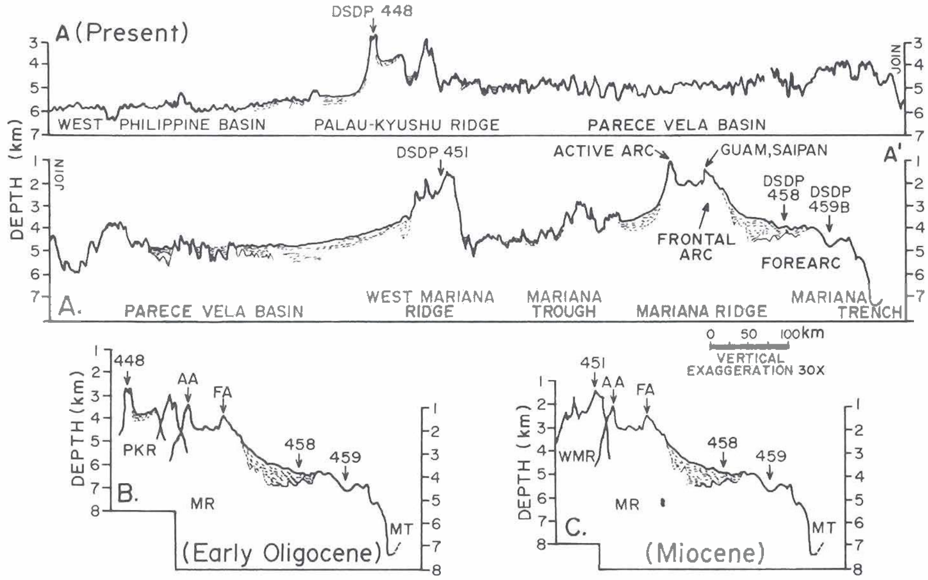

Figure 3. (A) Bathymetric profile through the Mariana Trench, fore arc, and arc; Mariana Trough (active back-arc basin); West Mariana Ridge (remnant arc); Parece Vela Basin (extinct back-arc basin); Palau-Kyushu Ridge (remnant arc); and West Philippine Basin (trapped oceanic crust), from Karig (1971). Location of profile is shown as dotted line in Figure 1; vertical lines mark bends in profile. Projected locations of DSDP sample sites and islands are also shown. (B) Reconstruction of Palau-Kyushu Ridge (PKR) and Mariana Ridge (MR) and Trench (MT), approximating the situation prior to the opening of the Parece Vela Basin during the Oligocene epoch. (C) Reconstruction of the West Mariana Ridge (WMR) with the MR and MT, approximating the situation prior to the opening of the Mariana Trough in the Mio-Pliocene. AA = active arc; FA = frontal arc. Note that Eocene volcanic rocks and associated plutonic rocks are known from the frontal arc as far east as DSDP site 459; basement exposed in the trench wall is probably similar in age.

1984) composed of both boninitic and tholeiitic components (Hawkins and Evans, 1983; Geary and others, 1989). It is also late Eocene in age and can be interpreted as having formed during the initial stages of subduction along the Manila Trench, during the major plate reorganization that led to the formation of the IBM subduction zones (Fig. 2). Estimated crustal thicknesses are 5 km for the Coto Block and 8 km for the Acoje Block (Hawkins and Evans, 1983). We thus estimate a minimum crustal thickness of 6 km for the infant IBM arc, a typical ocenaic crustal thickness, although the infant-arc crust may have been significantly thicker beneath what is now the frontal arc.

These parameters indicate that between 1,200 and 1,800 km^3 of fore-arc crust was produced per kilometer of IBM arc during late Eocene time. We know that the system was active by 45 Ma. Because it is not clear whether or not the volcanic episode continued into early Oligocene time, we allow 10 m.y. for the episode, yielding a crust production rate of 120 to 180 km^3/km-Ma. The inferred crustal production rate for the infant IBM arc is much greater than that inferred for modern arcs, none of which is in its infancy (Fig. 5). Modern arcs have mean eruption rates of 13 km^3/km-Ma (Gill, 1981; Sample and Karig, 1982; Wadge, 1984). Kay and Kay (1985) inferred that the ratio of eruption to plutonic emplacement for the Aleutian arc is between 0.69 and 0.81; using a mean of 0.75; this yields a mean crustal production rate of 30 km^3/km-Ma, a production rate identical to the "average arc accretion rate" of Reymer and Schubert (1984).

The inferred crustal production rate for the infant IBM arc is the same as that of a mid-oceanic ridge spreading at a half rate of 2–3 cm/yr. These rates, patterns of faulting in the fore arc, and the crustal thicknesses support the suggestion that IBM infant-arc crustal formation occurred in a strongly extensional environment. Poor exposure hinders identification of sheeted dikes, but Eocene boninitic sheeted dikes occur on Chichi-jima in the Bonin Islands (Arculus, 1992, personal commun.). The inferences from the geology—high crustal production rates, evidence of crustal extension, and association with thin lithosphere—are consistent with crustal formation mechanisms similar in style to those of sea-floor spreading centers. This is the "pre-arc spreading" stage inferred for SSZ ophiolites (Pearce and others, 1984). This conclusion is consistent with the observations of Bloomer (1987) that the association of peridotite, gabbro, boninite, and arc tholeiite recovered from the inner trench wall represents an autochthonous ophiolite with arc affinities, and with the observations of other investigators who have suggested that the IBM fore-arc crust formed quickly, during the earliest stages in the evolution of the subduction zone (for example, Kobayashi, 1983; Natland and Tarney, 1981; Meijer, 1980; Taylor, 1992; Pearce and others, 1992). We thus concur with Natland and Tarney (1981, p. 895) who argued ". . . that at the earliest stages of arc volcanism, a type of sea-floor spreading may have occurred before volcanism focussed along the line of the arc."

MODEL FOR INFANT-ARC CRUST FORMATION: INFERENCES FROM IBM ARC INFANCY

We conclude that the earliest stages of IBM subduction were strongly extensional, in many ways indistinguishable from that of sea-floor spreading. This contrasts with many models that instead predict strong contraction during subduction initiation (McKenzie, 1977; Cloetingh and others, 1989; Mueller and Phillips, 1991). In this section, we attempt to explain the causes of this extension, drawing heavily on the thoughts of Donnelly and Rogers (1980), Natland and Tarney (1981), Kobayashi (1983), Hawkins and others (1984), Pearce and others (1984), and Leitch (1984). We suggest that sea-floor spreading occurs in the evolution of the fore arc and then speculate on why spreading stops, marking the end of arc infancy.

It is generally recognized that the beginning of subduction requires overcoming a significant stress threshold (for example, McKenzie, 1977). These stresses can be overcome most readily along pre-existing zones of lithospheric weakness, the most important of which are transform faults and fracture zones. A convincing case that these are generally sites of subduction initiation has been made by Casey and Dewey (1984) and Karig (1982) and specifically for the IBM system by Uyeda and Ben-Avraham (1972). It is essential to know whether subduction began because of regional forcing due to change of Pacific plate motion at 42 Ma or due to local gravitational instabilities developed across the precursor fracture zone. In the first instance, regional forces could result in strongly compressional stresses in the initial subduction zone, whereas in the second case, these stresses would

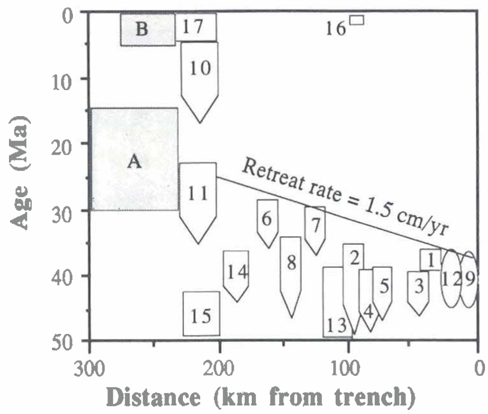

Figure 4. Summary of the spatial distribution of volcanism in the Mariana-Bonin arc system. Eocene volcanic rocks are known from DSDP sites 458 and 459B in the fore arc and from the frontal arc islands of Guam and Saipan; volcanism continued into the Oligocene epoch on Guam. Although Eocene rocks were not recovered from DSDP sites 448 and 451 on the Palau-Kyushu and West Mariana Ridges, rocks of this age have been dredged from the northernmost Palau-Kyushu Ridge (Ozima and others, 1977). Note that Eocene volcanic rocks erupted from at least the frontal arc to the inner trench wall, a distance of about 200 km. Note also that arc lavas of Oligocene and younger age were erupted from volcanoes near the present location of the active arc. Key: 1 = Mariana forearc (Johnson and others, 1991); 2 = DSDP site 458; 3 = DSDP site 459; 4 = ODP site 782; 5 = ODP site 786; 6 = ODP site 792; 7 = ODP site 793; 8 = Guam; 9 = Inner wall, Mariana Trench; 10 = West Mariana Ridge (DSDP site 451); 11 = Palau-Kyushu Ridge (DSDP site 448); 12 = Ogasawara Paleoland (Ishii, 1985); 13 = Bonin Islands; 14 = Saipan; 15 = Northern Palau-Kyushu Ridge (Ozima and others, 1977); 16 = ODP site 781 (Marlow and others, 1992); 17 = Mariana active arc; A = Parece Vela basin; B = Mariana Trough. Rectangles refer to areas where good age constraints exist. Pointed bottoms on some refer to the likelihood that older crustal sections exist. Ellipses refer to crustal sections with age assignments based on the abundance of boninitic rocks, inferred to be late Eocene in age. $^{40}Ar/^{39}Ar$ ages for DSDP site 448 are from Sutter and Snee (1980); biostratigraphic control for DSDP sites 448 and 451 is from Martini and others (1980). Biostratigraphic control for DSDP sites 458 and 459B is from Ellis (1981) and Kling (1981). K-Ar ages for Guam and Saipan are from Meijer and others (1983). Biostratigraphic control for Guam and Saipan is from Tracey and others (1964), Cloud and others (1956), and Meijer and others (1983). Biostratigraphic and geochronologic control for the Bonin Islands is from Hanzawa (1974) and Kaneoka and others (1970).

more likely be strongly extensional. Evidence that arc igneous activity began before 42 Ma (Ozima and others, 1977; Seno and Maruyama, 1984; Taylor, 1992) indicates that the change of Pacific plate motion could not have induced subduction initiation in the case of the IBM arc.

The age of the lithosphere at subduction zones is very important in controlling the motions of oceanic plates (for example, Elsasser, 1971; Forsyth and Uyeda, 1975; Carlson and others, 1983; Molnar and Atwater, 1978; Bercovici and others, 1989). New oceanic lithosphere is gravitationally stable because its density is dominated by that of the crust, which is less dense than the asthenosphere. Because the thickness of lithospheric mantle (which, because it is colder, is denser than asthenosphere) varies as the square root of its age (Parker and Oldenburgh, 1973), the lithosphere eventually becomes gravitationally unstable (Oxburgh and Parmentier, 1977). Molnar and Atwater (1978) calculated that this occurs when the lithosphere is 30×10^6 yr old. By this reasoning, the lithosphere east of the transform fault would have been gravitationally unstable prior to the initiation of subduction in the IBM system. At this time, the lithosphere to the west was very young, 0–20 m.y. old at most (Shih, 1980; Hilde and Lee, 1984). The lithosphere to the east was older, although it is difficult to specify how much. Jurassic magnetic lineations have been interpreted on the sea floor east of the IBM arc (Nakanishi and others, 1989; Larson and Chase, 1972), and this has been confirmed by the drilling of middle Jurassic sedimentary rocks (ca. 170 Ma) at ODP site 801C (Shipboard Scientific Party, 1990). Interpretation of spreading geometries from the anomaly patterns suggests that younger crust, perhaps early- to mid-Cretaceous (ca. 125 Ma) existed east of the transform fault when subduction began. This age inference finds further support in the age of accreted fragments preserved in the fore arc, previously discussed as indicating the accretion of Cretaceous sea floor during the Paleogene. Thus, the lithosphere to the east of the IBM system would have been about 65 or more m.y. old at the time of subduction initiation and so would have been gravitationally unstable (Fig. 6A).

Recognition of the "trench rollback" process indicates that an important component of the motion of subducted lithosphere is to fall vertically through the asthenosphere (for example, Elsasser, 1971; Garfunkel and others, 1986). Gurnis and Hager (1988) recognized that younger subduction zones have the steepest dips, whereas Davies (1980) calculated that gravitationally induced sinking forces become increasingly important from the surface to a maximum about 150-km depth. These results are in accord with earthquake focal mechanisms, indicating that extension in the subducted lithosphere is generally limited to depths of <200 km (Isacks and Molnar, 1971). These lines of evidence indicate that the maximum density contrast between asthenosphere and old lithosphere is present in the upper 100–200 km of the mantle. We thus suggest that subduction along the precursor transform fault was initiated by vertical "falling" of the gravitationally unstable lithosphere to the east. This inference agrees with the results of numerical simulations by Tomoda and others (1985). Note that these results (Fig. 7) indicate that the gravitational instability results in accelerated subsidence of the old lithosphere near the transform.

McKenzie (1977) argued that two stresses must be overcome for subduction to begin. The first is the shear stress on the fault plane of the embryonic subduction zone, and the second is the stress needed to bend the plate. Because the proto-trench forms along the line of a pre-existing fracture zone, the first stress may be negligible. Plate bending, however, may not be ignored. We assume for the purposes of developing the model that a pre-existing lithospheric weakness reduced the force required to accomplish plate bending, perhaps by the existence of another fracture zone or a region weakened by hot-spot igneous activity. The observations indicate that subduction initiation did occur; ergo, the plate-bending stresses were overcome.

As the old lithosphere subsides, the maximum displacement would be closest to the fault, with a hinge farther east (Fig. 6B). The downward displacement would be limited by rate of displacement of the underlying asthenosphere (Garfunkel and others, 1986). Substantial relief (~2–3 km; Parsons and Sclater, 1977) would initially exist across the precursor transform fault. As shown in Figure 7, subsidence-induced displacement of asthenosphere leads to extension of the young lithosphere, with young lithosphere eventually overriding old lithosphere. This is critical, because subduction can only be successful if asthenosphere overrides lithosphere (Turcotte and others, 1977). Once established, this will accelerate because it permits the transfer of lower-density asthenosphere from beneath, to above, the sinking lithosphere. Extension would

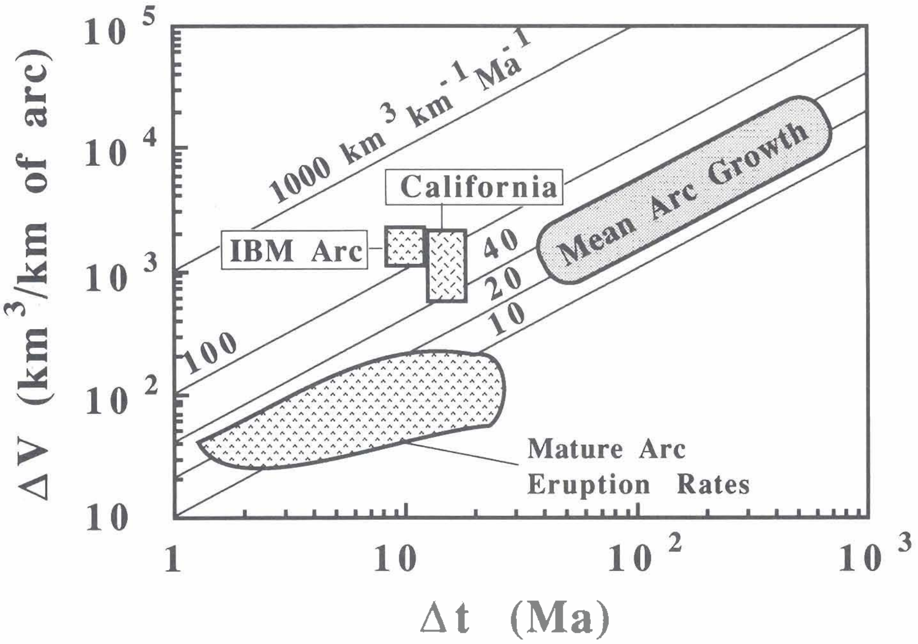

Figure 5. Diagram illustrating crustal production rates for the Eocene IBM and Jurassic of California after Reymer and Schubert (1984), along with their estimate for mean arc growth. The assumptions involved in calculating a mean crustal production rate of 120 to 180 km^3/km-Ma are discussed in the text. Field labeled "Mature Arc Eruption Rates" encompasses estimates from various sources (Gill, 1981; Sample and Karig, 1982; Wadge, 1984). Using Gill's (1981) compilation plus the estimates of Sample and Karig (1982), we calculate a mean arc eruption rate of 13 km^3/km-Ma. Using this estimate and the proportion for arcs of volcanic to plutonic additions of 0.75 (Kay and Kay, 1985), we obtain a mean crustal growth rate for mature arcs of 30 km^3/km-Ma. For the Jurassic of California, the lower estimate (60 km^3/km-Ma) is based on present width of the infant arc, from the eastern edge of the Western Belt of the Sierra Nevada to the Coast Range thrust, along 39°15′N. Upper estimate (130 km^3/km-Ma) includes 155 km of estimated shortening (Suppe, 1978). Duration of infant-arc episode is based on U-Pb zircon ages for Coast Range, Western Klamath, and Smartville ophiolites, 153–166 Ma. Volume calculations are based on a 6-km crustal thickness.

be accomplished first by stretching of young lithosphere, then by sea-floor spreading. Soon an equilibrium would be set up such that the subsidence of the lithosphere displaced a volume of asthenosphere matched by a similar volume of extension (Fig. 6B). At this point, the rate of lithospheric subsidence would accelerate and be limited by the rate of asthenospheric transfer from beneath, to above, the subsiding lithosphere. We expect that superposition of asthenosphere over lithosphere would first occur at one point along the proto-trench and then propagate rapidly north and south; this effect is consistent with the slightly greater age of arc infancy in the northern part of the IBM system.

Migration of asthenosphere over the subsiding lithosphere would entail adiabatic decompression. Simultaneously, widespread dehydration of the subsiding lithosphere would occur as especially its upper portions of altered basalt and sediments were squeezed and heated. Most of the water is released from subducted lithosphere and sediments by 40-km depth (Peacock, 1990), and this water would be introduced into the overlying asthenosphere. The two effects—adiabatic decompression (Klein and Langmuir, 1987) and hydration (Davies and Bickle, 1991)—should lead to extensive melting. Melting in such an extensional environment would lead to infant-arc crust formation by sea-floor spreading. This spreading would in many ways be similar to that of mid-ocean ridge and back-arc basin spreading regimes, with two significant differences. First, spreading is likely to be poorly organized, with many discrete ridge segments and asymmetric spreading. The infant-arc crust nearest the precursor transform fault might be first to form, although this may be replaced by younger infant-arc crust during later rifting events. Infant-arc crust and lithosphere in this region should be found in contact with the older oceanic crust and lithosphere farther away from the trench across what is essentially an intrusive contact, although in reality this contact would probably be so intensely metamorphosed, deformed, and intruded as to not easily be easily recognizable in the field (Fig. 6C). The second way in which infant-arc sea-floor spreading probably differs from mid-ocean-ridge spreading is the composition of lavas produced. IBM infant-arc lavas and residues show evidence for greater degrees of depletion than do MORB and their residues (Bloomer, 1983; Bloomer and Hawkins, 1987; Pearce and others, 1992). Boninites, as harzburgite melts, are the most spectacular examples of this. Melting of harzburgite does not happen at mid-ocean ridges; MORBs form by melting of much-less-depleted lherzolite. Whereas boninites are associated with arc tholeiites, their mere existence and restriction to the earliest stage of IBM arc formation indicates that the extremely unusual event of harzburgitic melting is restricted to infant-arc magmagenesis.

Spinel compositions in peridotite and associated lavas further indicates that this melting is unusually extensive. Dick and Bullen (1984) argued that the Cr# (Cr/Cr + Al) of spinels covaries with the degree of depletion of the mantle source. They noted that MORB-type abyssal peridotites have Cr# < 0.60, consistent with the inference that melting to form MORB (and back-arc basin basalts) was limited by exhaustion of clinopyroxene. Spinels from Eocene IBM arc lavas and peridotites are commonly much more Cr-rich (Cr# > 0.60) and, along with inferences from Ca contents of orthopyroxene in such peridotites, indicates that substantial melting continued after clinopyroxene exhaustion. The extreme depletion indicated may have two causes in the model advanced in Fig. 6B. First, the combined effects of adiabatic decompression of asthenosphere plus the lowering of the peridotite solidus by hundreds of degrees Celsius due to the influx of water leads to an extreme degree of melting only obtained beneath the infant arc. Second, the pre-existing and already depleted lithosphere mantle may be melted in the initial phases of subduction.

The end of arc infancy is marked by stabilization of the arc volcanic axis near the present magmatic front and by the cessation of boninitic igneous activity. This happened by the mid-Oligocene in the IBM system. This change re-

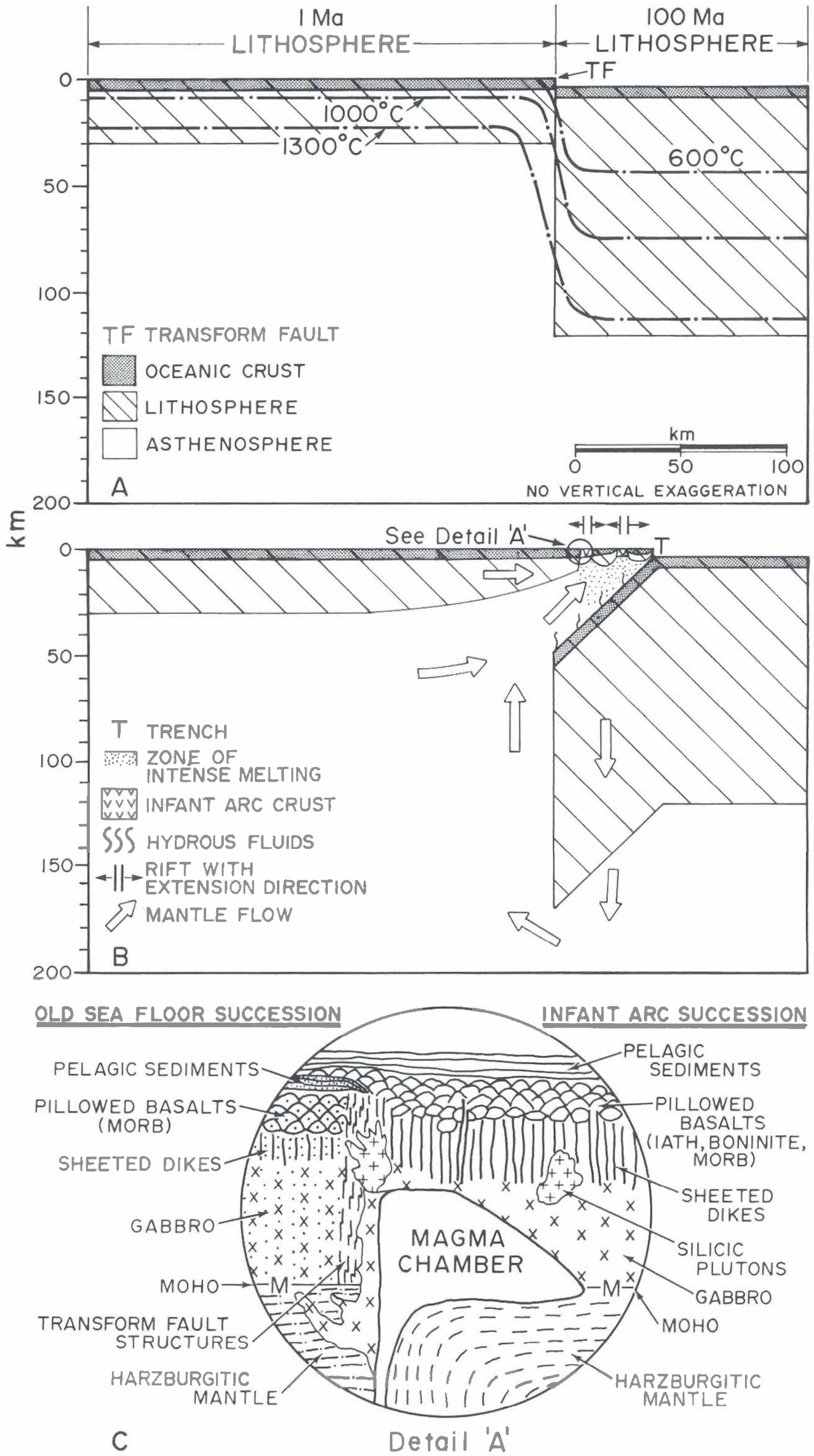

Figure 6. Section perpendicular to the transform fault/trench through the crust and upper mantle just prior to (A), and just after (B), initiation of subduction. C illustrates geologic relations that may be found at the junction between older oceanic and infant-arc crust and lithosphere.

flects a fundamental change in the thermal structure and in the extensional stresses experienced by the fore-arc wedge as the slab deepened. Cooling of the IBM fore arc is difficult to date, but some constraints come from the recent recognition of serpentinite diapirs. These diapirs form when lithospheric mantle is hydrated by water given off from the subducted lithosphere and are found in the IBM fore arc as far as 120 km away from the trench axis (Fryer and Fryer, 1987). ODP drilling results indicate that the first appearance of the diapirs was pre–Early Pliocene in the Mariana fore arc and pre–Middle Miocene in the Bonin fore arc (Fryer and others, 1990). Fryer and others (1985) inferred from this and other studies that the fore-arc thermal gradient is now 10–15 °C/km, an order of magnitude less than the 100–200 °C/km inferred for arc infancy.

The change in extensional stress in the fore arc probably reflects the change from lithospheric *subsidence* to lithospheric *subduction*. The reasons for this change may be due to the increasing difficulty of asthenospheric transfer from beneath the subsiding lithosphere. For example, for the lithosphere to subside from A to A′, the area AA′BB′ of asthenosphere must be removed (Fig. 8A). With continuing retreat of the hinge, both the area swept out by the subsiding lithosphere and the mean path of asthenospheric migration from beneath to above the subsiding lithosphere increase rapidly; these effects combine to increasingly slow subsidence. At some depth, basaltic crust transforms to eclogite, which will greatly increase the mass excess of the subsiding lithosphere. Ahrens and Schubert (1975) concluded that eclogite becomes stable at depths of not more than 20–30 km but further noted that the phase transition is limited by the reaction rate. The high temperatures in the surrounding mantle and high water contents of the subsiding lithosphere should cause this transition to occur at less than the 50 km depth observed in mature subduction zones such as the southwest Japan arc (Fukao and others, 1983). Eventually, the mass contrast between the subsiding lithosphere and the surrounding asthenosphere may lead to down-dip as well as purely vertical motion; when this happens, true subduction begins. The transition to subduction leads to the termination of fore-arc extension and magmatism (Fig. 8B). Extensional stresses will persist because of continued trench rollback, but rollback rates accompanying subduction are 10%–20% of total convergence rate (Garfunkel and others, 1986). This contrasts with the situation for lithospheric subsidence during subduc-

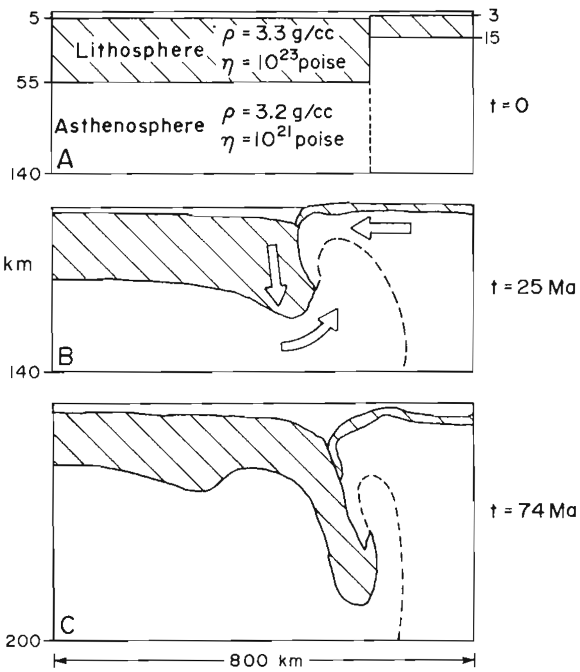

Figure 7. Diagram summarizing results from the numerical simulation of gravitational body forces acting on old lithosphere separated from young lithosphere by a transform fault; from Tomoda and others (1985). Note that the only differences between the lithosphere and asthenosphere are density and viscosity. All boundaries are stress-free and free-slip. Isostasy is assumed on both sides of the fracture zone. Note that the downward projection of the transform through the asthenosphere is a dotted line. (A) Initial conditions. (B) Configuration after 25 Ma. Note the trajectory of old lithosphere downward and the transfer of its underlying asthenosphere up and to the right. This also leads to extension of the young lithosphere. (C) Configuration after 74 Ma. Note that extension in the young lithosphere has ceased as transfer of asthenosphere from left to right is hindered by the descent of the old lithosphere.

tion zone infancy, where the rollback rate approximates the subsidence rate. This decrease in rollback rate results in a corresponding decrease in the extensional stresses across the fore arc and a greatly diminished volume of asthenosphere advected beneath the fore arc, which in turn leads to cooling of the fore arc. Cooling of the fore arc results from heat losses due to advection in the form of volcanism and hydrous fluid infiltration and conductive cooling by the slab (Andrews and Sleep, 1974; Davies and Bickle, 1991). This cooling results in the formation of thick lithosphere beneath the fore arc, such that plate rupture and asthenospheric upwelling will subsequently occur along the line of the magmatic arc (Fig. 8B). Subsequent tensional stresses will be accommodated by back-arc basin extension, demonstrated for the IBM system by the last major fore-arc rifting event in early Oligocene time being followed by the first back-arc basin rifting during late Oligocene time (Taylor, 1992).

The cessation of fore-arc spreading and transition from lithospheric subsidence to subduction marks the end of arc infancy. After the axis of extension is localized within, and behind, the volcanic arc, and the fore arc develops a subnormal geotherm, the system has achieved a thermal equilibrium that will change little over the lifetime of the arc. The location of the volcanic axis will be determined by the thickness of the lithosphere. This may thicken with time but induced convection in the asthenosphere may inhibit conductive cooling and thickening, such that retreat of the magmatic axis away from the trench may be relatively slow (Dickinson, 1973).

LATE JURASSIC CRUST FORMATION IN WESTERN CALIFORNIA: AN APPLICATION OF THE INFANT-ARC MODEL

There are many ophiolitic successions where the infant-arc model can be applied—including Troodos, Semail, Papua New Guinea, Zambales, Newfoundland, and the ophiolites of northern California. Because its tectonic setting is still controversial, the latter is especially appropriate for re-examination using the infant-arc paradigm. We discuss here the three principal components of what we interpret as a Late Jurassic infant-arc complex, including the Coast Range ophiolite, the Smartville ophiolite, and the Western Jurassic belt of the Klamaths (Fig. 9). These are the best preserved of an extensive band of mid- to Late Jurassic ophiolitic terranes that line the eastern Pacific margin, stretching at least from Baja California (Kimbrough, 1985) to Washington State (Whetten and others, 1980).

There are many models for the origin of the Coast Range ophiolite (for example, Hopson and others, 1981; Robertson, 1989; Shervais, 1990); for the Western Belt of the northern Sierra Nevada, the most important component of which is the Smartville Complex (for example, Xenophontos and Bond, 1978; Beard and Day, 1987; Edelman and others, 1989a); and for the Josephine ophiolite and related arc sequences of the Western Jurassic Belt of the Klamath Mountains (for example, Garcia, 1979; Harper and Wright, 1984; Harper, 1984; Wyld and Wright, 1988). Saleeby (1981) first argued that all are parts of one mid- to Late Jurassic convergent margin that developed on the fringes of North America. We argue in the following section that all of these components represent a broad and regionally extensive belt of infant arc that formed adjacent to the western margin of North America and that was subsequently buried beneath a prograding clastic wedge shed from the locus of active-arc magmatic activity when this migrated east during the latest Jurassic. We develop this argument first for the Klamaths, then for Smartville, and finally for the Coast Range ophiolite.

Arc Infancy in the Western Jurassic Belt of the Klamaths

The Western Jurassic Belt of the Klamath Mountains includes arc and ophiolitic crustal components that record extension, clastic sedimentation, and imbrication beneath older Klamath terranes, all occurring during Late Jurassic time. Consensus exists that this crust formed via extension adjacent to North America. Arguments for continental proximity include the following. (1) The 159 Ma Preston Peak ophiolite was intruded through and constructed on Paleozoic and Triassic basement (Saleeby and others, 1982). (2) Plagiogranite in the 164 Ma Devil's Elbow ophiolite contains 1.7 Ga

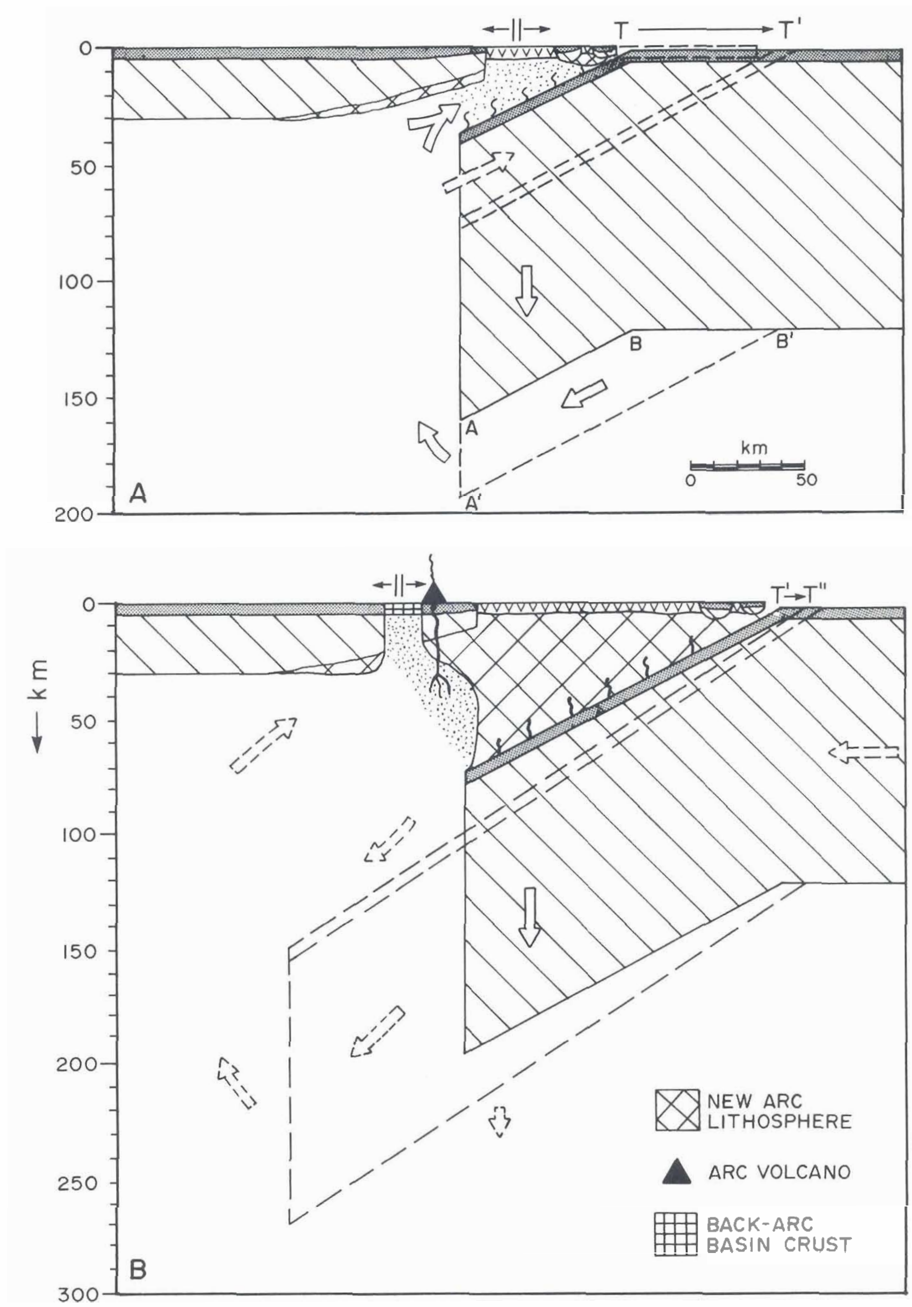

Figure 8. Section perpendicular to the arc showing why spreading in the fore arc might cease. Original lithosphere and mantle motions are outlined in solid lines, whereas new position and motions are outlined by dashed lines. (A) Continued subsidence of lithosphere results in fore-arc extension, with trench moving from T to T′ as the base of the lithosphere moves from A to A′. Retreat of subsidence hinge from T to T′ involves removal of asthenospheric material in trapezoid AA′BB′; further retreat involves removal of material progressively farther from the tip of the lithosphere (at B′), while progressive subsidence (from A to A′, and beyond) progressively cuts off route of asthenospheric transfer (shown by arrows). (B) Eventually, lithosphere begins to move with a down-dip component, and true subduction begins. This results in a greatly reduced rate of trench migration, which leads to a corresponding reduction in fore-arc extensional stresses. This allows the fore arc to cool and form thick lithosphere, forcing the magmatic axis to migrate away from the trench. Extensional strain resulting from trench rollback (T′–T″) is taken up in back-arc spreading.

xenocrystic zircons, interpreted to have been incorporated by melting of older continental crust during ophiolite formation (Wyld and Wright, 1988). (3) Conformably younger sedimentary rocks of the Galice Formation were derived in part from erosion of older Klamath terranes (Saleeby and others, 1982; Wyld and Wright, 1988); ophiolite and Galice strata were both deformed during the 145–150 Ma Nevadan Orogeny. (4) Paleomagnetic studies indicate no significant latitudinal displacement with respect to North America (Harper and Wright, 1984). It is worth noting in this regard that inferences based on radiolarian assemblages indicate transport of the 162 Ma Josephine ophiolite from Central Tethyan paleolatitudes (Pessagno and Blome, 1990), estimated to lie south of 22°N at that time, into the Southern Boreal Province by the beginning of Galice sedimentation, inferred to lie north of 30°N (Pessagno, 1991, personal commun.). The base of the Galice Formation contains Middle Oxfordian *Buchia* and Radiolaria, yielding an age of about 157 Ma (Pessagno and Blome, 1990). Taken at face value, these inferences indicate a minimum northward displacement rate of 18 cm/yr from eruption of the Josephine to deposition of the Galice Formation but conflict with previously cited evidence that the Western Klamaths formed on the margin of North America.

Whereas the Rogue Formation and Chetco Complex in the westernmost part of the Western Jurassic Belt are generally accepted to represent a magmatic arc (Garcia, 1979; Wyld and Wright, 1988), the most common model for Late Jurassic ophiolite farther east (Josephine, Preston Peak, and Devil's Elbow) is formation in a back-arc basin (for example, Harper, 1984; Harper and Wright, 1984; Wyld and Wright, 1988). Three observations contradict this conclusion and suggest instead that these ophiolites represent infant-arc crust. First, there are igneous rocks with boninitic affinities. Some dikes and lavas from the Devil's Elbow ophiolite (52%–56% SiO_2; 0.55%–0.98% TiO_2; 8.5%–10.5% MgO; 100–340 ppm Cr; 80–240 ppm Ni; Wyld and Wright, 1988) may be classified as Low-Ca, Type-2 boninites (Crawford and others, 1989). Some dikes and lavas from the Josephine ophiolite are also boninitic (Harper, 1984). Cumulate sequences in the Josephine ophiolite have abundant orthopyroxene and have a crystallization sequence of OL + SP, CPX, OPX, PLAG (Harper, 1984). This contrasts with the sequence OL, PLAG, CPX expected for crystallization of MORB and back-arc basin basalt magmas (Dick and Bullen, 1984; Pearce and others, 1984). Indeed, Harper (1984, p. 1022) noted,

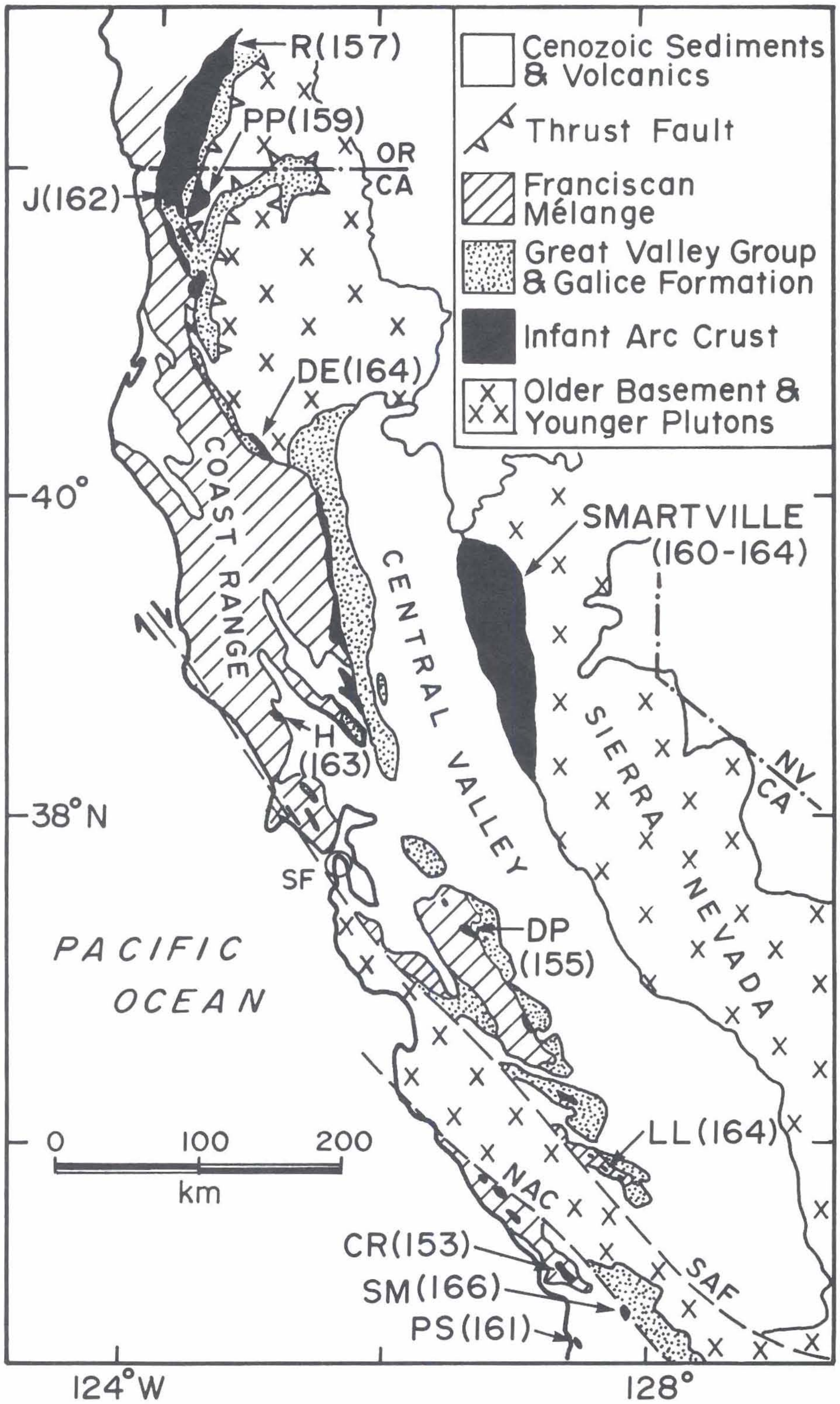

Figure 9. Late Jurassic ophiolitic and related rocks of northern California. Map modified after Hopson and others (1981), Wyld and Wright (1988), and Saleeby and others (1989). SF = San Francisco; SAF = San Andreas Fault; NAC = Nacimiento Fault. Numbers in parentheses correspond to U/Pb zircon ages, in Ma. Abbreviations and data sources are as follows (from north to south): R = Rogue Formation (Saleeby, 1984); J = Josephine ophiolite (Saleeby, quoted in Wyld and Wright, 1988); PP = Preston Peak (Saleeby and others, 1982); DE = Devil's Elbow (Wyld and Wright, 1988); Smartville (Edelman and others, 1989a; Saleeby and others, 1989); H = Healdsburg (Hopson and others, 1981); DP = Del Puerto (Hopson and others, 1981); LL = Llanada (Hopson and others, 1981); CR = Cuesta Ridge (Hopson and others, 1981); SM = Stanley Mountain (Pessagno and others, unpub. data); PS = Point Sal (Hopson and others, 1981).

"The crystallization sequence in the [Josephine] cumulates is the same as that of the Acoje Block of the Zambales ophiolite," an ophiolite with clear boninitic affinities. Second, spinels from the Josephine peridotite extend to very high Cr# (>80), consistent with the extreme melting responsible for generating infant-arc magmas and inconsistent with the lower degrees of melting expected to form back-arc basin basalts (Dick and Bullen, 1984). Third, the Klamath magmatic arc migrated from west to east following the formation of the Josephine and correlative ophiolites (Harper and Wright, 1984). This is not expected for a back-arc basin that formed behind a west-facing arc. The opening of a back-arc basin leaves an extinct, remnant arc on the portion of the rifted arc farthest from the trench, with the active arc persisting on the portion of the rifted arc nearest the trench (Karig, 1971). Eastward migration of the magmatic arc is, however, consistent with the infant-arc model.

In summary, a strong argument can be made that the arc sequences (Rogue Formation, Chetco Complex) and ophiolites (Josephine and correlatives) represent an infant-arc sequence constructed on the margin of western North America. Another similarity to the infant-arc model is the slow spreading rate inferred for the Josephine ophiolite (Harper, 1984). An interesting variation on the sequence of events observed in the IBM system is that infant-arc formation apparently involved older crust, as shown by the fact that the Preston Peak ophiolite was constructed on older crust and that an older Jurassic magmatic arc (159–174 Ma Hayfork Terrane) apparently was rifted away from the continent (Harper and Wright, 1984). Available geochronologic data suggest that this stage lasted about 10 m.y. (164–153 Ma) before the extensive west-verging thrust faulting collapsed the arc during the Nevadan Orogeny, about 150 Ma (Harper and Wright, 1984).

Arc Infancy in the Smartville Ophiolite, Western Belt, Sierra Nevada

Basement rocks of the northern Sierra Nevada have been subdivided into four belts. The westernmost belt, which borders the Central Valley, is known as the Smartville Complex and consists of a well-preserved ophiolitic sequence (Fig. 9; Day and others, 1985). Minor gabbro, more-abundant gabbro-diorite and tonalite tabular plutons, abundant diabasic dikes, and pillowed basalt comprise the sequence; the complex differs from classic ophiolites in that it lacks basal cumulate and tectonized peridotite (Beard and Day, 1987). U/Pb zircon ages are between 160 and 164 Ma (Saleeby and others, 1989; Edelman and others, 1989a). The volcanic rocks have been subdivided into older tholeiitic and younger calc-alkaline units (Menzies and others, 1980). Intrusive rocks include much granodiorite and tonalite, both as plutons and as part of a sheeted dike complex. All Smartville igneous rocks plot on a Sm-Nd isochron with an age of 178 ± 21 Ma and an initial $\epsilon_{Nd} = +9.2 \pm 0.6$, indicating derivation by mantle melting with no discernible contamination by continental crust (Saleeby and others, 1989). The geochemistry of the igneous rocks and composition of the sediments has led to a consensus that the ophiolite formed in an arc undergoing strong extension (Menzies and others, 1980; Beard and Day, 1987), and that the sediments deposited upon it were shed from a nearby arc (Xenophontos and Bond, 1978).

Until recently, the Smartville Complex was thought to be allochthonous, with the suturing of it—and terranes to the west—to North America at the end of the Jurassic being responsible for the ca. 150 Ma Nevadan Orogeny (for example, Schweickert and Cowan, 1975; Schweickert and others, 1984). Subsequent U-Pb zircon geochronologic investigations indicate that the Smartville Complex was intruded through rocks of the ca. 200 Ma Slate Creek Complex (Edelman and others, 1989a), which were already part of the North America (Edelman and others, 1989b). The Smartville Complex must now be regarded as an autochthonous part of North America, similar in this respect to the Preston Peak ophiolite of the Klamaths.

The presence of a sheeted dike complex, consisting of 100% mafic and felsic dikes trending 320° to 345°, indicates that strong extension generally occurred perpendicular to the present outcrop trend (Xenophontos and Bond, 1978; Beard and Day, 1987). Extension was accompanied by the intrusion of tonalitic and gabbro-diorite plutons (Beard and Day, 1987), and it is likely that many of the intrusions of similar age just to the east, such as the 158–161 Ma Yuba River Pluton (Edelman and others, 1989a) and the Folsom Dike Swarm and Pine Hill Intrusive Complex (Saleeby, 1982), are related to Smartville extensional magmatism. Indeed, passive emplacement was the characteristic mode of emplacement of all Late Jurassic plutons in the Foothills terrane (Tobisch and others, 1989).

The Smartville Complex may be regarded as the easternmost part of an infant-arc tract. There is strong evidence for extension, and that the extension direction was mostly perpendicular to the Late Jurassic continental margin. The Smartville Complex lies close to where the magmatic front retreated as the mantle beneath the fore arc cooled. This is demonstrated by the fact that 140 Ma tonalite-granodiorite plutons intrude the Smartville (Saleeby and others, 1989) and that the Early Cretaceous (ca. 135–95 Ma) magmatic front lay only slightly farther east (Ingersoll, 1979). In this regard, Smartville occupies a position analogous to the Mariana frontal arc islands of Guam and Saipan. It is noteworthy that Smartville igneous rocks include no boninites (although related rocks have boninitic affinities; Thy and Dilek, 1987), that orthopyroxene is not an important part of the crystallization sequence, and that Cr-Y discriminant diagrams show that the Smartville Complex was derived from a much-less-depleted source than that of Coast Range Ophiolitic rocks to the west (Pearce and others, 1984; see discussion below).

Arc Infancy in the Coast Range Ophiolite

The tectonic setting for the formation of the Coast Range ophiolite (CRO) is controversial. The CRO yields U-Pb zircon ages that range from 153 to 164 Ma, remarkably similar to the ages obtained for the Smartville Complex and the ophiolites of the western Klamaths (Fig. 9). Many workers argue that the CRO is far traveled (for example, Hopson and others, 1981; McWilliams and Howell, 1982). Critical to this discussion is the nature of the basement beneath the Central Valley and the depositional history of the Great Valley sedimentary sequence. Cady (1975) interpreted the gravity and magnetic highs over the Central Valley to indicate that it is underlain by Late Jurassic oceanic crust. This interpretation is consistent with the abundance of mafic rocks recovered from drill holes into the basement beneath the Great Valley, although Sierran felsic plutons are present beneath the eastern third of the Great Valley (Cady, 1975).

Constraints on the likelihood and timing of posited collisional events depend on understanding the strata deposited on the CRO. Dickinson and Seely (1979) argued that the Great Valley formed as a fore-arc basin, initially built on trapped oceanic crust. This basin began in the Late Jurassic period as a deep marine trough, flanked east and west by the arc massif of the Sierra Nevada and the Franciscan subduction complex; the basin now contains as much as 6 to 8 km of Late Jurassic and younger sedimentary rocks (Dickinson and Seely, 1979). Depositional environments along the west side of the Central Valley have been interpreted to show the progressive filling of the basin, without any significant hiatus (Ingersoll, 1982). The western basin edge was flanked by CRO that was uplifted accompanying subduction of Franciscan mélange; the effect of this was progressive uplift and rotation of the ophiolite and older Great Valley strata and migration of the basin axis to the east (Suchecki, 1984). Petrographic data indicate that the provenance changed with time, beginning with sediments eroded from the Klamath terranes and becoming progressively dominated by detritus eroded from unroofing Sierran plutons (Ingersoll, 1982). Therefore, the sedimentologic record indicates that the CRO east of the San Andreas fault was in its present position relative to North America when deposition of the Great Valley sequence began during latest Kimmeridgian and Tithonian time (ca. 153–146 Ma; Harland and others, 1989).

A heterogeneous sequence of tuff, volcaniclastic rocks, tuffaceous radiolarian chert, and minor volcanic rocks of the Lotta Creek unit lies between the turbidite beds of the Great Valley sequence and the CRO (Robertson, 1989). This succession has a gradational contact with the Great Valley sequence at Stanley Mountain, Del Puerto, and Llanada (Robertson, 1989). The age of the Lotta Creek unit is controversial. According to the biostratigraphy of Pessagno and others (1984, 1991, personal commun.), it ranges in age from Middle Oxfordian through upper Tithonian (ca. 157 to 146 Ma; Harland and others, 1989; Pessagno and Blome, 1990), whereas Baumgartner and Murchey (1987) argued that these radiolarian assemblages extend down through the Bathonian (ca. 166 Ma; Harland and others, 1989). Subject to these uncertainties, any suturing of the CRO to North America must have happened prior to the beginning of Lotta Creek sedimentation, ca. 166 to 157 Ma.

The CRO underlies the Lotta Creek unit everywhere except at Llanada, where about 1,300 m of volcano-sedimentary strata (the Bitterwater Canyon unit of Robertson [1989], Member II of Lagabrielle and others [1986]) intervene. This unit contains minor basalts, more abundant intermediate and felsic volcanic rocks, and volcaniclastic strata erupted in an arc setting (Hopson and others, 1981), probably on the flanks of a submarine volcano (Robertson, 1989).

The nature of the contact between the CRO and the overlying units is controversial. Hopson and others (1981) argued that a hiatus of 10–20 m.y. existed between the ophiolite and its sedimentary cover; correlation of the biostratigraphy with improved time scales has reduced this gap to 6–12 m.y. (Pessagno and others, 1991, personal commun.). The suggestion that some hiatus exists finds additional support in the presence of an erosional unconformity between the CRO and Member II at Llanada (Lagabrielle and others, 1986) and by a change in depositional environments represented by ophiolitic interpillow sediments and those of the overlying strata (Hopson and others, 1981). Ophiolitic breccia overlies the CRO north of San Francisco (Phipps, 1984) and indicates that uplift followed sea-floor spreading. Robertson (1989) argued, however, that an important unconformity does not exist for two reasons. First, the uncertainties in correlations between the biostratigraphy of the Lotta Creek Unit and the isotopic ages for the CRO are sufficiently large that the 6–12 m.y. time gap may be an artifact of the correlation. Second, the presence of tuff and felsic volcanic rocks in the upper part of the CRO indicates proximity to an arc throughout its evolution. MacPherson and Phipps (1988) reported that olistostromal units at the base of the Great Valley Group contain blocks that are similar to those in the Franciscan mélange and the CRO, and MacPherson and others (1990) argued that underplating of the Franciscan mélange beneath the fore arc may have caused uplift of the CRO, a conclusion that agrees with Suchecki (1984).

Progressive cooling of the fore-arc region following an infant-arc stage of sea-floor spreading may be inferred from the alteration history of the Great Valley sequence. There is evidence that the CRO was still hot when deposition of the Great Valley Group began. Suchecki and Land (1983) reported that the lowest several hundred meters of Great Valley strata were altered by hydrothermal fluids emanating from the subjacent ophiolitic rocks, and that the Upper Jurassic and Lower Cretaceous strata were subjected to a geothermal gradient of 25 °C/km. Dumitru (1988) concluded that the geothermal gradients decreased to about 15 °C/km at 90 Ma and to no more than 9 °C/km by 65 Ma.

Paleomagnetic and paleontologic evidence has been offered that the CRO is far traveled. Hopson and others (1981) summarized paleomagnetic data that indicated that the ophiolite west of the San Andreas fault formed near the equator. McLaughlin and others (1988) argued on this basis that the CRO moved north at a rate of about 30 cm/yr before becoming accreted to North America at 145 Ma. Paleomagnetic data for ophiolitic fragments east of the San Andreas fault—those unaffected by San Andreas right-lateral motions—give conflicting results, with Williams (1983) arguing for no significant latitudinal displacement and Beebe and Luyendyk (1983) arguing that the ophiolite came from far to the south. Frei and Blake (1987) examined four localities and concluded that there had been a complete remagnetization of the ophiolite, a result that is consistent with paleomagnetic data for the Great Valley Group. Mankinen and others (1991) argued that where primary remanent magnetization could be measured, the CRO had experienced no significant latitudinal displacement with respect to North America. Finally, Butler and others (1991) concluded that the paleomagnetic evidence for a near equatorial origin for Mesozoic western California and Baja California could be accounted for by a combination of compaction in sediments, rotational uplift of the Peninsular Ranges, and better estimates of the position of North America during the Mesozoic. We conclude that the existing paleomagnetic data do not require the CRO to be a terrane that is exotic to North America.

The faunal assemblages overlying the CRO have been interpreted as indicating rapid northward motion of the ophiolite during the Late Jurassic, a time when North America was moving very slowly (<1 cm/yr) to the north (for example, Hopson and others, 1981). Pessagno and Blome (1986) draw attention to the fact that the radiolarian assemblages immediately overlying the ophiolite correspond to the Central Tethyan Province and are succeeded by assemblages with Southern Boreal affinities. We are impressed by the similarity of this transition with that of the Western Jurassic Belt of the Klamaths. The fact that the Klamath strata that formed on the margin of North America show similar faunal shifts as strata overlying the CRO suggest that these assemblages may not track latitudinal displacements. For this reason, we do not believe that the latitude of origin of the CRO is constrained by the studied faunal assemblages.

If the CRO is allochthonous, then there must be a suture between it and basement to the east. Because the Smartville Complex is autochthonous, any suture must lie beneath the Great Valley; such a structure has not been identified, nor is any evidence for collision preserved in the sedimentary record. Furthermore, one of the most important arguments supporting an allochthonous interpretation for the CRO is the interpretation of the Nevadan Orogeny as a collisional event. Structural studies (Tobisch and others, 1987, 1989) indicate that Nevadan deformation is most intense in the western, and disappears in the eastern, Sierras and that Nevadan structures record brittle, not ductile, deformation. These data are interpreted to indicate that Nevadan deformation in the Sierra Nevada was much less intense than heretofore thought, so that a major collision may not be responsible. Sharp (1988) suggested an analogy with modern deformation of the Andes, where extensive shortening in the arc and back-arc regions is unrelated to collision of allochthonous terranes.

The mineralogical and chemical composition of the CRO was originally interpreted as indicating an origin as normal oceanic crust (Bailey and others, 1970; Hopson and Frano, 1977; Hopson and others, 1981). The interpretation that the CRO formed at a mid-oceanic spreading ridge was first challenged by Evarts (1977), who noted the abundance of felsic rocks, inferred a high content of H_2O for all components of the ophiolite, and concluded that it formed in an arc or back-arc setting. The conclusion that the CRO is an SSZ ophiolite has become increasingly compelling (Blake and Jones, 1981; Pearce and others, 1984; Shervais and Kimbrough, 1985; Lagabrielle and others, 1986; Shervais, 1990). It is difficult to resolve on chemical or sedimentological bases whether the CRO formed in an arc or back-arc basin setting, but several lines of evidence favor the former. First, limited data indicate the presence of Cr-rich spinel (Cr# up to 82; Evarts, 1977). Second, CRO volcanic rocks are depleted in high–field-strength cations and lack an increase in TiO_2 with fractionation, chemical features diagnostic for arcs (Lagabrielle and others, 1986; Shervais, 1990). Third, plutonic rocks include both orthopyroxene and hornblende cumulates and have very calcic plagioclase, characteristics of arc gabbros (Shervais, 1990). Finally, if the ophiolite is a back-arc basin, where is the arc with which it is associated? There is no evidence for it ever lying to the west, where the Franciscan mélange is today; ophiolitic blocks in the Franciscan are predominantly alkalic (Shervais, 1990), although blocks of CRO and MORB are also common (MacPherson and others, 1990). There is no evidence that it lay to the east; thickened arc crust is unlikely to become the site of a sedimentary basin such as the Great Valley, and support for models involving Late Jurassic sutures in the northern Sierran foothills is evaporating (Edelman and others, 1989a). The tectonic setting of volcanic blocks in olistostromal deposits at the base of the Great Valley Group indicate sources from the CRO and the Franciscan mélange, with no evidence for an arc distinct from the CRO (MacPherson and Phipps, 1988). Nevertheless, the presence of interpillow limestone at some localities (Robertson, 1989; Hopson and others, 1981) succeeded by radiolarian chert indicate that the CRO formed at water depths near to the carbonate compensation depth, about 3–4 km deep throughout the Mesozoic.

We believe that sufficient evidence warrants application of the infant-arc model to the Late Jurassic ophiolites of California. The region from the eastern edge of the Smartville Complex across the mafic-ultramafic crust beneath the Great Valley to the westernmost exposure of Coast Range ophiolite formed by slow sea-floor spreading, in a position that was fundamentally autochthonous to North America (Ernst, 1984). This infant-arc crust extended to the north to join up with now-telescoped correlative ophiolites in the western Klamaths (Fig. 9) and probably extended much farther north and south, as far as Washington and Baja California. The time span 164 to 153 Ma encompassed by these ophiolites bracket the infant-arc episode, with subsequent magmatic activity shifting farther east as the infant arc nearest the trench became a depocenter for sediments shed first from the Klamaths and then from the unroofing Sierra Nevada batholith. The first stages of Franciscan metamorphism probably manifest an initially hot mantle; Mattinson (1988) concluded that high-grade (eclogitic) blocks were metamorphosed during a very limited period between 165 and 150 Ma. Mattinson (1988, p. 1024) further noted that the oldest ages ". . . clearly reflect initiation of subduction, as indicated by various types of thermal modelling." Wakabayashi (1992) argued that this was because the hanging wall of the Franciscan subduction zone cooled very rapidly during the first 15×10^6 years after subduction initiation. We further suggest that regional extension was taken up as back-arc spreading along the 148–150 Ma Independence dike swarm (Chen and Moore, 1979; Lahren and others, 1990) after the infant-arc episode was over.

The dimensions of the IBM infant-arc crust compared to that of the California Jurassic are similar and support our model (Fig. 10). The similar widths and time spans inferred for both episodes (ca. 15–10 Ma) indicate similarly high crustal growth rates for each (Fig. 5). Application of the infant-arc model solves one of the greatest stumbling blocks to accepting the Late Jurassic ophiolites of California as autochthonous, SSZ ophiolites that formed over an east-dipping subduction zone: ". . . the apparent lack of actualistic models . . ." (Robertson, 1989, p. 215).

The most serious argument against the proposed model concerns the tectonic setting of California prior to the infant-arc episode. In the case of the IBM system, there is evidence that the subduction zone nucleated at a transform fault. The tectonic setting in California during the Jurassic is less clear, although the southwestern margin of North America was truncated by strike-slip faulting during latest Paleozoic and/or early Mesozoic times (Burchfiel and Davis, 1972; Anderson and Schmidt, 1983; Stewart, 1988). The Kings River ophiolite formed ca. 200 Ma (Saleeby, 1982) and has been inter-

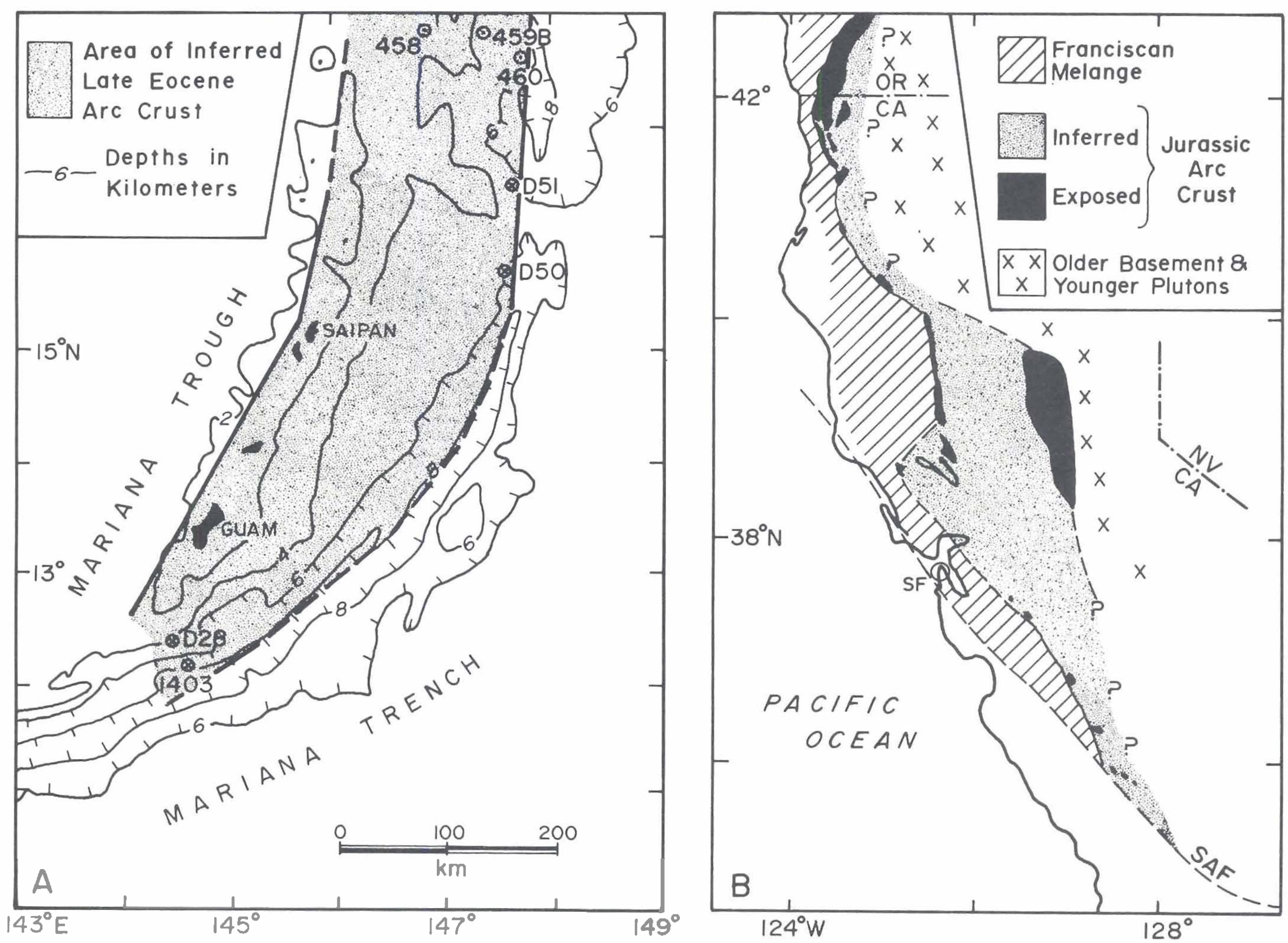

Figure 10. Comparison of extent of infant-arc crust in the southern Mariana arc (A) with that inferred for California (B), at the same scale. Telescoping of the infant-arc crust in the Klamath Mountains occurred during the Cretaceous and, if unrolled, would be similar in width to that across the northernmost Central Valley.

preted as being displaced along a major strike-slip fault near the oceanic-continental interface (Saleeby, 1977). Saleeby (1981) and Harper and others (1985) stressed the idea that the Mesozoic arcs and sedimentary basins of the Sierran foothills formed on the sites of older strike-slip faults. There are also indications that subduction was underway before the arc infancy phase identified here: granitic plutons in the eastern Sierra Nevada interpreted to be related to subduction are of pre–Late Jurassic and Triassic age (Chen, 1982), and lawsonite-bearing blueschist, of probable Triassic age along the northern Foothills suture (Schweickert, 1976). It is noteworthy that monzonite and syenite characterize the pre–Late Jurassic plutons (Miller, 1977; Barton and others, 1988), compositions generally not produced early in the magmatic evolution of an arc. Gabbro, tonalite, and quartz diorite are the expected products of primitive arcs, and these are not common in California until Late Jurassic time (Barton and others, 1988). A reconciliation may come from the observations of Busby-Spera (1988), who noted that the early Mesozoic magmatic axis occupied a continuous depression, characterized by high subsidence rates and extensive faulting. This axis was quite different from that of typical Andean-type margins in that it did not block sediment transport from the interior of North America. Busby-Spera (1988) suggested that a modern analogue may be the arc-graben of Central America, but entertained a transtensional setting as well.

We conclude that, although evidence for early Mesozoic arc activity must be recognized, there are also compelling indications for strike-slip faulting at this time. A better understanding of the tectonic setting of California before Late Jurassic arc infancy is needed before the model can be embraced, but we believe that the essential elements of our model will withstand critical evaluation. With this in mind, we summarize and compare the evolution of the IBM and California infant arcs in Figure 11.

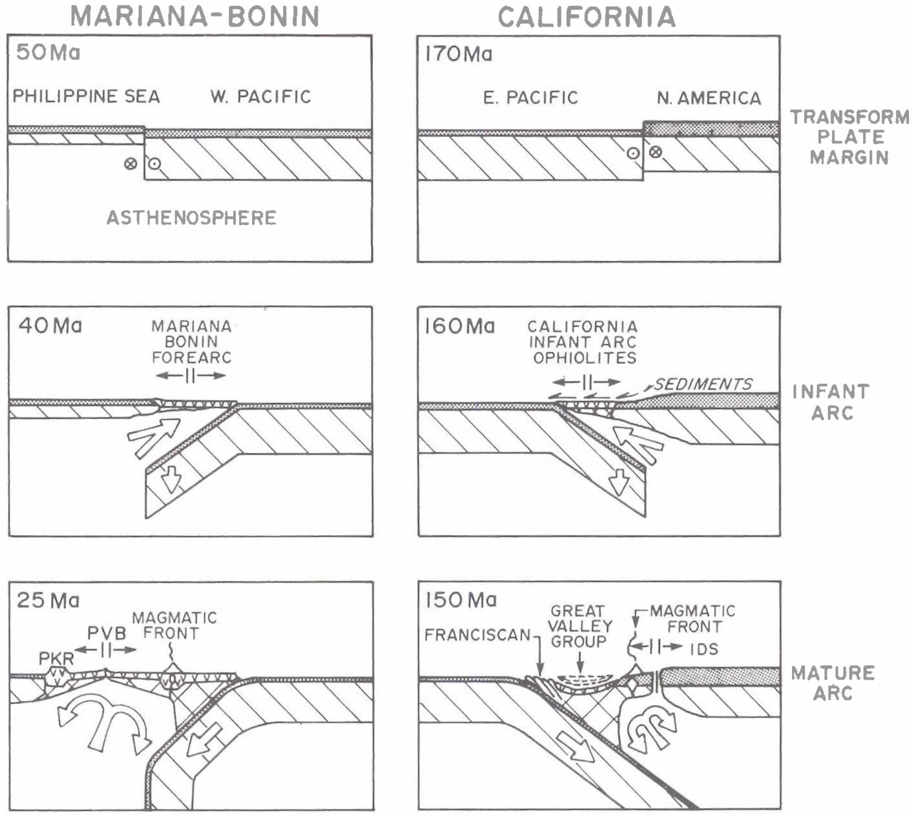

Figure 11. Summary diagram comparing the pre–infant-arc, infant-arc, and mature-arc stages for the Mariana-Bonin and California systems. Symbols are those of Figure 6. Asymmetric arrows show sediment transport from the east onto, and across, the infant-arc crust to form wackes of the Great Valley and Franciscan Groups. IDS = Independence Dike Swarm.

CONCLUSIONS

Application of the infant-arc model to the late Eocene IBM system explains several otherwise puzzling features. Most important among these are (1) the much greater width affected by arc activity compared to modern arcs, none of which are in a comparable stage of development to that of the IBM system during Eocene time; (2) indications of early sea-floor spreading; (3) the generation of boninite; and (4) the much higher crustal growth rate of the arc compared to its present crustal growth rate. Application of the infant-arc model to the Late Jurassic crustal evolution of California is more controversial. We conclude that the Late Jurassic ophiolites of California comprise a single tract of fore-arc crust that formed in its present position relative to North America, and that the similarities in the nature and duration of igneous activity between the late Eocene IBM arc and the Late Jurassic of California are so striking that the infant-arc model must be seriously entertained. Application of the model satisfactorily explains similarities in age of the earliest and highest grade metamorphic rocks in the Franciscan and the time of formation of these ophiolites, and it provides new insights into the depositional history of the Great Valley Group.

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