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Multi-Stage Origin of the Coast Range Ophiolite, California: Implications for the Life Cycle of Supra-Subduction Zone Ophiolites

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Abstract

The Coast Range ophiolite of California is one of the most extensive ophiolite terranes in North America, extending over 700 km from the northernmost Sacramento Valley to the southern Transverse Ranges in central California. This ophiolite, and other ophiolite remnants with similar mid-Jurassic ages, represent a major but short-lived episode of oceanic crust formation that affected much of western North America. The history of this ophiolite is important for models of the tectonic evolution of western North America during the Mesozoic, and a range of conflicting interpretations have arisen. Current petrologic, geochemical, stratigraphic, and radiometric age data all favor the interpretation that the Coast Range ophiolite formed to a large extent by rapid extension in the forearc region of a nascent subduction zone. Closer inspection of these data, however, along with detailed studies of field relationships at several locales, show that formation of the ophiolite was more complex, and requires several stages of formation.

Our work shows that exposures of the Coast Range ophiolite preserve evidence for four stages of magmatic development. The first three stages represent formation of the ophiolite above a nascent subduction zone. Rocks associated with the first stage include ophiolite layered gabbros, a sheeted complex, and volcanic rocks with arc tholeiitic or (more rarely) low-K calc-alkaline affinities. The second stage is characterized by intrusive wehrlite-clinopyroxenite complexes, intrusive gabbros, Cr-rich diorites, and volcanic rocks with high-Ca boninitic or tholeiitic ankaramite affinities. The third stage includes diorite and quartz diorite plutons, felsic dike and sill complexes, and

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calc-alkaline volcanic rocks. The first three stages of ophiolite formation were terminated by the intrusion of mid-ocean ridge basalt dikes, and the eruption of mid-ocean ridge basalt or ocean-island basalt volcanic suites. We interpret this final magmatic event (MORB dikes) to represent the collision of an active spreading ridge. Subsequent reorganization of relative plate motions led to sinistral transpression, along with renewed subduction and accretion of the Franciscan Complex. The latter event resulted in uplift and exhumation of the ophiolite by the process of accretionary uplift.

Introduction

THE WESTERN MARGIN of North America is characterized by extensive tracts of ophiolitic basement with radiometric ages of 173 to 160 Ma or younger (Hopson et al., 1981; Saleeby et al., 1982; Wright and Sharp, 1982; Saleeby, 1983; Coleman, 2000). The Coast Range ophiolite of California is the most extensive of these tracts, which also include the Fidalgo ophiolite complex in Washington state (Brown et al., 1979), the Smartville ophiolite in the Sierra Foothills (Xenophontos and Bond, 1978), the Josephine ophiolite of southern Oregon and northern California (Harper, 1984; Harper et al., 1985), the Preston Peak ophiolite in the Klamath Mountains (Snoke, 1977), and the Cedros ophiolite in Baja California (Kimbrough, 1982). The regional extent of these ophiolite belts, and the narrow range in their ages of formation, make their petrogenesis one of the more important (and contentious) tectonic problems in the Cordillera (e.g., Dickinson et al., 1996).

The Coast Range ophiolite (CRO) was first recognized by Bailey et al. (1970) and later studied in detail by Hopson et al. (1981). Early studies of the CRO emphasized its similarity to oceanic crust formed at mid-ocean ridge spreading centers, and tectonic interpretations focused on the obduction of intact oceanic lithosphere (e.g., Bailey et al., 1970; Page, 1972; Hopson and Frano, 1977; Hopson et al., 1981, 1992, 1997). These studies provided the fundamental petrologic and structural framework for later investigations, and established the overall "oceanic" nature of the ophiolite.

Later petrologic and geochemical studies demonstrated that the Coast Range Ophiolite formed in a supra-subduction zone environment, possibly by fore-arc or intra-arc rifting (Evarts, 1977; Blake and Jones, 1981; Saleeby, 1982; Shervais and Kimbrough, 1985a, 1985b; Lagabrielle et al., 1986; Shervais, 1990, 2001; Stern and Bloomer, 1992). Petrologic evidence for this interpretation includes the presence of both keratophyre and quartz keratophyre (metamorphosed andesite and rhyolite) in the hypabyssal and volcanic sections, magmatic hornblende in the plutonic section, and the crystallization of clinopyroxene before plagioclase in the cumulates. Geochemical evidence includes low Ti, Na, and Cr concentrations in relict clinopyroxene, relative depletion of high-field-strength trace elements in extrusive rocks, and a trend toward silica-saturation with increasing FeO/MgO (Bailey and Blake, 1974; Evarts, 1977; Blake and Jones, 1981; Shervais and Kimbrough, 1985a; Shervais, 1990, 2001; Giaramita et al., 1998).

More recent work on the CRO suggests that none of the simple, single-stage interpretations advanced earlier are entirely correct, and that the CRO had a brief but complex history. In particular, the occurrence of MORB-like volcanic and hypabyssal rocks within stratigraphic or tectonic settings characteristic of the CRO suggests that models that propose a supra-subduction zone origin are too simple. New data gathered over the last 10 years suggest that although a supra-subduction zone setting was the dominant focus of CRO origin and development, the interaction between tectonic and igneous processes that form mid-ocean ridges and convergent plate boundaries was important, especially during the later stages of ophiolite development (Shervais, 1993, 2001; Murchey and Blake, 1993). These observations suggest that our current concepts on the development of supra-subduction zone ophiolites are too simple and need to be revised.

In this contribution we synthesize the results of recent and ongoing studies of the CRO and reexamine previous ideas about its development in light of this synthesis. We then generalize these observations to place the CRO within the context of other supra-subduction zone ophiolites, and assess the implications of our results for the tectonic evolution of the western Cordillera.

Geology and Setting of the Coast Range Ophiolite

Field setting

The Coast Range ophiolite is represented by a series of dismembered ophiolitic fragments scattered along the Coast Range fault (the tectonic

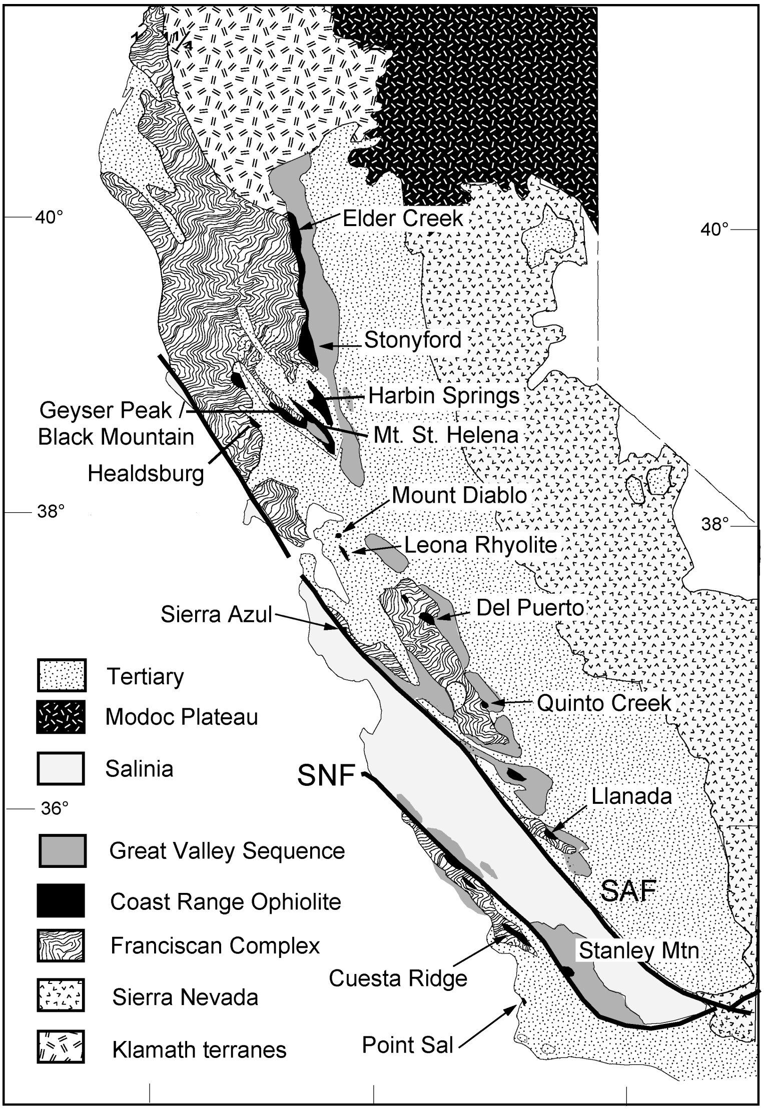


FIG. 1. Simplified geologic map of California showing Coast Range ophiolite localities discussed in this paper (black), along with the main geologic subdivisions of the state: batholiths and metamorphic rocks of the Sierra Nevada, Klamaths, and Salinia, sediments of the Jurassic–Cretaceous Great Valley Sequence, the Jurassic–Paleogene Franciscan Complex, Tertiary rocks of the Modoc Plateau, and Cenozoic rocks unrelated to the other units. Abbreviations: SAF = San Andreas fault system; SNF = Sur-Nacimiento fault system.

contact between the CRO and the underlying Franciscan Complex; see Jayko et al., 1987) from Elder Creek in the north to Point Sal in the south (Fig. 1). When Neogene wrench faulting along the San Andreas and Sur-Nacimiento fault systems is restored, the ophiolite belt defines an original extent of over 1300 km (Hopson et al., 1981). The ophiolite is overlain depositionally by Upper Jurassic strata of the basal Great Valley Series (Knoxville Formation in northern California, Toro and Espada formations in central California, Riddle Formation in southwest Oregon), a flysch deposit derived from emergent arc

TABLE 1. Comparison of Major Occurrences of Coast Range Ophiolite in California and Their Correspondence with the Proposed Life Cycle for Suprasubduction Zone (SSZ) Ophiolites

Life cycle:	Stage 1: Birth	Stage 2: Youth	Stage 3: Maturity	Stage 4: Death	Stage 5: Resurrection	Primary references
Events: Ophiolite	Initial spreading hinge rollback? Arc tholeiite	Refractory melts, second-stage melting	Stable arc (calc-alkaline or arc tholeiite)	Ridge subduction	Accretionary uplift or obduction	For all: Hopson et al., 1981; Shervais and Kimbrough, 1985a
Elder Creek volcanics	Dike complex, pillows, clasts in Crowfoot Point Breccia	Clasts in Crowfoot Point Breccia (ol-px phyric)	Clasts in Crowfoot Point Breccia (andesite)	MORB dikes cut gabbro-diorite	Accretionary uplift by Franciscan Complex	Hopson et al, 1981; Shervais and Kimbrough, 1985a; Lagabrielle et al, 1986; Shervais and Beaman, 1989, 1991, 1994
Elder Creek plutonics:	Layered ± foliated gabbro, dunite	Gabbro pegmatoid–wehrlite–clinopyroxenite high-Cr diorites	Hb diorite, qtz diorite plutons	Franciscan high-grade blocks (U/Pb, Ar/Ar ≈162–159 Ma)		
Stonyford Complex volcanics	Melange blocks of pillow lava	Melange blocks of pillow lava		Stonyford volcanics	Accretionary uplift by Franciscan Complex	Hopson et al, 1981; Shervais and Kimbrough, 1985a; Shervais and Hanan, 1989; Zoglman, 1991; Zoglman and Shervais, 1991
Stonyford Complex plutonics	Layered gabbro blocks in mélange	Wehrlite-clinopyroxenite blocks in mélange	Qtz diorite blocks in mélange below volcanics	Franciscan high-grade blocks (U/Pb, Ar/Ar ≈162–159 Ma)		
Black Mountain volcanics	Diabase sill complex	?	?	MORB pillow lava, diabase breccia	Accretionary uplift by Franciscan Complex	McLaughlin and Pessagno, 1978; McLaughlin and Ohlin, 1984; McLaughlin, 1978; Giaramita et al., 1998
Black Mountain plutonics	Layered gabbro	?	?			
Mount Diablo volcanics	Arc basalt pillow lava, diabase sill complex		Keratophyre (andesite)	MORB dikes cut gabbro-diorite	Accretionary uplift by Franciscan Complex	Williams, 1983; Hagstrum and Jones, 1998
Mount Diablo plutonics						
Leona Rhyolite volcanics			Rhyolite vitrophyre	Mid-Ocean Ridge Basalt glass inclusions	Accretionary uplift by Franciscan Complex	Jones and Curtis, 1991; Shervais, 1993
Leona Rhyolite plutonics	Layered gabbros underlie rhyolite					

Del Puerto Canyon	Pillow lava, massive flows, sills (cpx-plg phyric)	Pillow lava, massive flows (ol-cpx phyric), boninites	Volcanics: Lotta Creek tuff, Leona rhyolite; qtz keratophyre lavas, sills	late MORB dikes	Accretionary Uplift by Franciscan Complex	Evarts, 1977; Evarts and Schiffman, 1983; Hopson et al, 1981; Evarts et al, 1992, 1999; Robertson, 1989
Plutonics:	Plutonics: dunite, layered gabbro	Plutonics: wehrlite, clinopyroxenite, dunite	Plutonics: Hb qtz diorite plutons, 200m+ thick	Franciscan High grade blocks (U/Pb, Ar/Ar ~ 162–159)		
Quinto Creek	Pillow Lava Layered Gabbros				Accretionary Uplift by Franciscan Complex	Bailey et al, 1970, Hopson et al, 1981
Llanada Volcanics:	Pillow lava (cpx-plg phyric)	Pillow lava (ol-cpx phyric)	Lotta Creek tuff; qtz keratophyre lavas; andesite breccia, lahars	Not seen	Accretionary Uplift by Franciscan Complex	Giaramita et al, 1998; Lagabrielle et al, 1986; Robertson, 1989; Hopson et al, 1981; Emerson, 1979; Bailey et al, 1970
Plutonics:	Layered gabbro	Feldpathic wehrlite, dunite	Keratophyre sills			
Sierra Azul Volcanics:	Pillow basalt	?	Keratophyre, andsite breccias and tuffs	MORB dikes cut gabbro-diorite	Accretionary Uplift by Franciscan Complex	McLaughlin et al., 1991, 1992
Plutonics:	Layered gabbro	Ultramafics?	Diorite, Qtz diorite,			
Cuesta Ridge Volcanics:	??	Boninite lavas	Basaltic andesite to dacite lavas	MORB dikes cut gabbro-diorite	Accretionary Uplift by Franciscan Complex	Page 1972; Pike, 1974
Plutonics:	Rare gabbro	Wehrlite, lherzolite, clinopyroxenite	Qtz diorite, keratophyre Qtz gabbro, diabase			
Point Sal Volcanics:	Lower pillow lavas (plag-cpx phyric)	Upper pillow lavas (olivine-phyric)	Tuffaceous chert	Not seen	Accretionary Uplift by Franciscan Complex	Hopson and Frano, 1977; Hopson et al, 1981; Mattinson and Hopson, 1992
Plutonics:	Layered, foliated gabbro, dunite	Wehrlite-clinopyroxenite	Keratophyre sill complex, plagiogranite			
Snow Camp Mtn. (Oregon) Volcanics:					Accretionary Uplift by Dothan Formation Franciscan Complex	Blake et al., 1985; Harper, personal communication, 1999
Plutonics:	Layered gabbro		Hornblende quartz diorite dikes			
Wild Rogue Wilderness (Oregon) Volcanics:	Pillow lava Dike Complex Group 1	basalt, basaltic andesite: (lower volcaniclastics) Dike Complex Group 2	Hornblende andesite, dacite dikes, lavas (upper volcaniclastics)	Rare MORB dikes	Accretionary Uplift by Dothan Formation	Blake et al., 1985; Harper, personal comm., 1999; Kosanke, 2000
Plutonics:	Layered gabbro	High Cr diorites	Hornblende quartz diorite stocks, dikes			

terranes to the north and east (Bailey et al., 1970; Ingersoll and Dickinson, 1981). Basal Great Valley sediments in northern California contain *Buchia rugosa* of late Kimmeridgian to early Tithonian age (~152–148 Ma; Palfey et al., 2000), which places an upper limit on the time of ophiolite formation (Jones, 1975; Imlay, 1980).

Exposures of CRO vary greatly in their character from place to place (Table 1). Some CRO remnants feature more or less complete ophiolite stratigraphy (although all are complexly faulted and dismembered), volcanic rocks with arc-like geochemistry, and depositional contacts with overlying Upper Jurassic sediments of the Great Valley Sequence. At other locales, plutonic rocks are scarce, the volcanic rocks are more like "oceanic" basalts (that is, basalts erupted at oceanic spreading centers or within oceanic plates), and contacts with sediments of the Great Valley Sequence may be faulted. The differences between these occurrences suggest that the CRO represents a complex composite terrane formed through the interplay of different tectonic processes, as originally proposed by Shervais and Kimbrough (1985a). In order to determine what processes were involved, and their influence on ophiolite formation, we need to review in some detail the results of work carried out over the last decade at various CRO localities. Much of this work has not been published except in abstracts or theses, so this summary represents the first description of geologic relations in several of these locales.

The CRO in California is divided geographically into three distinct belts: the Northern or Sacramento Valley belt, the Central or Diablo Range belt, and the Southern or Coastal belt, which is separated from the other belts by two major wrench fault systems (the San Andreas and Sur-Nacimiento fault systems) and by the Salinia terrane. Ophiolites in the Sacramento Valley belt are typically (but not everywhere) overlain by an ophiolite-derived sedimentary breccia, whereas ophiolites in the other belts are typically overlain by tuffaceous chert or cherty tuff (Hopson et al., 1981). Lesser known remnants of Jurassic ophiolite in southwest Oregon that are correlated with the CRO of California crop out west of the Rogue-Chetco arc complex (Blake et al., 1985; Kosanke et al., 1999).

Northern (Sacramento Valley) belt

The Northern or Sacramento Valley belt of CRO crops out along the western margin of the Sacramento Valley and in small remnants that have been offset by late Neogene wrench faulting (Fig. 1). The CRO and overlying Great Valley Sequence in this belt are separated from blueschist-facies schists and semi-schists of the Franciscan Complex by serpentinite-matrix mélange that crops out continuously along the western edge of the valley (Bailey et al., 1964; Jayko and Blake, 1987). This serpentinite mélange is commonly considered to be part of the CRO, but recent data suggest that a mixed CRO/Franciscan provenance is more likely (Jayko et al., 1987; MacPherson et al., 1990; Huot and Maury, 2002). Pillow lava knockers in this serpentinite matrix mélange resemble backarc basin basalts, and are overlain conformably by banded radiolarian cherts resembling those in the Franciscan Complex (Huot and Maury, 2002).

There are three main ophiolite locales in this belt—the Elder Creek ophiolite, the Stonyford Volcanic Complex, and the Black Mountain/Geyser Peak ophiolite—plus three smaller remnants: Mount Saint Helena, Harbin Springs, and Healdsburg. These ophiolites collectively exhibit the entire range of lithologic and tectonic diversity observed in the CRO. Additional fragments not discussed here include the Fir Creek, Wilbur Springs, and Bennett Creek ophiolite remnants (e.g., Hopson et al., 1981). Ophiolite remnants in the Clear Lake–Sonoma County area (Black Mountain/Geyser Peak, Mt. St. Helena, Harbin Springs, Healdsburg) have all been displaced northwestward during the Neogene along faults related to the San Andreas system (Fig. 1).

Elder Creek. The Elder Creek ophiolite remnant is the northernmost exposure of CRO in California, and also one of the best preserved (Fig. 2). It comprises a nearly complete ophiolite sequence, with cumulate mafic and ultramafic rocks, non-cumulate gabbro and diorite, dike complex, and volcanic rocks (Bailey et al., 1970; Hopson et al., 1981; Shervais, 2001).

Igneous rocks of the ophiolite are overlain by the Crowfoot Point Breccia, a coarse, unsorted fault-scarp talus breccia which varies from <10 m to over 1000 m in thickness (Lagabrielle et al., 1986; Robertson, 1990). This unit was deposited on an eroded surface that crosscuts all other units of the ophiolite (from cumulate ultramafics through dike complex). The Crowfoot Point Breccia contains clasts that range in composition depending on the level of erosion, such that volcanic clasts are most common where the breccia sits on eroded dike complex, and plutonic clasts are most common where the breccia sits on eroded plutonic complex. The Crowfoot Point

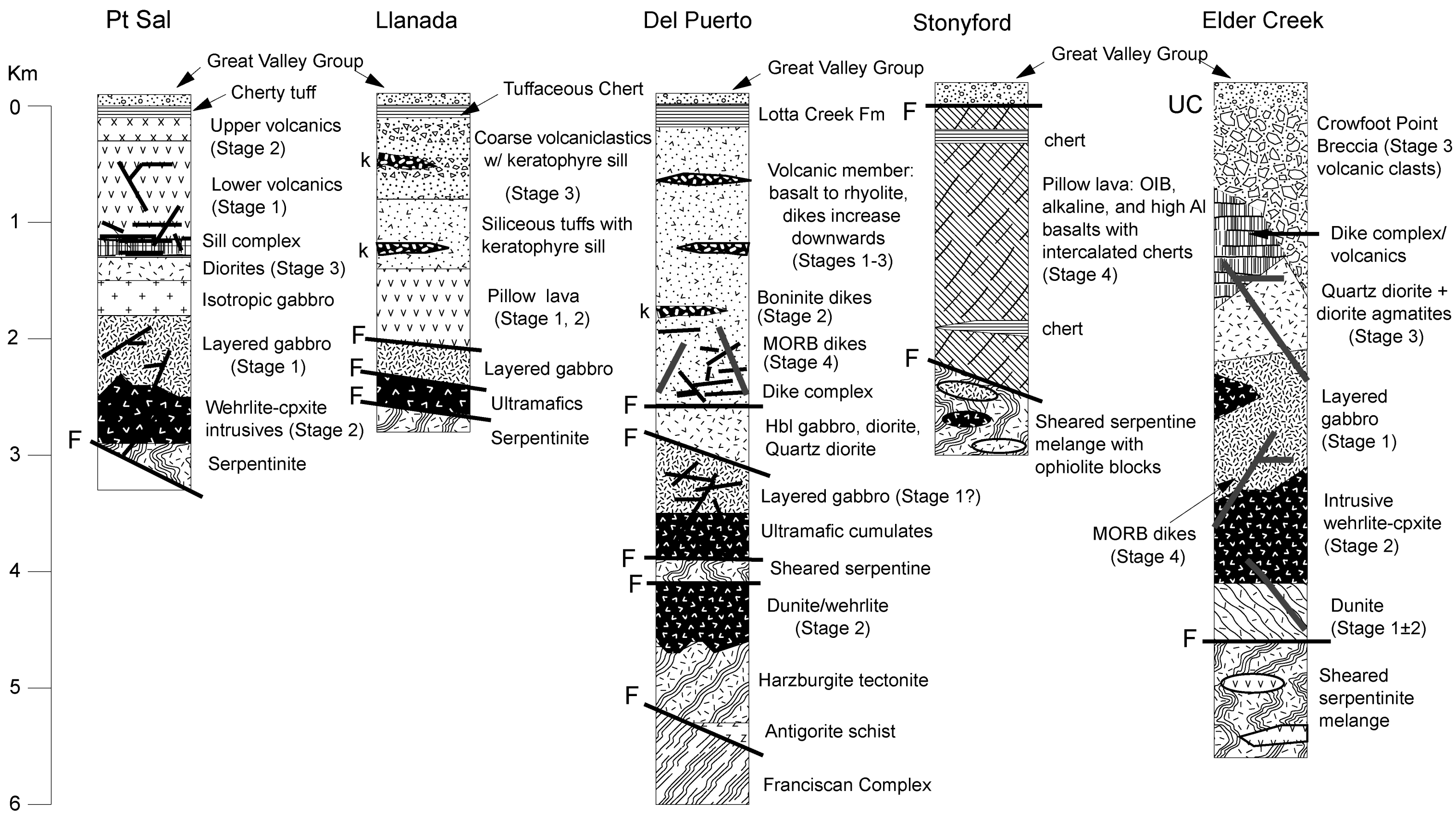


FIG. 2. Simplified stratigraphic columns of five representative Coast Range ophiolite localities: Point Sal, Llanada, Del Puerto Canyon, Stonyford, and Elder Creek. Thick lines crossing sections and labeled "F" are faults; crosscutting lines within sections are dikes of various generations and crosshatched wedges labelled "k" are keratophyre sills or dikes. Point Sal section after Hopson et al. (1981); Llanada section after Emerson (1979) and Hopson et al. (1981); Del Puerto section after Evarts et al. (1999); Stonyford Volcanic Complex and Elder Creek from unpublished data of Shervais and master's theses by Zoglman (1991) and Beaman (1989).

Breccia is itself overlain depositionally by the Upper Jurassic Great Valley Sequence, which here contains fossils of *Buchia rugosa* of late Kimmeridgian age (Bailey et al., 1970; Jones, 1975; Blake et al., 1987).

Field relations and geochemistry of the Elder Creek remnant indicate that four magmatic episodes are required for its formation (Beaman, 1989; Shervais 2001). The first magma series comprises cumulate dunite and gabbro, isotropic gabbro, dike complex, and by some of the massive volcanics. The cumulate gabbros may exhibit foliation formed by extension, or be cut by ductile faults. The second magmatic series comprises clinopyroxenite-wehrlite intrusions with less common gabbro and gabbro pegmatoid. Intrusive relations are shown by xenoliths of layered gabbro or dunite within pyroxenite and gabbro pegmatoid, and by truncation of layering along intrusive contacts between the two series. The third magmatic episode is represented by isotropic gabbro, agmatite composed of xenoliths of cumulate or foliated gabbro, volcanic rock, and dike complex in a diorite or quartz diorite matrix; diorite and quartz diorite stocks and dikes that intrude all of the older lithologies; and felsite dikes that are marginal to the quartz diorite plutons. This episode is also represented by volcanic clasts in the Crowfoot Point breccia that are calc-alkaline in composition. The fourth magma series is represented by basaltic dikes that crosscut rocks of the older episodes, and basaltic pillow lavas that form volcanic lenses within the Crowfoot Point Breccia.

Geochemical data are consistent with formation of the first three series in a supra-subduction zone (arc) environment, but all require distinct parent magmas. Rocks of the first magmatic series represent primitive arc tholeiite magmas, with a narrow range of silica contents (53–66% SiO_2), low TiO_2 and Cr concentrations (Fig. 3), low Ti/Zr and Ti/V ratios (Fig. 4), and LREE-depleted REE patterns. In contrast, parent magmas for the second magmatic series were more primitive in their major-element characteristics, similar to boninites and tholeiitic ankaramites found in many western Pacific arcs. These rocks have high Cr concentrations over a wide range of silica contents (Fig. 3) but also have low Ti/Zr and Ti/V ratios (Fig. 4). Rocks of the third magmatic series (diorites, quartz diorites, basaltic andesites, andesites, dacites) form a classic low-K calc-alkaline magma series (Fig. 3). These rocks do not represent a later, superimposed volcanic arc, but are an integral part of the ophiolite stratigraphy.

Dikes and pillows of the final magma series are characterized by MORB-like major and trace element compositions: low silica, high TiO_2 that increases with silica and Zr, and high Ti/V ratios (Figs. 3 and 4). The dikes crosscut all of the older suites, and the pillows are intercalated within the Crowfoot Point Breccia, showing that this magma series represents the final magmatic event in the life cycle of the Elder Creek ophiolite.

Stonyford. The Stonyford Volcanic complex (SFVC), which lies just 35 miles south of Elder Creek, is unique within the CRO (Fig. 1). This seamount complex consists almost entirely of volcanic flows (pillow lava) with subordinate diabase, hyaloclastite breccia, and minor sedimentary intercalations of chert and limestone (Fig. 2; Shervais et al., 2003). The rocks are exceptionally fresh for the CRO, as shown by the preservation of primary igneous plagioclase, clinopyroxene, and basaltic glass in most of the volcanic rocks, and (rarely) olivine phenocrysts within volcanic glass (Shervais and Hanan, 1989). The preservation of unaltered basaltic glass shows that these rocks could not have been subducted, and therefore must have formed in the upper plate of the subduction zone.

The SFVC crops out as four large tectonic blocks (up to 5 × 3 km in areal extent) within or overlying a sheared serpentinite-matrix mélange. Structurally below the largest blocks of SFVC are dismembered remnants of CRO plutonic and volcanic rock, including dunite, wehrlite, clinopyroxenite, gabbro, diorite, quartz diorite, and keratophyre pillow lava. Quartz diorite also occurs as dikes within mélange blocks of isotropic gabbro. At other locations in the mélange, tectonic blocks include unmetamorphosed volcanogenic sandstones (correlative with the Crowfoot Point breccia near Elder Creek), foliated metasediments (possibly correlative with the Galice Formation), and pale green metavolcanic rocks (Zoglman and Shervais, 1991; Shervais et al., 2003). The occurrence of foliated Galice-like metasediments suggests tectonic transport of the CRO across Sierran basement (e.g., Jayko and Blake, 1987).

Volcanic rocks of the SFVC comprise three distinct petrologic groups: (1) oceanic tholeiite basalts; (2) transitional alkali basalts and basaltic glasses; and (3) high-alumina, low-Ti tholeiites (Zoglman, 1991). These compositionally distinct magma series are intercalated at all stratigraphic levels of the seamount. The oceanic tholeiite suite has high concentrations of TiO_2 (Fig. 3) with MORB-like Ti/Zr and Ti/V ratios (Fig. 4). The alkali-suite lavas have lower

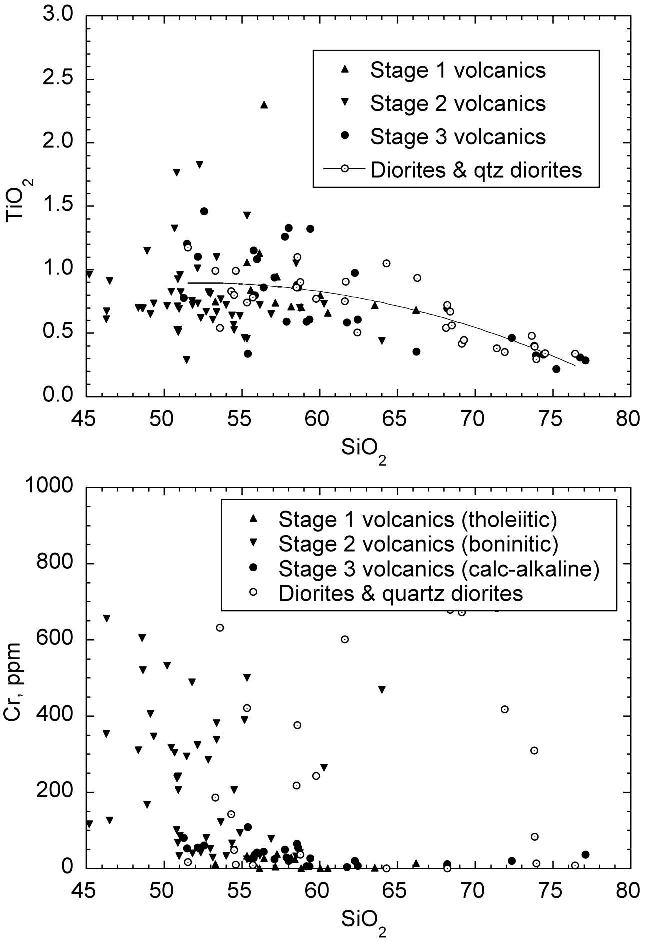


FIG. 3. Harker diagrams for TiO_2 and Cr in volcanic rocks of the Coast Range ophiolite. Upright triangles = stage 1 arc tholeiite lavas (Elder Creek, Healdsburg, Del Puerto, Cuesta Ridge, Stanley Mountain, Point Sal); inverted triangles = stage 2 refractory lavas (tholeiitic ankaramites or boninites: Elder Creek, Del Peurto, Llanada, Quinto Creek, Cuesta Ridge, Point Sal); filled circles = stage 3 calc-alkaline lavas (Elder Creek, Llanada, Cuesta Ridge); open circles = stage 4 mid-ocean ridge basalt/ocean island basalt (Elder Creek, Stonyford, Black Mountain). Data sources: Shervais and Kimbrough (1985a), Lagabrielle et al. (1986), Beaman (1989), Shervais (1990), Zoglman (1991), and Giaramita et al. (1998).

TiO_2 but higher Ti/V ratios (Fig. 4). The high-Al, low-Ti lavas have low Ti/V ratios that are typical of arc-volcanic rocks (Shervais, 1982). Initial Pb isotopic ratios (at 164 Ma) for the volcanic glasses show a wide range in $^{206}Pb/^{204}Pb$ (Hanan et al., 1991, 1992), similar to Pacific oceanic basalts currently found along the East Pacific Rise. The trace-element and Pb isotopic data indicate that the oceanic tholeiites and alkalic lavas of the Stonyford complex were derived from a heterogeneous mantle source comprising two components: depleted MORB-asthenosphere and an enriched OIB-like component (Hanan et al., 1991, 1992). The high-Al, low-Ti lavas resemble high-Al island-arc basalts; second-stage melts of MORB asthenosphere that form by melting of plagioclase lherzolite at low pressures have similar compositions (Shervais et al., 2003). The occurrence of the SFVC structurally overlying dismembered ophiolite lithologies that resemble the Elder Creek ophiolite remnant suggests that the

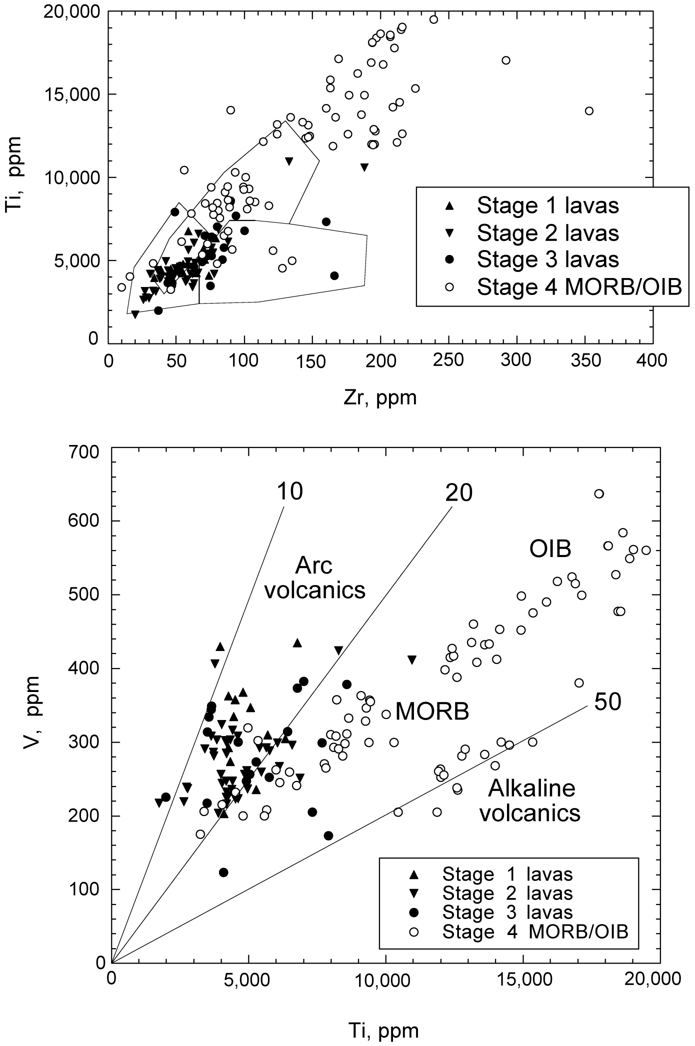


FIG. 4. Ti-Zr plot (Pearce and Cann, 1973) and Ti/V plot (Shervais, 1982) for volcanic rocks of the Coast Range ophiolite. Symbols and data sources are the same as Figure 3.

volcanic complex was built on older, subduction-related ocean crust prior to or during deformation of the older crust.

Shervais and Kimbrough (1985a, 1987) did not distinguish between the Stonyford Volcanic Complex and the Snow Mountain Volcanic Complex of MacPherson (1983). However, as noted by MacPherson and Phipps (1985), these units are distinct and unrelated. The Snow Mountain Volcanic Complex contains incipient blueschist metamorphic phases and lies west of the serpentinite belt that marks the Coast Range fault (MacPherson, 1983, 1986; MacPherson and Phipps, 1985). In contrast, the SFVC contains unaltered volcanic glass, lies entirely within the Coast Range fault serpentinite belt, and is closely associated with (but is not directly overlain by) sediments of the Jurassic basal Great Valley Series (Shervais and Hanan, 1989; Zoglman, 1991; Zoglman and Shervais, 1991). The SFVC also contains radiolarian chert intercalations with faunal assemblages that are similar to those found at other CRO locations, but distinct from faunal assemblages derived from Franciscan cherts. Thus, the Snow Mountain Volcanic Complex is clearly part of the Franciscan assemblage, whereas the SFVC is best regarded as part of the CRO (contrary to our former correlation of both units with the Franciscan assemblage, e.g., Shervais and Kimbrough, 1987; Shervais and Hanan, 1989; Shervais, 1990). This distinction is important because it has significant tectonic implications for the origin and evolution of the CRO.

Black Mountain/Geyser Peak. The Black Mountain/Geyser Peak ophiolite remnant consists largely of pillow lava and diabase dikes and sills (McLaughlin, 1978; McLaughlin and Pessagno, 1978; McLaughlin & Ohlin, 1984; McLaughlin et al., 1988). The western edge of this remnant comprises a thin strip of serpentinized peridotite and an adjacent strip of uralitic gabbro, in contact with a diabase sill complex that underlies much of Geyser Peak (McLaughlin, 1978; McLaughlin and Pessagno, 1978). A steep fault juxtaposes this sill complex against pillow lava and pillow breccia that underlie much of Black Mountain to the east. The pillow lava is overlain in turn by ophiolitic breccias containing clasts of microgabbro, diabase, diorite, and basalt, in a fine, sandy matrix (McLaughlin and Pessagno, 1978; Lagabrielle et al., 1986). These breccias are similar to the Crowfoot Point Breccia and occupy the same stratigraphic horizon.

Geochemical investigations of the Black Mountain pillow lavas show that they are MORB-like basalts that contain a minor subduction component (Lagabrielle et al., 1986; Giaramita et al., 1998). The basalts are characterized by increasing TiO_2 relative to increasing silica (Fig. 3), high Ti relative to Zr, and Ti/V ratios of ~22–30 (Fig. 4; Giaramita et al., 1998). These characteristics are nearly identical to the "oceanic tholeiite" suite in the SFVC, but without the accompanying alkali basalt and low-Ti/high-Al basalt suites. The only hint of a subduction component in these rocks are the elevated Th concentrations relative to MORB.

Mount Saint Helena, Harbin Springs. The Harbin Springs ophiolite remnant crops out about 10 miles north of Mount Saint Helena and the Mount Saint Helena ophiolite remnant (Fig. 1). Both ophiolites have the same fault-bounded assemblages: serpentinized harzburgite ± dunite or sheared serpentinite, cumulate gabbro, and gabbro/diabase breccia (Bezore, 1969; McLaughlin and Pessagno, 1978; Hopson et al., 1981; Robertson, 1990). Contacts between units are faults, and the breccia is overlain by tuffaceous sediments of the lower Great Valley Sequence. The cumulate gabbros include both ol + cpx gabbros and cpx gabbros with well-developed planar laminations and adcumulus textures, and may be cut by dikes. The breccia is dominantly clast supported, crudely stratified, with intercalations of finer-grained sand and silt of similar composition (Robertson, 1990). The breccia at these locales correlates with the Crowfoot Point breccia at Elder Creek. No geochemistry has been reported from either location.

Healdsburg. The Healdsburg ophiolite remnant comprises sheared serpentinite (mostly harzburgite), cumulate gabbro and olivine gabbro, hornblende diorite and hornblende quartz diorite, diabase dike complex, and up to 300 m of volcanic rocks (Hopson et al., 1981). Contacts are either faults or covered. The volcanic section includes both spilite (basalt) and keratophyre (andesite-dacite). Geochemical data for two samples from Healdsburg in Shervais (1990) show that these rocks range from 56% to 71% SiO_2, with low TiO_2 and Ti/V ratios of 10–15. These values are characteristic of arc volcanics, but it is not clear how they relate to one another or how well they reflect the true proportions of volcanic rocks present.

Central (Diablo Range) belt

The Central or Diablo Range belt crops out in the Diablo Range, along the western margin of the San Joaquin Valley (Fig. 1). Major CRO locales in this belt include Mount Diablo, the Leona Rhyolite, Del Puerto Canyon, Quinto Creek, and Llanada. Also included in this belt here is the recently described Sierra Azul ophiolite in the Santa Cruz Mountains, which is separated from the Diablo Range locales by the Hayward-Calaveras fault system. Ophiolites of the Diablo Range belt are typically overlain conformably by silicic tuff, tuffaceous chert, and sediments of the Great Valley Sequence (Hopson et al., 1981; Evarts et al., 1999).

Mount Diablo. The Mount Diablo ophiolite remnant consists of serpentinite, diabase sill complex, and pillow lava, juxtaposed against Franciscan mélange (Williams, 1983). Cumulate plutonic rocks are absent, and all contacts between adjacent units are faults (Williams, 1983; Hagstrum and Jones, 1998). Zircons from small keratophyre dikes yield a U-Pb age of 165 Ma (J. M. Mattinson, reported in Mankinen et al., 1991), essentially identical to U-Pb zircon dates from other CRO locales (Mattinson and Hopson, 1992).

Two geochemically distinct suites of igneous rocks can be distinguished in the Mount Diablo ophiolite remnant: an older suite of island arc basalts and andesites (keratophyres), and a younger suite of pillow lava and diabase with geochemical affinities to mid-ocean ridge basalts (Williams, 1983). The older island-arc suite volcanic rocks are cross-cut by dikes and sills of MORB-like basalt; similar cross-cutting relations are observed at Elder Creek and Del Puerto Canyon.

Leona Rhyolite. The Leona Rhyolite crops out extensively in the Berkeley Hills, forming prominent local landmarks such as Indian Rock and the Cragmont Crags. Originally mapped as a Miocene felsic volcanic unit, the Leona Rhyolite is now interpreted to correlate with the Lotta Creek Formation in Del Puerto Canyon because it rests depositionally on mafic rocks of the CRO, and is overlain depositionally by Jurassic sediments of the Great Valley Sequence (Jones and Curtis, 1991). A more robust correlation may be with felsic volcanic rocks in Del Puerto Canyon that underlie the Lotta Creek Formation, based on the absence of hornblende phenocrysts in the Leona Rhyolite (Evarts, pers. commun., 1999). Our data show that rhyolite samples from outcrops near Leona Regional Park in east Oakland consists of clear volcanic glass with small phenocrysts of plagioclase ($\approx An_{55\text{-}75}$), augite ($\approx Wo_{38}En_{33}$), hypersthene ($\approx Wo_{3.6}$ En_{48}), and ferropigeonite ($\approx Wo_{14}En_{13}$). In thin section, the glass contains domains that resemble collapsed pumice, suggesting that it may have formed as a submarine pumice-obsidian flow.

Another interesting feature of the Leona Rhyolite is the occurrence of immiscible blobs (= 3 mm in diameter) of brown basaltic glass within the clear rhyolite glass (Shervais, 1993). Electron microprobe and ion microprobe analyses of these brown glass inclusions show that they have major-element chemistry similar to mid-ocean ridge basalts of the Pacific basin, LREE-depleted trace element patterns, and no dissolved water (microprobe totals = 99.5%; Shervais, unpubl.). The brown glass inclusions may simply represent the product of liquid immiscibility; alternatively, they may reflect the mixing of MORB magmas into the rhyolite prior to eruption. If so, this implies the intersection of an active ridge segment with the CRO forearc and intrusion of mid-ocean ridge basalt magmas into magma chambers of the overlying plate (Shervais, 1993).

Del Puerto Canyon. The Del Puerto Canyon ophiolite remnant is the largest and best preserved ophiolite fragment in the Diablo Range (Evarts, 1977; Evarts and Schiffman, 1983; Evarts et al., 1999). It crops out as four large, fault-bounded blocks that together preserve a relatively complete ophiolite stratigraphy and are overlain depositionally by sediments of the Great Valley Sequence, which contain *Buchia piochii* of Tithonian age (Evarts et al., 1999). These blocks are underlain tectonically by blueschist-facies rocks of the Franciscan assemblage. Despite its relatively complete stratigraphy (Fig. 2), much of the plutonic section appears to be missing, along with portions of the volcanic section (Evarts et al., 1999).

Four magmatic series are evident, especially in the plutonic section of the ophiolite. The oldest is represented by cumulate dunites and gabbros; associated volcanic rocks may include low-K, LREE-depleted massive flows in the volcanic section, but there is no clear distinction between "early" lavas and "late" lavas as there is at Point Sal. The second magmatic series is represented by pyroxenite, wehrlite, feldspathic peridotite, and some gabbronorites, and by sparse high-Mg andesites with "boninitic" affinities (high Mg, Cr, and Ni relative to Si; high Cr spinels; Shervais, 1990; Evarts et al., 1992, 1999). The third magmatic series comprises a calc-alkaline suite of pyroxene and hornblende gabbro, hornblende diorite, hornblende quartz diorite, keratophyre dikes and sills, and calc-alkaline volcanic rocks ranging from basalt to rhyolite in composition. The fourth magmatic suite observed here comprises rare basaltic dikes with MORB geochemistry ($Ti/V \approx 30$, $La_n/Yb_n \approx 0.5$, $Nb \approx 8$ ppm; Evarts et al., 1992). Tuffaceous radiolarian cherts, tuffs, and volcanic breccias of the Lotta Creek Formation seem to postdate the MORB dikes, and may represent volcanic-arc activity unrelated to the ophiolite.

Age relations in the Del Puerto ophiolite remnant are based on a range of techniques and rock types. Hornblende gabbro has been dated at ≈157 Ma ($^{40}Ar\text{-}^{39}Ar$ hornblende; Evarts et al., 1992), and a plagiogranite dike that crosscuts quartz diorite has the same age (≈157 Ma, U-Pb zircon; Hopson et al., 1981). Felsic dikes that cut cumulates, rhyolite lava, and andesite boulders in a Lotta Creek Formation conglomerate bed have all been dated at ≈150 Ma (zircon fission track and $^{40}Ar\text{-}^{39}Ar$ hornblende; Evarts et al., 1992). These are the youngest ages yet reported from the CRO and overlap in age with basal Knoxville strata of the Great Valley Sequence; they are clearly much younger than other CRO locales (e.g., Point Sal, ≈165–173 Ma; Mattinson and Hopson, 1992).

The Del Puerto ophiolite remnant bears many similarities to the Elder Creek remnant in that: (1) the first magma series is represented primarily within the plutonic section; (2) the third magmatic series dominates the volcanic section and is well represented by dikes, sills, and sill-like plutons in the plutonic section; and (3) the fourth magmatic event, which may have overlapped with stage 3 volcanism, is represented by rare basaltic dikes with MORB-affinity. But Del Puerto differs from other CRO locales in several important ways as well: (1) the first magmatic suite is low-K calc-alkaline rather than arc tholeiite; (2) Del Puerto extends to younger ages than other well-dated CRO remnants; and (3) calc-alkaline volcanism seems to have continued (or restarted) after the late MORB dikes, as shown by the hornblende-phyric Lotta Creek tuff (Evarts et al., 1999). The extended magmatic history of this ophiolite remnant may be related to late Jurassic arc activity in the western Sierran foothills, or may reflect variations in plate margin geometry.

Quinto Creek. This small ophiolite remnant comprises serpentinized harzburgite, cumulate and noncumulate gabbro; quartz gabbro/diorite; plagiogranite dikes;, a sheeted complex of diabase, basalt, andesite, and microdiorite; and a volcanic complex of pillowed and massive basalts (Bailey et al., 1970; Hopson et al., 1981). Volcanic breccia horizons are intercalated with the pillows, similar to the volcanic breccias at Cuesta Ridge (Robertson, 1989). Limited geochemical data suggest that the volcanic rocks here are borderline between island-arc lavas and MORB (Shervais and Kimbrough, 1985a; Shervais, 1990), but the data are insufficient to identify distinct suites or relationships.

Llanada. The Llanada ophiolite remnant was originally mapped by Enos (1963) and was later mapped in detail by Emerson (1979). It has been studied in reconnaissance by many investigators, including Bailey et al. (1970), Hopson et al. (1981), Robertson (1989), Lagabrielle et al. (1986), and Giaramita et al. (1998). The Llanada remnant contains the following members: (1) cumulate plutonic rocks, including gabbro, olivine gabbro and melagabbro, feldspathic wehrlite, and possibly dunite or harzburgite; (2) a lower volcanic member consisting of massive and pillowed lava flows; and (3) an upper volcanic member consisting of siliceous tuffs, basaltic breccias, and massive, coarsening upwards laharic breccias and tuff breccias composed mainly of andesite (Fig. 2; Emerson, 1979; Hopson et al., 1981; Lagabrielle et al., 1986; Robertson, 1989).

The upper volcanic section at Llanada corresponds to a calc-alkaline volcanic arc sequence, grading upwards from distal to proximal in depositional facies; this is similar to the third magmatic series observed at other CRO localities (e.g., Elder Creek, Del Puerto). The lower volcanic series of massive and pillowed basalt has the geochemical characteristics of supra-subduction zone volcanism: low TiO_2; low Ti/V ratios; high Th, Rb, and Ba; and REE abundances that range from slightly LREE depleted to slightly LREE enriched (Shervais and Kimbrough, 1985a; Lagabrielle et al., 1986; Shervais, 1990; Giaramita et al., 1998). Although not true boninites, these lavas are also enriched in Cr and Ni relative to normal arc tholeiites, and contain Cr-spinels that have Cr/Al ratios higher than typical MORB spinels (Giaramita et al., 1998). Thus these lavas, along with the feldspathic wehrlites of the plutonic section, may correspond to the second magmatic series observed at other CRO localities (e.g., the upper pillow lavas and wehrlites at Point Sal (see below), and the wehrlite intrusions at Elder Creek). There is no clear evidence for a typical "arc tholeiite" series that might correspond to the lower pillow lavas or the early cumulate gabbros at Point Sal, although the Llanada cumulate gabbros may represent this magma series. There is also no indication of a later MORB-like magma series.

Sierra Azul. This recently discovered ophiolite locality in the Santa Cruz Mountains near Los Gatos contains a nearly complete ophiolite stratigraphy, with cumulate ultramafic rocks, gabbros, diorites, quartz diorites, andesites, massive basalt flows, and overlying andesite/quartz keratophyre tuffs and breccias (McLaughlin et al., 1991, 1992). Intrusive relations of the plutonic rocks have not been described in detail, so it is not possible to tell how many magmatic series are present. Almost all of the volcanic rocks and dike rocks have arc-like geochemical features (low TiO_2, high SiO_2) but a few are apparently MORB-like in character (McLaughlin et al., 1991, 1992). Overall, this ophiolite remnant appears similar to the occurrences found at Elder Creek and Del Puerto Canyon.

Southern (Coastal) belt

The Southern or Coastal belt of the CRO is separated from the other belts by the San Andreas and Sur-Nacimiento fault systems (Fig. 1). There are three main ophiolite locales (Cuesta Ridge, Stanley Mountain, and Point Sal) plus several smaller remnants not described here (Marmalejo Creek, San

Simeon, and Santa Cruz Island; Hopson et al., 1981). All of these ophiolites were likely derived from locations much farther to the south, based on the aggregate offset along these fault systems.

Cuesta Ridge. The Cuesta Ridge ophiolite (Page, 1972; Pike, 1974; Snow, 2002) is exposed in a large, open syncline along Cuesta Ridge, just north of San Luis Obispo. It consists of harzburgite ± dunite tectonite, cumulate mafic and ultramafic rocks, and up to 1.2 km of pillow lava and breccias. Layered gabbros are rare; olivine clinopyroxenites, wehrlites, and lherzolites with interstitial hornblende and plagioclase are more common. The dominant plutonic rock is quartz diorite, which forms the entire sheeted sill complex mapped at the east end of Cuesta Ridge and a 400 m thick high-level intrusive complex (quartz gabbro, quartz diorite, keratophyre, and quartz diabase sills 5–10 m thick) near Cerro Alto Peak (Snow, 2002).

Arc tholeiite volcanics are present low in the volcanic sequence but are less common than other lavas. More common are volcanic rocks with boninitic affinities (high Mg, Cr, Ni, and silica contents) that are intercalated with calc-alkaline–series volcanics up-section. These include massive volcanic flows, pillow lavas, and breccias with basaltic andesite to dacite compositions (Snow, 2002). Late-stage dikes of olivine tholeiite that crosscut the quartz diorite sill complex have major and trace element composition similar to MORB (e.g., ≈ 1.1 to 2.2 wt% TiO_2); late pillow lavas that crop out just below the chert in some places have similar MORB-like compositions (Snow, 2002).

Stanley Mountain. The Stanley Mountain ophiolite remnant crops out along Alamo Creek in southern San Luis Obispo County (Fig. 1). This ophiolite comprises a thick section of pillow lava and volcanic breccias overlain by more than 100 m of chert, tuffaceaous chert, and mudstone, in turn overlain conformably by basal sediments of Great Valley Series (Hopson et al, 1981; Robertson, 1989). The radiolaria assemblages from the cherts and mudstones have been described in great detail by Hull (1995, 1997), and by Hull and Pessagno (1994, 1995). A paleomagnetic study Hagstrum and Murchey (1996) showed that these cherts did not form south of the equator, but relatively close to their present latitude (see also Murchey and Hagstrum, 1997). Little information is available on the chemistry of the underlying pillow lava; Shervais (1990) published one whole-rock analysis: it is an andesite with low Mg, Cr, and Ni, and low Ti/V ≈ 12, i.e., a fairly typical low-K arc tholeiite.

Point Sal. The Point Sal ophiolite remnant is one of the best exposed and most thoroughly studied of all the CRO fragments (Hopson and Frano, 1977; Hopson et al., 1981; Mattinson and Hopson, 1992; Fig. 1). Coastal exposures of the plutonic section reveal mantle harzburgite, dunite, layered and in some cases foliated cumulate gabbros, intrusive wehrlite-clinopyroxenite bodies that crosscut layering in the gabbros, isotropic high-level gabbros, diorites and quartz diorites with miarolytic cavities that form arrays of vertically oriented pipes, thin sheets of plagiogranite, and a sheeted sill complex with basaltic, diabase, microdiorite, and epidosite sills and dikes that penetrate the base of the volcanic section (Fig. 2). The volcanic section consists of a lower volcanic zone of augite-plagioclase microphyric basaltic-andesite to andesite, and an upper volcanic zone of olivine-augite microphyric basalt. The upper volcanic zone is overlain by a condensed sequence of grey-green to brown tuffaeous radiolarian cherts.

All of the volcanic rocks at Point Sal have trace element systematics characteristic of supra-subduction zone basalts (Figs. 3 and 4; Shervais and Kimbrough, 1985a; Shervais, 1990). The lower volcanic zone lavas are arc tholeiites, whereas the upper volcanic zone lavas have high-Ca boninitic affinities. Based on phenocryst assemblages in the volcanic rocks and cumulate phase assemblages in the plutonic rocks, the lower pillow lavas correlate with the early cumulate gabbros, whereas the upper pillow lavas correlate with the intrusive wehrlite-clinopyroxenites. These magma groups correspond to the first and second magma series observed at Elder Creek. High-level diorites, microdiorites, and plagiogranites are rare, and there is no evidence for a final, MORB-like lava series.

Southwest Oregon

Ophiolite remnants in the Wild Rogue Wilderness area have been correlated with the CRO of California (Gray et al., 1982; Saleeby et al., 1984; Blake et al., 1985; Kosanke et al., 1999; Kosanke, 2000). These rocks crop out west of the Rogue-Chetco arc complex and its back-arc basin, the Josephine ophiolite complex (Harper, 1984; Harper et al., 1985, 1994), and structurally overlie rocks of the Franciscan Complex. The structural position of these remnants overlying Franciscan units, their stratigraphic position below Great Valley equivalent rocks (Riddle Formation, Days Creek Formation), and their lack of Nevadan deformation or metamorphism all support their correlation with the Coast Range ophiolite in California (Saleeby, 1984; Blake et al., 1985; Kosanke et al., 1999). A similar ophiolite remnant, the Snow Camp Mountain ophiolite, appears to represent an outlier of the Josephine ophiolite (Schoonmaker et al., 2003).

The Wild Rogue Wilderness ophiolite remnant has been studied in detail by Kosanke, from whom this description has been abstracted (Kosanke et al., 1999; Kosanke, 2000). Cumulate plutonic rocks are found only as screens in the sheeted complex, but isotropic gabbros, diorites, tonalites, and volcanic rocks are well represented. The volcanic rocks include pillow lava and the overlying Mule Mountain metavolcanics, which consists of basalt/basaltic andesite flows in its lower part and andesite/dacite flows in its upper part. The pillow lavas are LREE-depleted basaltic andesites with island-arc tholeiite affinities, similar to the first magma series at Elder Creek, Point Sal, and other CRO locales in California. Basalts in the lower part of the Mule Mountain metavolcanics include pyroxene porphyries with high Cr contents similar to high-Ca boninites (e.g., Crawford, 1989); these rocks correspond to the second magma series at Elder Creek and Point Sal. Andesites and dacites in the upper part of the Mule Mountain metavolcanics are calc-alkaline lavas and volcaniclastics similar to those found a Llanada and Del Puerto; these are inferred to correspond to the third magma series at Elder Creek and other California CRO locales.

Similar groupings are found in the hypabyssal rocks: isotropic gabbros and diabase dikes have transitional MORB/IAT compositions that correspond to the earliest magma series, whereas felsic diorites include both high-Cr varieties that correspond to the refractory magma series and hornblende gabbros and diorites that appear to correspond to later calc-alkaline magmas (Kosanke, 2000). Where cross-cutting relations can be observed, the diabase dikes are always older than the dioritic dikes. Two dikes with MORB-like compositions were described by Kosanke, but their timing is not clear. The latest dikes are garnet-muscovite tonalites with "S-type" granite affinities that are similar to late biotite granites in the Semail ophiolite (Lippard et al., 1986).

Age of the Coast Range ophiolite

Age relations in the CRO are critical for understanding the tectonic evolution of the Western Cordillera, especially Jurassic orogeny in the Sierra Nevada and Klamath mountains. Evolution of the CRO is closely tied in space and time to arc volcanism, deformation, and fabric development in the western Sierra foothills terranes. A number of models for Jurassic orogeny in the foothills metamorphic belt are specifically linked to the igneous age of the CRO and subsequent biostratigraphic ages of overlying sediments (e.g., Hopson et al, 1981; Dickinson et al., 1996).

Published ages for the Coast Range ophiolite are summarized in Table 2. These dates can be evaluated in three principal groups: (1) K-Ar dates on igneous hornblendes from gabbros and diorites (Lanphere, 1971; McDowell et al., 1984); (2) older U-Pb zircon dates, which predate high-precision, multi-collector mass spectrometers (Hopson et al., 1981); and (3) newer U-Pb zircon ages, which are generally significantly older than both the K-Ar ages and the older U-Pb zircon ages (J. M. Mattinson reported in Mankinen et al., 1991 and in Pessagno et al., 1993; Evarts et al., 1992; Mattinson and Hopson, 1992). Unfortunately, almost all of the more recent ages are reported either in abstracts (Mattinson and Hopson, 1992; Evarts et al., 1992) or as personal communications in other papers (J. M. Mattinson, reported in Mankinen et al., 1991, and Pessagno et al., 1993), with no supporting data and in some cases, no assignment to specific locations within the ophiolite. As a result, these ages are of limited value at this time.

The data suggest that CRO formation began circa 172 Ma and ended in most areas by ~160–164 Ma. Volcanism persisted until ~150 Ma in the Del Puerto ophiolite, possibly in response to progradation of arc volcanics from the Sierra foothills. Early K-Ar ages are now considered unreliable in light of pervasive low-temperature metamorphism, whereas the older U-Pb zircon ages (Hopson et al., 1981) are considered to be too young by 5–10 million years (Mattinson and Hopson, 1992).

Discussion

Multi-stage, supra-subduction zone origin for the Coast Range ophiolite

It now seems relatively well established that the CRO of California formed in a fore-arc setting,

TABLE 2. Summary of Published Age Dates for the Coast Range Ophiolite[1]

Locale	Rock type	Age	System	Reference
Elder Creek	Hb gabbro	154 ± 5[2]	K-Ar Hb	Lanphere, 1971
	Hb gabbro	166 ± 3, 143–144 ± 3	K-Ar Hb	McDowell et al., 1984
Wilbur Springs	Hb gabbro	143–144 ± 3	K-Ar Hb	McDowell et al., 1984
Harbin Springs	Plagiogranite	169[3]	*U-Pb zircon*	*J. M. Mattinson, reported in McLaughlin and Ohlin, 1984*
Healdsburg	Plagiogranite	163 ± 2	U-Pb zircon	Hopson et al., 1981
Mt Diablo	Plagiogranite	**165**[3]	*U-Pb zircon*	*J. M. Mattinson, reported in Mankinen et al., 1991*
Del Puerto (Red Mtn.)	Hb gabbro	162 ± 5[2]	K-Ar Hb	Lanphere, 1971
	Hb peridotite	164 ± 5[2]	K-Ar Hb	Lanphere, 1971
	Plagiogranite	154–157 ± 2	U-Pb zircon	Hopson et al., 1981
	Hb gabbro	**157**[3]	*^{40}Ar-^{39}Ar Hb*	*Evarts et al., 1992*
	Rhyolite, andesite	**150**[3]	*^{40}Ar-^{39}Ar, fission track*	*Evarts et al., 1992*
Llanada	Hb albitite	163–165 ± 2		Hopson et al., 1981
Cuesta Ridge	Plagiogranite	152–153 ± 3	U-Pb zircon	Hopson et al., 1981
Stanley Mtn	Plagiogranite	**166**[3]	*U-Pb zircon*	*J. M. Mattinson, reported in Pessagno et al., 1993*
Point Sal	Plagiogranite	160–162 ± 2	U-Pb zircon	Hopson et al., 1981
	Plagiogranite	**165-173**[3]	*U-Pb zircon*	*Mattinson and Hopson, 1992*
Santa Cruz Island	Plagiogranite	161–167 ± 2	U-Pb zircon	Hopson et al., 1981
	Plagiogranite	144–148 ± 2	U-Pb zircon	Hopson et al., 1981

[1]Older K-Ar dates recalculated by McDowell et al., 1984. Most of the recent (post-1990) dates are reported in abstracts or as personal communications, with no supporting data or uncertainties (references shown in italics). Dates in bold are considered to be the most reliable at this time (dates based on modern U-Pb zircon work [since 1990].
[2]Ages recalculated by McDowell et al., 1984, using new decay constants.
[3]Ages reported in abstracts or as personal communication in unrelated paper, with no supporting data.

probably above the east-dipping proto-Franciscan subduction zone (e.g., Evarts, 1977; Shervais and Kimbrough, 1985a, 1985b; Lagabrielle et al., 1986; Robertson, 1989, 1990; Shervais, 1990, 2001; Shervais and Beaman, 1989, 1991, 1994; Stern and Bloomer, 1992; Giaramita et al., 1998). Three distinct magmatic suites can be found in many remnants of the Coast Range ophiolite, all of which formed in this supra-subduction zone setting; these suites and their occurrence are summarized in Table 1.

The first magmatic suite includes dunites, layered and foliated cumulate gabbros, some isotropic gabbros, portions of the sheeted complex, and a lower or older volcanic series (arc tholeiite or, less commonly, low-K calc-alkaline), where these are preserved. Volcanic rocks of this suite are characterized by low Ti and Cr over a range of silica contents (dominantly basaltic andesite to andesite in composition; Fig. 3), and low Ti/Zr and Ti/V ratios (Fig. 4). The second magmatic suite includes intrusive wehrlites, lherzolites, clinopyroxenites, primitive gabbros, and gabbronorites in the plutonic sections, portions of the sheeted complex, and an upper volcanic series with high-Ca boninitic affinities. The volcanic rocks of this suite are characterized by high Cr over a range of silica contents (Fig. 3), and by low TiO_2, Ti/Zr, and Ti/V ratios (Fig. 4). The third magmatic suite includes hornblende diorites, quartz diorites, plagiogranites, diorite-matrix agmatites, portions of the sheeted complex, and the volcaniclastic cover ("non-ophiolite volcanics") of andesite to rhyolite composition (Fig. 3) that rests on the underlying massive pillow lavas. The third magmatic suite is calc-alkaline in character and is often not included with the "true" ophiolite magma series, even though all three lie beneath the radiolarian cherts and clastic sediments that define the upper surface of the ophiolite.

The three magmatic episodes described above correspond to the three progressive stages in ophiolite formation as defined by Shervais (2001). These include: (1) birth—formation of incipient island-arc crust above a nascent subduction zone (e.g., Casey and Dewey, 1984; Hawkins et al., 1984; Stern and Bloomer, 1992); (2) youth—rifting and deformation of this crust, which allowed the intrusion of primitive boninitic magmas; and (3) maturity—transition to more normal calc-alkaline magmatism. Application of this model to the CRO is depicted schematically in Figure 5.

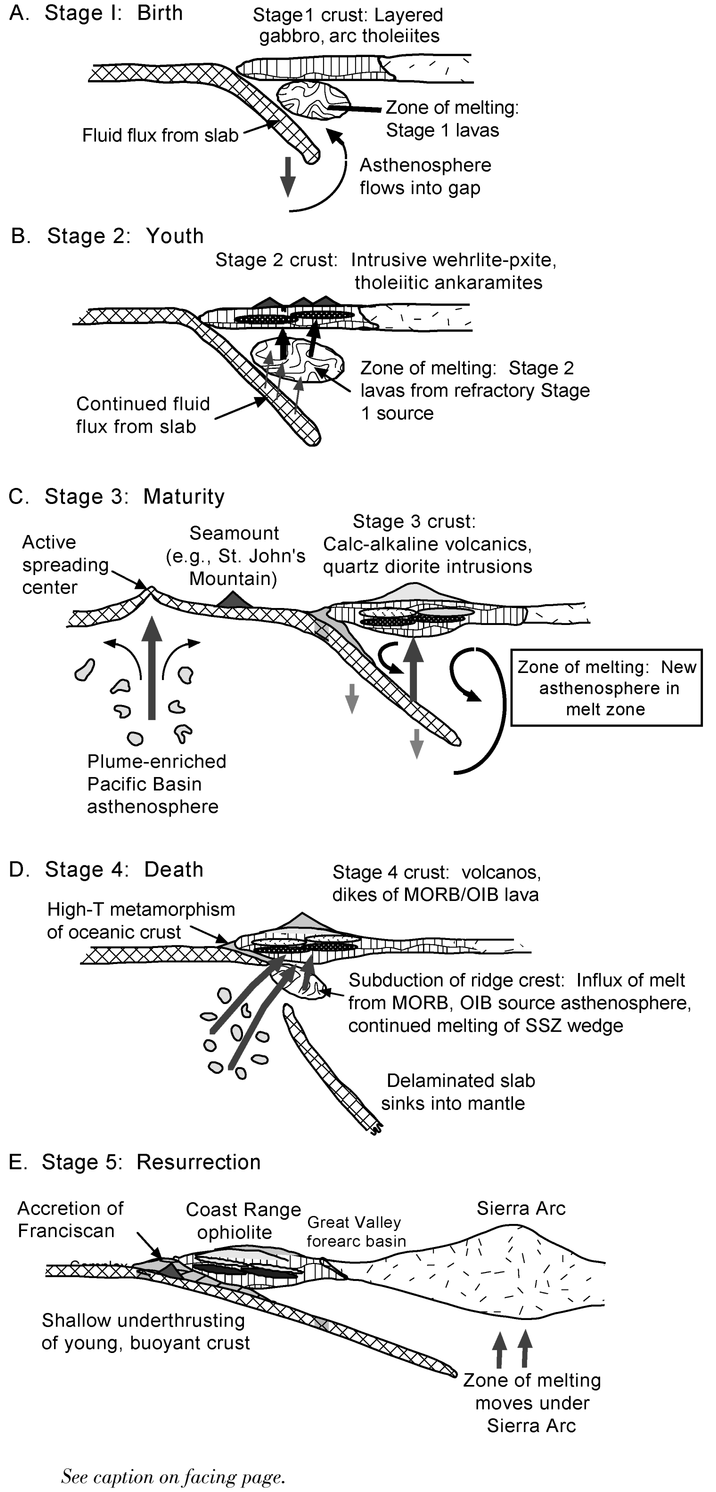
These three stages and their associated magma series form the classic ophiolite stratigraphy and comprise the bulk of all supra-subduction zone ophiolites. These stages and their associated magma series always form in the same relative order, and if evidence for one magma series is missing at any given locale, the other series still exhibit the same order of formation. They also seem to occur in almost all supra-subduction zone ophiolites. Shervais (2001) interpreted this sequence in terms of a model of progressive ophiolite formation, starting with a nascent subduction zone and progressing toward a stable calc-alkaline arc. Ophiolite formation may be terminated by collision with a spreading center, but this is not always apparent. Subsequent emplacement of the ophiolite may occur by obduction onto a passive continental margin (Tethyan ophiolites) or by accretionary uplift (Cordilleran ophiolites; Shervais, 2001).

Ridge collision: Death of an ophiolite

The fourth magma series, found in only a few CRO locales, represents the influx of true oceanic basalt magmas to form dikes and pillows that crosscut or overlie all of the older igneous rock series (Fig. 2; Table 1). Evidence for this late, oceanic basalt event includes late dikes of MORB composition that intrude older calc-alkaline intrusives or volcanics (Elder Creek, Mount Diablo, Del Puerto Canyon, Sierra Azul, Cuesta Ridge), lavas of MORB or oceanic basalt composition that overlie older subduction-related igneous suites (Stonyford volcanic complex, Geyser Peak/Black Mountain, Mount Diablo, Cuesta Ridge), intercalation of oceanic basalts with more arc-like lavas (Stonyford volcanic complex), and inclusions of MORB glass in rhyolite glass (Leona Rhyolite). There are three possible interpretations for these data: (1) the change to MORB-like chemistry represents continued rifting of the arc to form a backarc basin in which lavas have more oceanic compositions (e.g., Shervais and Kimbrough, 1985a, for the Mount Diablo ophiolite); (2) the propagation of a pre-existing backarc basin spreading center into the forearc (e.g., Harper, 2003); or (3) collision of the subduction zone with an active spreading center, which leaks oceanic magmas through the overlying plate (Shervais, 1993).

We favor the third explanation here (Fig. 5). Evidence for ridge collision includes the abrupt nature of the change from calc-alkaline to oceanic lavas, the intercalation of plume-enriched arc lavas with oceanic tholeiites and alkali basalts at Stonyford, the forearc location of the ophiolite (which would place the ophiolite directly over the mid-ocean ridge when it collides), and the presence of mafic crust under the Great Valley Sequence that is too thick to represent normal backarc basin crust (e.g., Godfrey and Klemperer, 1998). Further evidence comes from accreted terranes in the Klamath Mountains, the Sierra Nevada, and the Franciscan Complex, where formation ages and oceanic residence times show a reversal in the early Late Jurassic, indicating that a spreading center was consumed at that time (Murchey and Blake, 1993).

Ophiolite locales such as Black Mountain may represent "captured" segments of the subducting ridge axis, where the subtle geochemical overprint of subduction enrichment (e.g., elevated Th concentrations) could result from the backflow of subduction-zone fluids through a slab window (Casey and Dewey, 1984). Similar scenarios may apply to other ophiolite fragments in the CRO, where no clear arc-related basement can be demonstrated. In other cases (e.g., Mount Diablo, Stonyford) the demonstrated pre-existence of arc-related crust implies that the ophiolite remnant formed in the upper plate of the subduction zone, but was penetrated by magmas from the ridge axis. This seems to be the case for most CRO locales, where oceanic basalts



See caption on facing page.

FIG. 5. Schematic diagrams showing model for the development of the Coast Range ophiolite by a five-stage process, as recorded in the magmatic and tectonic record (after Shervais, 2001). These stages include the following. A. Stage 1, "Birth" = formation of the oldest magma series (arc tholeiite) in a nascent subduction zone during hinge rollback of the sinking slab. B. Stage 2, "Youth" = formation of a refractory magma series by remelting the mantle asthenosphere depleted during stage 1 melting, in response to fluid flux from the sinking slab. C. Stage 3, "Maturity" = onset of stable calc-alkaline volcanism as the subduction zone matures. D. Stage 4, "Death" = end of ophiolite formation caused by ridge subduction, and seen as oceanic magmas overlying the older ophiolite litholigies. E. Stage 5, "Resurrection" = uplift and emplacement of the ophiolite in response to accretionary uplift.

(MORB, alkali) intrude or overlie older arc-related basement. These locales include Elder Creek, Stonyford, Mount Diablo, the Berkelely Hills (Leona Rhyolite), Del Puerto Canyon, Sierra Azul, and Cuesta Ridge—that is, most of the known CRO remnants.

Collision with an active spreading ridge would likely end ophiolite formation in most locations because of the resulting change in relative plate motions, and because the young lithosphere adjacent to the spreading center is too bouyant to sink into the mantle (Cloos, 1993). As a result, continued subduction would take place at an angle too low to support magma formation in the forearc, and magmatic activity would shift inland, toward the volcanic arc, after a hiatus to establish normal subduction. The change from "hinge-rollback"–driven spreading (e.g., Stern and Bloomer, 1992) to shallow underthrusting of hot oceanic asthenosphere would initiate high-temperature metamorphism of the upper oceanic crust in the thrust zone, and compressional deformation of the forearc region.

As noted by Cloos (1993, p. 733), "The subduction of spreading ridges will cause vertical isostatic uplift and subsidence of as much as 2 to 3 km in the forearc region compared to when 80 m.y.-old oceanic lithosphere is subducted. The subduction of an active spreading center causes such a major perturbation in the margins thermal structure that evidence of the event is likely to be recorded widely in the geology of the forearc block." Thus, it may be no accident that high-grade metamorphic blocks in the Franciscan Complex in northern California (garnet amphibolite, eclogite) have maximum ages (≈160–162 Ma; Coleman and Lanphere, 1971; Mattinson, 1986, 1988; Ross and Sharp, 1986) that correspond to a ridge-subduction event (≈164 Ma) in this area, and not to initial ophiolite formation (~172 Ma). Models that link formation of the high-grade blocks to subduction initiation are not supported by the age data, which show that initiation of CRO formation began ~173 Ma and ended about 164–162 Ma.

Eclogite-facies metamorphism appears to overprint an older amphibolite-facies assemblage in many high-grade blocks (Moore and Blake, 1989); this would result from the cooling and continued subduction of amphibolites formed at hotter, shallower conditions. In addition, many CRO localities (especially those in the Northern/Sacramento Valley Belt) show evidence for uplift, high-angle faulting, and erosion prior to deposition of the overlying Great Valley Sequence (Hopson et al., 1981; Phipps and MacPherson, 1992).

The presence of oceanic magmas as the final magmatic event in ophiolite formation is not unique to the CRO. The Semail ophiolite (Oman) is underlain by garnet amphibolites with oceanic provenance that are only slightly younger than the third magmatic suite (diorites, plagiogranites), and is overlain by oceanic alkali basalts (Salahi volcanics) that reflect leakage of the oceanic spreading center through the overlying plate (Shervais, 2001).

Accretionary uplift as an alternative to obduction

One consequence of ophiolite formation in the upper plate of a subduction zone is that emplacement of the ophiolite onto a continental margin may occur by obduction (Gealey, 1977; Coleman, 1981). Obduction occurs when the ophiolite subduction zone attempts to subduct a passive continental margin (Gealey, 1977; Coleman, 1981; Pearce et al., 1981; Searle and Cox, 1999). Thrusting of the ophiolite over continental crust is terminated by the buoyant rise of the continental lithosphere, which lifts the ophiolite and exposes it subaerially (Gealey, 1977).

An alternative to obduction that may expose ophiolites is the process that has been defined previously as "accretionary uplift" (Shervais, 2001). Accretionary uplift occurs in response to continued subduction, regardless of whether or not a ridge collision event occurs. If a passive continental margin is not encountered, a thick accretionary complex will form and the ophiolite will be exposed by progressive uplift as the accretionary complex thickens.

When the accretionary complex becomes overthickened and gravitationally unstable, unroofing and exhumation may occur by extensional denudation (e.g., Platt, 1986; Jayko et al., 1987), by rapid erosion (Ring and Brandon, 1999), or both.

This is the process that we believe is responsible for uplift and emplacement of the CRO. Formation of the Franciscan complex throughout the late Mesozoic and early Paleogene underplated and thickened crust beneath the CRO, causing uplift and extensional deformation of the ophiolite (Jayko et al., 1987; Harms et al., 1992). Relatively early exhumation of the ophiolite is documented by the occurrence of sedimentary serpentinite debris flows in the lower Great Valley sequence (Moiseyev, 1970; Phipps, 1984). After subduction ceased in the Miocene, deformation was dominated by wrench faulting, compression, and tectonic wedging of Franciscan rocks under the CRO/Great Valley section (Wentworth et al., 1984; Glen, 1990; Blake et al., 1992; Unruh et al., 1995).

Implications for Jurassic tectonics of the Western Cordillera

It has been suggested that the CRO formed in a backarc basin behind Middle Jurassic arc rocks of the western Sierra Foothills belt and was emplaced during an arc-arc collision (the Late Jurassic Nevadan orogeny, ≈155 Ma) that resulted in formation of a new, proto-Franciscan subduction zone to the west (e.g., Schweikert and Cowan, 1975; Dickinson et al., 1996; Godfrey and Klemperer, 1998). These models are based on a view of Jurassic orogeny in the Cordillera that is no longer valid (e.g., Wright and Fahan, 1988). It is now known that orogenic activity during the Jurassic involved a complex series of events that began in the Early to Middle Jurassic and continued into the Late Jurassic, ending largely by ≈145 Ma (Harper and Wright, 1984; Wright and Fahan, 1988; Saleeby, 1982, 1983, 1990; Tobisch et al., 1989; Wolf and Saleeby, 1995). Many of the structures and fabrics attributed to Nevadan deformation in the Sierra Foothills are now known to be early Middle Jurassic in age (pre-168 Ma, post-190 Ma) and record deformation that occurred prior to formation of the CRO (Edelman et al., 1989; Edelman and Sharp, 1989; Saleeby, 1990). Girty et al. (1995) have shown that Nevadan age deformation took place within the arc, and that the forearc region lay to the west of this arc. They also suggest that a new subduction zone may have formed in the early Middle Jurassic (≈174 Ma) in response to an arc collision or collapse of a fringing arc against the continental margin (Girty et al., 1995). This is slightly older than the oldest ophiolitic rocks that have been dated in the CRO, and slightly younger than the initial opening of central Atlantic at about 175–180 Ma (Klitgord and Schouten, 1986).

Geochemical data show that the stage 2 volcanics found at many CRO locales are related to high-Ca boninites (Figs. 3 and 4), which are characteristic of forearc volcanism during brief episodes of rapid spreading. The occurrence of wehrlite intrusives (which represent ultramafic cumulates of high-Ca boninite magmas) into the older layered gabbros (arc tholeiite cumulates), and deformational fabrics within the older layered gabbros, imply that the CRO formed initially by rapid spreading in the forearc, as proposed by Stern and Bloomer (1992). In this model, extensions deformation of the older arc gabbros is common prior to intrusion of the later ultramafic cumulates (Shervais, 2001).

Ward (1995) has shown that subduction was not continuous during the Jurassic, as assumed previously, but varied in response to changes in plate motions and convergence. He has shown that both relative and absolute plate motions changed significantly at the beginning of the Late Jurassic—coincident with the ridge subduction event postulated here, and just prior to the Nevadan orogeny. Plate motion studies (Ward, 1995) and structural analysis of dike swarms in the Foothills terrane (Wolf and Saleeby, 1995) both show a change from relative convergence of North America and plates of the Pacific basin in the middle Jurassic to left-lateral transtension (in the upper plate) during the Late Jurassic. Ward (1995) suggested that this change must coincide with a ridge "collision" event.

In summary, the model presented here is consistent with our current understanding of tectonic evolution in the Western Cordillera during the mid- to Late Jurassic. Models that call for arc collision during the late Jurassic and formation of the CRO in a backarc basin are not compatible with these data, or with the chemical and petrologic characteristics of the CRO.

Implications for the origin and evolution of ophiolites

The data presented here show that the Coast Range ophiolite of California followed a consistent progression during its formation over a primary linear extent of at least 1300 km, from Elder Creek in the north to Point Sal in the south (when Neogene motion on the San Andreas fault system is restored), or 1600 km if the southwest Oregon remnants are included. This progression in magmatic evolution, from primitive arc tholeiite to boninitic to calc-alkaline, is the same progression followed by other ophiolites worldwide (Shervais, 2001). The global occurrence of this consistent progression implies that ophiolite formation is not a stochastic event, but is a natural consequence of the tectonic setting in which ophiolites form. The first three stages all form in a supra-subduction zone setting, that is, in the upper plate of a subduction zone (Shervais, 2001). The most likely setting appears to be subduction initiation and hinge rollback (Stern and Bloomer, 1992), but other settings are possible. In any event, models of ophiolite formation that call on unusual combinations of circumstances, or on non-uniformitarian interpretations of ocean crust formation, are not compatible with this observed progression.

The termination of ophiolite formation by ridge collision is a natural consequence of ophiolite formation above a subduction zone whenever the plate being consumed is actively generated at a spreading center and the convergence rate exceeds the spreading rate (Shervais, 2001). This termination will be diachronous unless the ridge is parallel to the trench. Evidence for ridge collision may include the intrusion of oceanic lavas into the ophiolite complex, the eruption of oceanic lavas on top of the ophiolite complex, or the occurrence of high-grade metamorphic soles with oceanic affinities. The occurrence of high-grade metamorphic soles in which the protoliths are oceanic basalts is common to many ophiolites, suggesting that ridge collision is more common than the current literature suggests.

Finally, the variations observed among different ophiolite remnants in the CRO shows that concentrating our collective efforts on studying a few occurrences within any given ophiolite, and ignoring those that are less well exposed, obscures the insights that can be gained by considering all ophiolite occurrences in sufficient detail to understand their fundamental characteristics and evolution. Indeed, focusing on one or two classic occurrences would lead to significant misconceptions regarding how the CRO formed and evolved. The CRO of California shows how detailed investigations of many ophiolite occurrences are needed to understand better the origin and evolution of all ophiolites.

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