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Birth, death, and resurrection: The life cycle of suprasubduction zone ophiolites

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[1] **Abstract:** Suprasubduction zone (SSZ) ophiolites display a consistent sequence of events during their formation and evolution that suggests that they form in response to processes that are common to all such ophiolites. This sequence includes the following: (1) birth, which entails the formation of the ophiolite above a nascent or reconfigured subduction zone; this stage is typically characterized by the eruption of arc tholeiite lavas and the formation of layered gabbros and sheeted dike complex; (2) youth, during which is continued melting of refractory asthenosphere (depleted during birth) occurs in response to fluid flux from the subducting slab, with extensional deformation of the older plutonic suite, eruption of refractory lavas, and the intrusion of wehrlite-pyroxenite; (3) maturity, with the onset of semistable arc volcanism, typically calc-alkaline, as the subduction zone matures and stabilizes, and the intrusion of quartz diorite and eruption of silicic lavas; and (4) death, which is the sudden demise of active spreading and ophiolite-related volcanism, which in many cases is linked to collision with an active spreading center and the onset of shallow underthrusting of the buoyant spreading axis; expressed as dikes and lavas with oceanic basalt compositions that crosscut or overlie rocks of the older suites; (5) resurrection, with emplacement by obduction onto a passive margin or accretionary uplift with continued subduction. The early stages (1–3) may be diachronous, and each stage may overlap in both time and space. The existence of this consistent progression implies that ophiolite formation is not a stochastic event but is a natural consequence of the SSZ tectonic setting.

Keywords: Suprasubduction zone ophiolites; oceanic lithospheric; Oman; Coast Range ophiolite; island arcs.

Index terms: Geochemical cycles; plate tectonics; mineralogy and petrology; dynamics of lithosphere and mantle general.

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1. Introduction

[2] Detailed studies of well-exposed ophiolites over the last 20 years have shown that many formed in suprasubduction zone (SSZ) settings; that is, in the upper plate at a convergent plate boundary, and not at mid-ocean ridge spreading centers [e.g., *Evarts*, 1977; *Miyashiro*, 1973; *Pearce et al.*, 1981]. The suprasubduction zone setting encompasses the early evolution of nascent or reorganized subduction zones, prior to the onset or renewal of emergent arc volcanism and plutonism, and includes processes that lead to arc rifting and the formation of forearc and intra-arc basins [e.g., *Hawkins et al.*, 1984; *Stern and Bloomer*, 1992]. These ophiolites have many features in common that indicate a consistent sequence of events during their formation and evolution. In this contribution I review the sequence of plutonic and volcanic units that are common to most suprasubduction zone ophiolites, their petrologic and geochemical characteristics, and their inferred origins within the suprasubduction zone environment. I will also examine events that may lead to the death of these ophiolites, and their subsequent resurrection as they are emplaced at structural levels in the crust that expose them for our study.

[3] Suprasubduction zone ophiolites sensu lato include both those formed in back arc basins (e.g., the Josephine ophiolite of northern California and southern Oregon) and those that contain rocks typically associated with forearc extension (Troodos, Oman, and the Coast Range ophiolite of California). Back arc basins are characterized by lithologic associations and geochemical systematics grossly similar to mid-ocean ridge basalts (MORB); ophiolites formed in this setting are for the most part indistinguishable from MORB geochemically and can only be associated with a back arc origin by careful study of the regional geologic setting [e.g., *Harper*, 1984; *Harper et al.*, 1985]. I will focus here on those ophiolites thought to be associated with forearc rifting. These constitute most of the major ophiolite occurrences of the world, including those that are commonly used as structural analogues for oceanic crust.

[4] One of the most robust tectonic models for SSZ ophiolite formation is that of *Stern and Bloomer* [1992], which builds on earlier work by *Hawkins et al.* [1984], *Casey and Dewey* [1984], *Leitch* [1984], and others. This model proposes that ophiolites generally form during subduction zone initiation, when old, relatively dense, oceanic lithosphere begins to sink into the asthenosphere (Figure 1). Lithosphere in the upper plate adjacent to the sinking lithosphere must extend rapidly into the gap left as the dense lithosphere sinks. Crustal formation is fed by melts from hot asthenosphere that must flow upward into the region above the sinking plate margin, even as the sinking plate displaces the asthenosphere below itself [*Stern and Bloomer*, 1992]. Melting of the hot asthenosphere that flows into the gap created by the sinking plate margin is enhanced by a massive fluid flux from the sinking lithosphere. This combination of rapid decompression melting with fluid enhanced lowering of the solidus leads to extensive melting of the shallow asthenospheric wedge, creating refractory lavas such as boninites and high-Mg andesites and leaving an even more refractory residue of harzburgite tectonite [*Stern and Bloomer*, 1992]. A similar progression could also form during major reorganizations of plate boundaries.

[5] This model has implications for the petrologic development of ophiolites that I will explore in some detail below. In particular, I will show that the sequence of magmatic events in all SSZ ophiolites is consistent with the progressive stages of this model, and this sequence may lead to the emplacement of an

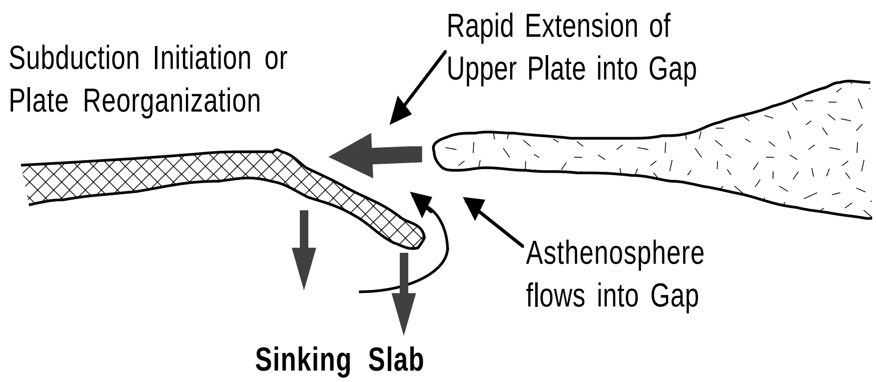


Figure 1. Schematic model for ophiolite formation by rapid extension in the upper plate of a nascent subduction zone, in response to sinking of the lower plate lithosphere [after *Stern and Bloomer*, 1992]. MORB-source asthenosphere flows into the wedge beneath the extending lithosphere and is fluxed with fluids from the sinking slab. Melting occurs in response to decompression of the lithosphere and the aqueous flux from the slab.

ophiolite or to its burial as the basement of a superjacent island arc complex.

2. Petrologic and Geochemical Signatures of Suprasubduction Ophiolites

[6] The petrologic and geochemical signatures of suprasubduction zone ophiolites have been reviewed many investigators, including *Miyashiro* [1973], *Wood* [1980], *Pearce* [1982], *Shervais* [1982], *Pearce et al.* [1984], *Rautenschlein et al.* [1985], *Harris et al.* [1986], and *Pearce and Parkinson* [1993]. Some general features of subduction zone magmatism found in suprasubduction zone ophiolites and other arc rocks include the following:

1. Enrichment in large ion lithophile elements (LILE: K, Rb, Cs, Th, and the light rare earth elements, LREE) relative to normal MORB (NMORB) in response to aqueous fluids or melts expelled from the subducting slab [e.g., *Pearce*, 1982; *Wood*, 1980]. These elements correspond in general to the low field strength elements of *Saunders et al.* [1980]. Some LILE (e.g., K, Rb, Ba) tend to be soluble in aqueous solutions or melts during slab dewatering reactions; others (Th, LREE) are relatively immobile during alteration.
2. Depletion in the high field strength elements (HFSE: Ti, Nb, Ta, Hf) relative to NMORB, which may be caused by larger fractions of partial melting, also in response to aqueous fluids or melts expelled from the subducting slab [e.g., *Pearce*, 1982; *Pearce and Norry*, 1979; *Shervais*, 1982; *Wood*, 1980].
3. The common occurrence of refractory, second stage melts with high MgO (more olivine in source), high silica (more enstatite in source), and high LILE (slab component added with aqueous fluids or melts expelled from the subducting slab) [*Crawford et al.*, 1989; *Ernewein et al.*, 1988; *Juteau et al.*, 1988a, 1988b; *Malpas*, 1990; *Robinson and Malpas*, 1990].
4. Higher oxygen fugacities than NMORB, as reflected by low Ti/V ratios in arc volcanics and in suprasubduction zone ophiolites, and by the occurrence of calc-alkaline fractionation trends in some arc volcanic suites; V partitioning is controlled by fO_2 during melting, whereas Ti partitioning is

independent of oxidation state [*Shervais*, 1982].

5. The typical occurrence of clinopyroxene before plagioclase during crystallization, resulting in the crystallization sequence olivine-clinopyroxene-plagioclase instead of the typical MORB crystallization sequence olivine-plagioclase-clinopyroxene [*Cameron et al.*, 1980; *Hebert and Laurent*, 1990].
6. The common occurrence of highly calcic plagioclase and/or orthopyroxene in some cumulate plutonic rocks caused by changes on phase equilibria during hydrous melting and crystallization [e.g., *Beard*, 1986; *Hebert and Laurent*, 1990].
7. The association with refractory lithosphere comprising harzburgite tectonites and dunite in contrast to the more fertile abyssal lherzolites commonly found with oceanic crust [e.g., *Dick*, 1989; *Dick and Bullen*, 1984].
8. Mineral compositions in the lavas and mantle tectonites that are refractory compared to those found in oceanic basalts and abyssal peridotites, e.g., Cr spinels with high Cr/Al ratios and low Mg/Fe ratios [*Cameron*, 1985; *Crawford et al.*, 1989; *Dick*, 1989; *Dick and Bullen*, 1984; *Umino et al.*, 1990].
9. Enrichment in the radiogenic isotopes of Sr and Pb, resulting in higher $^{87}Sr/^{86}Sr$, $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ ratios, and depletion in radiogenic Nd, resulting in lower $^{143}Nd/^{144}Nd$ ratios, in the volcanic rocks relative to MORB [e.g., *Cohen and O'Nions*, 1982; *Hamelin et al.*, 1984, 1981; *Hanan and Schilling*, 1989; *Ito et al.*, 1987; *Jacobsen and Wasserburg*, 1979; *McCulloch and Cameron*, 1983; *McCulloch et al.*, 1980, 1981; *Noiret et al.*, 1981; *Sun*, 1980].

[7] These geochemical signatures result from processes or conditions that are unique to subduction zones or that are enhanced within the subduction zone environment. As a result, ophiolites that are characterized by the geochemical and isotopic signatures listed above are likely candidates for formation within a suprasubduction zone setting.

2.1. MORB With Arc Signatures

[8] There are a few circumstances in which oceanic igneous rocks with these signatures are erupted in mid-ocean spreading ridge settings. One is when a subduction zone "flips" polarity, leaving relict back arc crust and its subduction-enriched lithosphere and asthenosphere in front of the new subduction zone. If new crust is formed at a spreading center in this location (which is likely if back arc spreading was active prior to the polarity flip), this crust will carry the memory of its former position in the upper plate of a subduction zone. The classic example of this process is the Woodlark Basin in the SW Pacific, which was a back arc basin to the Solomon Islands arc prior to its collision with the Ontong Java plateau [*Perfit et al.*, 1987; *Staudigel et al.*, 1987; *Taylor and Exon*, 1987].

[9] The other special circumstance is found where subduction of an active spreading center occurs at high angles to the subduction zone. In this case, a "slab window" opens inside the subduction zone in response to extinction of the spreading center when it enters the trench. Active spreading in front of the subduction zone continues to form new oceanic crust, leaving a wedge-shaped opening or "window" in the subducting slab [*Karsten et al.*, 1996; *Klein and Karsten*, 1995; *Lytwyn et al.*, 1997; *Sherman et al.*, 1997; *Sturm et al.*, 1999]. This window allows subduction-enriched asthenosphere to rise through the slab into the region of active spreading, in the same manner that plume-enriched mantle flows outward along a sublithospheric conduit beneath active spreading centers [e.g., *Hanan and Schilling*, 1989].

The best example of this process is where the South Chile Rise is being subducted beneath South America slab [*Karsten et al.*, 1996; *Klein and Karsten*, 1995; *Sherman et al.*, 1997; *Sturm et al.*, 1999].

[10] These observations show that it is possible to create oceanic crust at mid-ocean ridge spreading centers that resemble arc tholeiites and may explain the origin of some SSZ ophiolites that are structurally associated with the lower plate in a subduction zone. However, these settings do not explain the occurrence of later magma suites in SSZ ophiolites, nor do they solve the problem of how large intact sheets of ocean crust are transferred to the upper plate and obducted onto passive continental margins.

[11] Nonetheless, these examples show that it is not possible to establish the origin of any particular ophiolite on the basis of geochemical discriminants alone. Each ophiolite must be evaluated based on its structural setting, its associated sedimentation, and its relationship to adjacent terranes, in addition to any geochemical fingerprinting.

3. Life Cycle of SSZ Ophiolites

[12] SSZ ophiolites display a consistent sequence of events during their formation and evolution that suggests that they form in response to processes that are common to all such ophiolites and that are characteristic of their mode of formation. This series of events may be summarized briefly as follows (Table 1):

1. Birth, which is the formation of the ophiolite in a nascent or reconfigured subduction zone and the initiation of subduction-related volcanism and plutonism (Figure 2a);
2. Youth, which is the continued melting of refractory asthenosphere (depleted during birth) in response to fluid flux from the subducting slab (Figure 2b);
3. Maturity, which is the onset of semistable arc volcanism, often calc-alkaline in character, as the subduction zone matures and stabilizes (Figure 2c);
4. Death, which is the sudden demise of active spreading and ophiolite-related volcanism, which in many (but not all) cases is linked to collision with an active spreading center, and its partial or complete subduction by the over-riding plate (Figure 2d);
5. Resurrection, which is the emplacement of the ophiolite onto a passive continental margin (“obduction”) in the case of “Tethyan” ophiolites, or “emplacement” through the process I will call “accretionary uplift” in the case of many “Cordilleran” ophiolites (Figure 2e).

[13] These events generally progress in an orderly fashion from birth through death and resurrection, but not all ophiolites display all stages of this proposed life cycle (Tables 2 and 3). In particular, some SSZ ophiolites never reach maturity but skip directly to death and resurrection. In others, death and resurrection are coincident, with no evidence for a prolonged interval between these events. The “birth” and “youth” stages seem to be common to all SSZ ophiolites and are in fact the most characteristic phases of ophiolite growth in a SSZ setting. The structural aspects of resurrection are also characteristic of all SSZ ophiolites and reflect their original formation in the upper plate of a subduction zone. In all cases, the relative progression of events is consistent from ophiolite to ophiolite, so that events occur in the same relative order even if evidence for some stages of development is missing.

3.1. Stage 1: Birth

[14] It is now generally agreed that many SSZ ophiolites represent crustal spreading above

Table 1. Life Cycle of Ophiolites, Correlating Volcanic, Plutonic, and Metamorphic Rock Series With the Five Stages of Ophiolite Evolution

	Stage 1: Birth	Stage 2: Youth	Stage 3: Maturity	Stage 4: Death	Stage 5: Resurrection
Events	initial spreading, hinge rollback?	refractory melts, second stage melting (LILE enriched)	calc-alkaline, "normal arc"	ridge subduction or directly to obduction	obduction onto passive margin (Tethyan) or accretionary uplift (Cordilleran)
Volcanic rocks	primitive arc tholeiites (basalt to basaltic-andesite)	high-Mg andesites, boninites, tholeiitic ankaramites	andesite, dacite, basaltic andesite	MORB-like or OIB	none
Plutonic rocks	layered gabbro, troctolite, dunite	wehrlite-Cpxite sill complex,	quartz diorite hornblende diorite, agmatites	none	rare granitoids; anatexis of lower plate
Metamorphic rocks	hydrothermal alteration of volcanics	hydrothermal alteration of volcanics	hydrothermal alteration of volcanics	high-grade soles (amphibolites, granulites with MORB affinities)	obduction may be cold or hot, depending on time gap; new subduction zone may form (accretionary uplift)

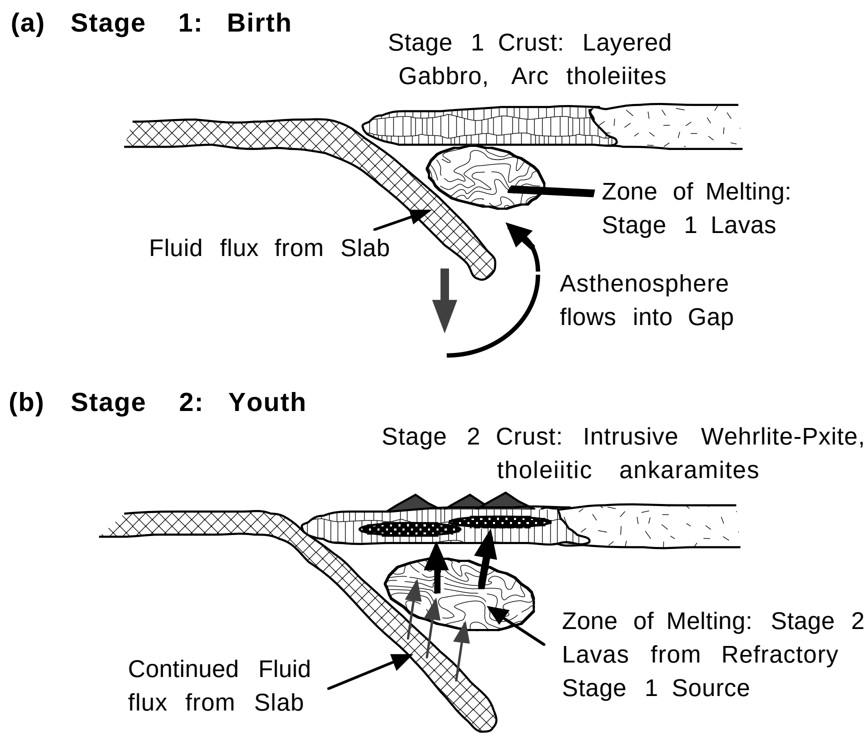


Figure 2. Idealized life cycle of a suprasubduction zone ophiolite, based on the example of the Coast Range ophiolite in California. (a) Stage 1: Birth. Flow of MORB-source asthenosphere into the wedge beneath the extending lithosphere, which is fluxed with fluids from the sinking slab. Melting occurs in response to decompression of the lithosphere and the aqueous flux from the slab, as in Figure 1; (b) Stage 2: Youth, which is the flow of new MORB-source asthenosphere into the wedge beneath the extending lithosphere slows as slab stabilizes, sinks more slowly; massive fluid flux from slab lowers solidus of previously melted mantle wedge (stage 1) to form refractory, second stage melts; (c) Stage 3: Maturity, which is the onset of semistable subduction regime, with balance between influx of virgin asthenosphere and fluids from slab, transition to calc-alkaline magma series; note approach of active spreading center with variably-enriched asthenosphere, seamounts; (d) Stage 4: Death, which is the collision of ridgecrest/spreading center with subduction zone; partial subduction of former spreading center site. Leakthrough of MORB, OIB melts, formation of very high *T* metamorphic sole, and onset of shallow thrusting of ophiolite over oceanic lithosphere; end of active, arc-related magmatism in ophiolite; (e) Stage 5: Resurrection, which may occur by accretionary uplift (illustrated here for the CRO of California) or by obduction onto a passive continental margin.

nascent or reconfigured subduction zones [e.g., *Casey and Dewey*, 1984; *Hawkins et al.*, 1984; *Karig*, 1982; *Leitch*, 1984; *Stern and Bloomer*, 1992]. The initiation of subduction typically occurs in response to major plate reorganizations and changes in either or both convergence directions and rates [*Casey and Dewey*, 1984; *Hawkins et al.*, 1984]. The most widely accepted model at this time is the hinge roll-back model [*Stern and Bloomer*, 1992], in which old lithosphere sinks into the asthenosphere, leaving a “gap” that is filled by rapid spreading at the leading edge of the lithosphere in the upper plate (Figure 1). The spreading rate in this gap is determined by the rate at which the older lithosphere sinks, the gap itself being filled continuously by the newly formed crust and lithosphere. In general, the spreading rate

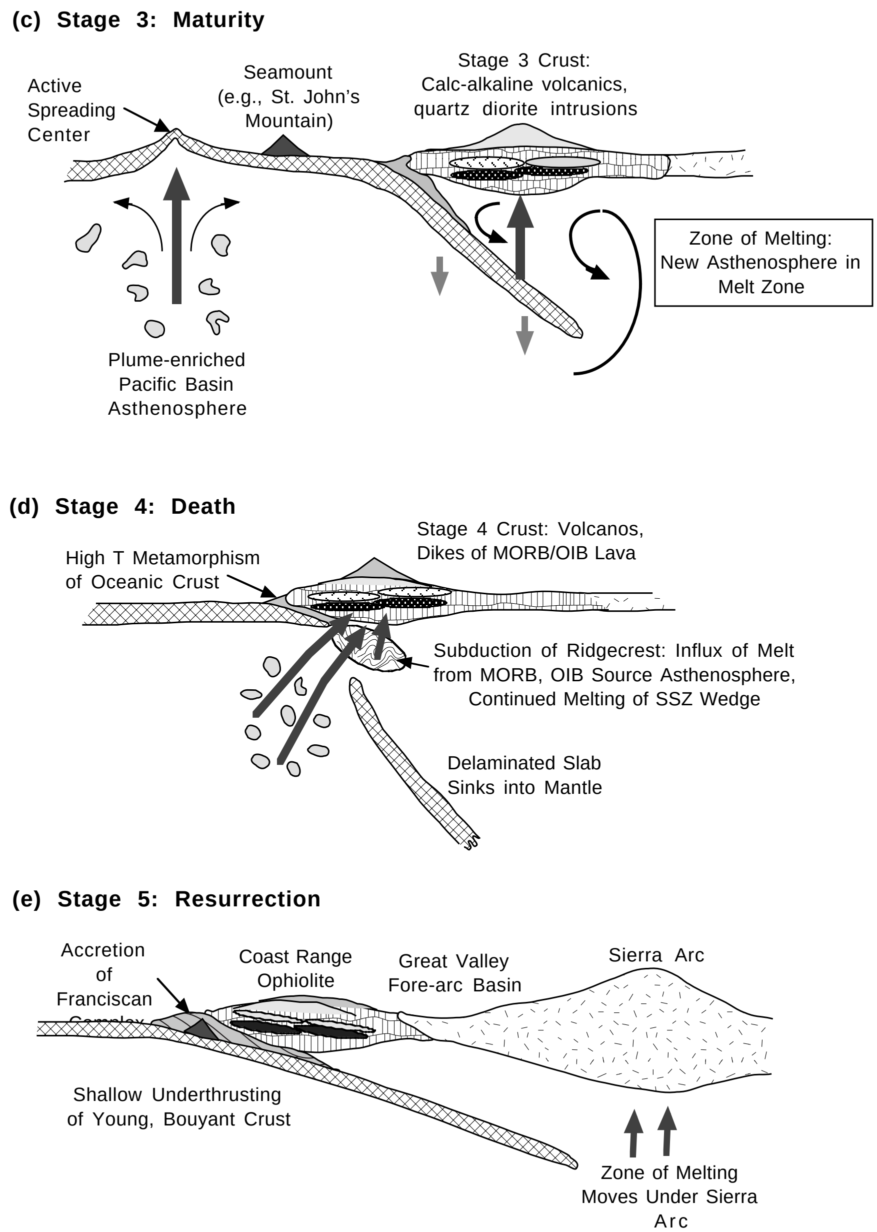


Figure 2 (continued)

must be quite rapid to keep pace with the sinking lithosphere. This model explains both the occurrence of spreading (often rapid) within what is normally considered a compressional environment, and the occurrence of ophiolite lithologies in the forearc region of active volcanic arcs [e.g., *Stern and Bloomer*, 1992].

[15] Rocks associated with the initial phase of ophiolite formation include the layered gabbros and some isotropic gabbros of the plutonic section, the "lower" volcanic units, and a large portion (but not all) of the sheeted dike complex (Table 1; Figure 2a). Volcanic rocks associated with this phase of ophiolite formation are

typically low-K arc tholeiites ranging in composition from basalt to basaltic andesite or even dacite (Table 2). Because their parent magmas form by melting of MORB-source asthenosphere that flows laterally into the gap created during sinking of the downgoing lithosphere, before this asthenosphere can be significantly enriched by fluid flux from the subducting plate, these low-K tholeiites strongly resemble MORB in their overall geochemical character. They are typically LREE depleted and do not show the strong enrichments in LILE that characterize later melts. Nonetheless, these rocks generally display distinct geochemical signatures that distinguish them from true MORB, including low Ti/V ratios, minor enrichment in the LILE (especially Th), and depletions in the high field strength elements (Ti, Nb, Ta, Hf) relative to the LILE. These melts cannot form by the same pressure release melting mechanism as true MORB because they form from asthenosphere that is already at relatively shallow levels in the mantle. This shallow asthenosphere requires at least some flux of aqueous solutions or vapor-rich melts from the sinking slab in order to melt.

[16] Layered and isotropic gabbros that form during the initial birth phase of ophiolite generation have the same parent magmas as the low-K tholeiite volcanic rocks that form the lower or high-Ti volcanic series in these ophiolites and much of the dike complex. The layered gabbros are often ductiley deformed (foliated or boudinaged), possibly in response to syn-magmatic extension of the nascent crust. This has been documented in Troodos [*Malpas*, 1990], Oman [*Juteau et al.*, 1988a; *Juteau et al.*, 1988b; *Nicolas*, 1992], and California [*Shervais and Beaman*, 1991]. This deformation is rarely seen in later magmatic series so it must relate to the extremely rapid early extension of the SSZ crust. This implies that true "sheeted dike complex" formation must be associated with this same early stage of ophiolite formation. It should be noted, however, that most sheeted dike complexes may contain up to five generations of dikes with different compositions and that cross-cutting relations show that different magma series overlap in time (e.g., Troodos [*Baragar et al.*, 1990; *Staudigel et al.*, 1999]).

3.2. Stage 2: Youth

[17] Youth, the second stage of SSZ ophiolite formation, is characterized by second stage melts of the mantle wedge overlying the new subduction zone (Table 1; Figure 2b). These melts form in response to continued melting of previously depleted asthenosphere brought about by the increasing flux of fluids and melts from the subducting slab. A decrease in the rate at which the subducting lithosphere sinks reduces the influx of new asthenosphere and strands the previously depleted asthenosphere within the "melting zone" beneath the ophiolite.

[18] Rocks associated with this second phase of ophiolite formation include mafic and ultramafic rocks which are intrusive into the older plutonic section, "upper" or "low-Ti" volcanic units that overlie stage 1 volcanics, and parts of the dike complex (Table 1). Volcanic rocks associated with this phase of ophiolite formation vary from primitive Ol + augite phyric tholeiites to tholeiitic ankaramites, high-Ca boninites, high-Mg andesites, and clinoenstatite-phyric, low-Ca boninites (Table 2) [*Alabaster et al.*, 1982; *Cameron et al.*, 1980; *Hopson and Frano*, 1977; *Malpas and Langdon*, 1984; *Umino et al.*, 1990]. Because their parent magmas form by melting of previously depleted asthenosphere that was stranded within the zone of melting in the shallow mantle wedge above the subduction zone, these lavas are strongly depleted in the high field strength trace elements (Ti, Nb, Ta, Hf) and in the HREE compared to oceanic basalts. The

Table 2. Comparison of Major Ophiolite Occurrences and Their Correspondence to the Proposed Life Cycle for SSZ Ophiolites[a]

	Stage 1: Birth	Stage 2: Youth	Stage 3: Maturity	Stage 4: Death	Stage 5: Resurrection
Events Ophiolite	initial spreading hinge rollback	refractory melts, second stage melting	stable arc (calc-alkaline or arc tholeiite)	ridge subduction or obduction	obduction (Tethyan) or accretionary uplift
Troodos	volcanics: "lower pillows", high-Ti, "group A", arc tholeiites	volcanics: "upper pillows", low-Ti, groups B, C (high MgO andesite, boninites	volcanics: not found	not observed direct to obduction	obduction onto passive margin
	plutonics: layered, foliated gabbro, dunite, isotropic gabbro	plutonics: wehrlite, lherzolite, olivine gabbro, websterite	plutonics: not found		
Oman (Semail)	volcanics: Geotimes, Lasail arc tholeiites	volcanics: Opx-phyric basalts of Alley volcanics (boninitic)	volcanics: basalt, andesite, dacite, and rhyolite of Alley volcanics	ridge subduction (≈95 Ma) Sahali volcanics (alkalic, oceanic) 85–90 Ma	obduction onto passive margin circa 75 Ma (Juaizi Formation, late Campanian molasse)
	plutonics: layered gabbros	plutonics: wehrlite-pyroxenite	plutonics: felsite dikes, plagiog ranites (U/Pb ≈ 96 Ma)	metamophic sole (MORB amphibolite) (Ar/Ar ≈ 95 Ma)	postemplacement cover ≈ 70 Ma (late Maastrichtian)
Vourinos	volcanics: Krapa series (Cpx + Plg phyric)	volcanics: Asprokambo series (Cpx + Opx ± Plg phyric)	volcanics: Rhyolite, dacite		obduction onto passive margin
(Greece)	plutonics: layered gabbro pxite- wehrlite	plutonics: websterite, gabbronorite	Plutonics: Hb gabbro, diorite, granophyre, agmatites	metamorphic sole (amphibolites)	
Elder Creek Coast Range Ophiolite	volcanics: pillows, dike complex		volcanics: clasts in Crowfoot Point Breccia (andesites)	ridge subduction (MORB dikes, Stonyfordvolcanics) (Ar/Ar ≈ 163 Ma)	accretionary uplift by Franciscan Complex
(California)	plutonics: layered, foliated gabbro, dunite	plutonics: wehrlite-clinopyroxenite-gabbro	plutonics: Hb diorite, qtz diorite plutons (U/Pb ≈ 166–172)	Franciscan high grade blocks (U/Pb, Ar/Ar ≈ 162–159)	
Del Puerto Canyon Coast Range Ophiolite	Volcanics: Pillow lava, massive flows, sills (Cpx-Plg phyric)	volcanics: Pillow lava, massive flows (Ol-Cpx phyric), boninites	volcanics: Lotta Creek tuff, Leona rhyolite; Qtz kerato-phyre lavas, sills	ridge subduction (late MORB dikes)	accretionary uplift by Franciscan Complex

Table 2. (continued)

	Stage 1: Birth	Stage 2: Youth	Stage 3: Maturity	Stage 4: Death	Stage 5: Resurrection
(California)	Plutonics: dunite, layered gabbro	plutonics: wehrlite, clinopyroxenite, dunite	plutonics: Hb Qtz diorite plutons, 200 m + thick	Franciscan high grade blocks (U/Pb, Ar/Ar ≈ 162–159)	
Llanada Coast Range Ophiolite	Volcanics: Pillow lava (Cpx-Plg phyric)	volcanics: pillow lava (Ol-Cpx phyric)	volcanics: Lotta Creek tuff; Qtz keratophyre lavas, sills; andesite breccia, lahars	not reported	accretionary uplift by Franciscan Complex
(California)	plutonics: layered gabbro	plutonics: feldpathic wehrlite, dunite	keratophyre sills		
Point Sal Coast Range Ophiolite	volcanics: lower pillow lavas (Plg-Cpx phyric)	volcanics: upper pillow lavas (olivine-phyric)	volcanics: tuffaceous chert	not seen at Point Sal (late Ol-basalts crosscut diorites at Cuesta Ridge)	accretionary uplift by Franciscan Complex
(California)	plutonics: layered, foliated gabbro, dunite	plutonics: wehrlite-clinopyroxenite	keratophyre sill complex, plagiogranite		
Bay of Islands	volcanics: Intermediate Ti lavas of Betts Head Formation. (Cpx + Plg phyric)	volcanics: boninites in Betts Head Formation (Ol ± Opx ± Cpx phyric)	volcanics: Snooks Arm Formation, arc tholeiites	not reported	obduction onto passive margin
(Newfoundland)	plutonics: Layered, folded olivine gabbro (unit 3), isotropic Hb gabbro (unit 4)	plutonics: sill complex of wehrlite, harzburgite, Cpxite (unit 2)	Qtz gabbro, Hb diorite, trondhjemite stocks		
Muslim Bagh	not preserved	not preserved	not preserved	ridge subduction late dolerite dikes	obduction onto passive margin (=Maastrichtian limestone)
(NW Pakistan)	plutonics: layered gabbro, dunite	plutonics: pyroxenite, wehrlite	Qtz diorite, Hb diorite, trondhjemite (dike complex)	metamophic sole (garnet amphibolite, greenschist, phyllite, marble)	postemplacement cover ≈ 55 Ma (early Eocene)
Ballantrae Complex	island arc tholeiites, 501 ± 12 Ma	boninites, tholeiites 476 ± 14 Ma	calc-alkaline volcanic sandstones	ocean island tholeiite Bennae Head,	obduction during arc-continent collision 478 ± 8 Ma;
(Scotland)	Mains Hill Sequence	Games Loup Sequence	Kilranny Hill Formation	Pinbain Sequences	Caledonian orogeny

[a]Data sources are given in Table 3.

Table 3. Sources of Data for Table 2

Ophiolite	Sources
Troodos (Cyprus)	*Baragar et al.* [1990], *Cameron* [1985], *Hebert and Laurent* [1990], *Kay and Senechal* [1976], *Kidd and Cann* [1974], *Laurent* [1990], *Malpas* [1990], *Malpas and Langdon* [1984], *Malpas et al.* [1990], *McCulloch and Cameron* [1983], *Miyashiro* [1973, 1975a, 1975b], *Moores* [1975], *Rautenschlein et al.* [1985], *Robertson* [1977], *Robinson and Malpas* [1990], *Robinson et al.* [1983], *Schmincke et al.* [1983], *Schmincke and Bednarz* [1990], *Staudigel et al.* [1999], *Taylor* [1990]
Semail (Oman)	*Alabaster et al.* [1982], *Benn et al.* [1988], *Bernoulli et al.* [1990], *Boudier and Coleman* [1981], *Boudier and Nicolas* [1988, 1995], *Briqueu et al.* [1991], *Browning* [1984], *Chen and Pallister* [1981], *Coleman* [1981], *Coleman and Hopson* [1981a, 1981b], *Dunlop and Fouillac* [1985], *El-Shazly et al.* [1990], *El-Shazly* [1994], *El-Shazly and Lanphere* [1992], *El-Shazly and Liou* [1991], *El-Shazly et al.* [1997], *Ernewein et al.* [1988], *Gealey* [1977], *Ghent and Stout* [1981], *Glennie et al.* [1990], *Gnos* [1998], *Gnos and Peters* [1993], *Gregory* [1984], *Gregory et al.* [1998], *Hacker* [1991], *Hacker et al.* [1996], *Hopson et al.* [1981a], *Hopson and Pallister* [1980, 1981], *Juteau et al.* [1988a, 1988b], *Kelemen et al.* [1997], *Korenaga and Kelemen* [1997], *Lanphere* [1981], *Lanphere et al.* [1981a, 1981b], *Lippard* [1983], *Lippard et al.* [1986], *Montigny et al.* [1988], *Nicolas and Boudier* [1995], *Nicolas et al.* [1994, 1988], *Pallister* [1981], *Pallister and Knight* [1981], *Pearce et al.* [1981], *Robertson* [1986], *Searle and Cox* [1999], *Searle* [1985], *Searle and Cooper* [1986], *Searle and Graham* [1982], *Searle et al.* [1980], *Tilton et al.* [1981], *Umino et al.* [1990], *Vetter and Stakes* [1990], *Woodcock and Robertson* [1982], *Yanai et al.* [1990]
Vourinos, Pindos (Greece)	*Beccaluva et al.* [1984], *Dostal et al.* [1991], *Dupuy et al.* [1984], *Harkins et al.* [1980], *Jackson et al.* [1975], *Moores* [1969], *Noiret et al.* [1981], *Ross et al.* [1980], *Ross and Zimmerman* [1996]
Coast Range Ophiolite (California)	*Bailey and Blake* [1974], *Bailey et al.* [1970], *Blake et al.* [1987], *Blake et al.* [1992], *Dickinson et al.* [1996], *Evarts* [1977], *Evarts et al.* [1999], *Evarts et al.* [1992], *Giaramita et al.* [1998], *Hagstrum and Murchey* [1996], *Hopson et al.* [1992], *Hopson et al.* [1997], *Hopson et al.* [1981b], *Hull et al.* [1993], *Jackson and Schiffman* [1990], *Jayko and Blake* [1986], *Jayko et al.* [1987], *Lagabrielle et al.* [1986], *Mattinson and Hopson* [1992], *McLaughlin et al.* [1988], *McLaughlin et al.* [1992], *Page* [1972], *Phipps and Macpherson* [1992], *Robertson* [1989], *Robertson* [1990], *Schiffman et al.* [1991], *Shervais* [1990], *Shervais and Kimbrough* [1985a, 1985b], *Taylor et al.* [1992], *Williams* [1984], *Yule* [1997]
Newfoundland	*Bedard and Hebert* [1996], *Bedard and Hebert* [1998], *Bedard et al.* [1998], *Cawood and Suhr* [1992], *Church and Stevens* [1971], *Dewey and Bird* [1971], *Dunning and Krogh* [1985], *Elthon et al.* [1984], *Jacobsen and Wasserburg* [1979], *Jamieson* [1980], *Jenner et al.* [1991], *Karson and Dewey* [1978], *Karson* [1984], *Suen et al.* [1979]
Muslim Bagh (NW Pakistan)	*De Jong* [1982], *Kojima et al.* [1994], *Sarwar* [1992]
Ballantrae Complex (Scotland)	*Smellie and Stone* [1992], *Thirlwall and Bluck* [1984]

large flux of slab-derived fluids required to lower the solidus of this refractory mantle, however, compensates for this depletion in by adding LILE (K, Rb, Th) and LREE to the melts, resulting in low Ti/V ratios, LREE-enriched rare earth patterns, and MORB-normalized incompatible element plots that are enriched in the LILE and have negative anomalies in the high field strength elements [*Cameron et al.*, 1983; *Crawford et al.*, 1989; *Hickey and Frey*, 1982; *McCulloch and Cameron*, 1983; *Shervais*, 1982].

[19] Plutonic rocks associated with this second phase of ophiolite formation include wehrlites, pyroxenites, primitive gabbros, and some diorites (Table 2). These rocks intrude the older plutonic section of layered and foliated gabbro, in places forming thin sills intruded parallel to the older layering and elsewhere forming stocks and dikes that clearly crosscut the older layering and foliation (e.g., Oman [*Ernewein et al.*, 1988; *Juteau et al.*, 1988a; *Juteau et al.*, 1988b]; Troodos [*Laurent*, 1992; *Malpas*, 1990]; Coast Range ophiolite [*Evarts et al.*, 1999; *Hopson and Frano*, 1977; *Shervais and Beaman*, 1991]; Bay of Islands [*Bedard and Hebert*, 1996; *Bedard et al.*, 1998]). Age relations relative to the older layered gabbros are shown by xenoliths of the older rocks in the younger intrusions, by intrusive contacts that crosscut layering and foliation in the older rocks, and by the lack of fabric in the younger rocks.

3.3. Stage 3: Maturity

[20] Maturity comes to SSZ ophiolites (as it comes to us all) when the subduction zone becomes relatively stable and the rate of crustal spreading slows. Melts formed at this stage begin to resemble those found in established intraoceanic island arcs, where a balance is achieved between the flow of relatively undepleted asthenosphere into the zone of melting and the flux of slab-derived fluids and melts. Magmas formed during this third phase of ophiolite generation tend to be more silica saturated, richer in LILE, and depleted in high field strength elements (Table 1; Figure 2c).

[21] Volcanic rocks associated with this phase of SSZ ophiolite evolution include basaltic andesites, andesites, dacites, and rhyolites, which are typically calc-alkaline or transitional toward calc-alkaline in composition (Table 2). These rocks are characterized by a wide range in silica contents (up to 73% SiO_2), evolved Mg/Fe ratios, and higher alkalis than in previous magma suites. The lower silica volcanics may occur as pillows or massive flows, but the more evolved rocks typically occur as volcanic breccias and other volcaniclastic deposits (e.g., Del Puerto Canyon ophiolite [*Evarts et al.*, 1999; *Evarts et al.*, 1992]). In many cases these rocks are not considered part of the subjacent ophiolite and are referred to as the "postophiolite" volcanic series. In most cases, however, these volcanics lie beneath siliceous sediments that are considered to cap the ophiolite in other locations (e.g., Llanada ophiolite [*Giaramita et al.*, 1992; *Hopson et al.*, 1981b]).

[22] Plutonic rocks formed during stage 3 include hornblende diorites, quartz diorites, tonalites, and trondhjemites. These rocks are generally grouped under the generic term "plagiogranite" [*Coleman and Peterman*, 1975]. Stage 3 plutonic rocks typically intrude at high levels within the plutonic complex, forming tabular, sill-like plutons between the layered gabbros and the sheeted dike complex. These tabular plutons may be up to 1 km thick and over several kilometers in length [e.g., *Lippard et al.*, 1986; *Pike*, 1974]. They also form dikes and sills at both deeper and shallower levels of the complex. Dioritic dikes within the sheeted complex probably fed the volcanic sequences. Geochemically, the stage 3 plutonic rocks represent a calc-alkaline intrusive suite that corresponds to the stage 3 volcanic rocks [*Briqueu et al.*, 1991; *Floyd et al.*, 1998; *Lachize et al.*, 1996; *Shervais and Beaman*, 1991].

[23] A common feature of these late dioritic intrusives is their propensity to form agmatites (igneous intrusion breccias formed by the stopping and partial dissolution of the wallrocks). Typical stopped blocks include layered gabbro, isotropic gabbro, dike complex, and volcanic

rocks. In places the agmatites are so choked with xenoliths that they exceed the dioritic matrix in volume. Intrusion is interpreted to have occurred at relatively shallow depths, as shown by the abundance of angular xenoliths within the breccias and by miarolytic cavities partially filled with coarse aggregates of hornblende and feldspar.

[24] Only a few ophiolites live to maturity. That is, only a few of the ophiolites that are emplaced where we can study them exhibit this stage of development. It is likely that many ophiolites form and go through all three of these stages, only to end up as the basement to an active, long-lived island arc like the Marianas or the Izu-Bonin arc. In order to preserve this stage of maturity the ocean basin being subducted must be large enough to complete the first two stages without disappearing, but not so large that the ophiolite evolves into the basement of a long-lived arc complex.

3.4. Stage 4: Death

[25] Many ophiolites tend to have short active lives followed by a sudden termination in activity (e.g., Bay of Islands). Others have remained active for longer times, but the nature of that activity changes drastically (e.g., Oman). Whether any given ophiolite has a long life expectancy or not depends largely on its mode of formation; this determines whether death and resurrection are coincident or are separated by a distinct period of time (Table 1; Figure 2d).

[26] Pericollisional ophiolites form when a subduction zone encounters promontories on the passive continental margin being subducted [*Harris*, 1992]. Hinge rollback occurs as the subduction zone spreads rapidly into the adjacent reentrants (e.g., Banda forearc [*Harris*, 1992]; Bay of Islands [*Bedard and Hebert*, 1996; *Bedard et al.*, 1998]). There is no significant time gap between formation of the ophiolite and its obduction onto the passive continental margin. In this case, death and resurrection are essentially simultaneous: ophiolite formation ceases as it is being "resurrected" by obduction onto the continental margin (Table 2).

[27] Other ophiolites (e.g., Oman, Turkey, Troodos) form within fairly large ocean basins. As a result, death may occur by simple obduction onto a passive continental margin after extensive subduction of oceanic crust (if no active ridge crest intervenes), or death may occur when an intervening active spreading center is encountered by the subduction zone (Figure 2d). In this case, thrusting of the ophiolite over the formerly active ridge crest will cause a change in relative plate motions, convergence directions, and convergence rate. It will also introduce under the ophiolite a partially molten upwelling of MORB source asthenosphere that has seen no influence of fluids from the subducting slab (Figure 2d). This can result in the eruption of MORB-like or even ocean island basalt (OIB) lavas on top of the earlier "arc-related" lavas of stages 1 to 3 or the intrusion of these magmas into the plutonic section to form dikes (Table 2). Examples of this include the Coast Range ophiolite of California (Elder Creek, Stonyford, Del Puerto, and other massifs adjacent to the Great Valley [*Shervais and Beaman*, 1991], Oman (Salahi volcanics) [*Alabaster et al.*, 1982; *Ernewein et al.*, 1988; *Umino*, 1995; *Umino et al.*, 1990], and the Kizildag ophiolite of Turkey [*Dilek and Thy*, 1998; *Dilek et al.*, 1999; *Juteau*, 1980; *Juteau et al.*, 1977; *Lytwyn and Casey*, 1993].

[28] Because oceanic lithosphere adjacent to the spreading center is hot and buoyant and overlies even hotter and more buoyant asthenosphere, there are two additional characteristics of ridge collision/subduction: the formation of

high-temperature metamorphic soles (amphibolite, garnet amphibolite, or granulite) with inverted metamorphic aureoles and the onset of shallow thrusting of the ophiolite over the subjacent oceanic lithosphere (Figure 2d).

[29] Formation of the high-grade metamorphic sole requires temperatures (±800°C) that are too hot to achieve by overthrusting within the subduction zone; refrigeration of the basal ophiolite lithosphere by the cold subducting crust quickly cools the upper plate to temperatures well below this [*Hacker and Gnos*, 1997; *Searle and Cox*, 1999; *Wakabayashi*, 1996]. In contrast, subduction of an active spreading center places extremely hot crustal material at the base of the ophiolite [*Cloos*, 1993]. This hot oceanic crust may be sheared off to form a schüppenzone, which is cooled from below by juxtapositioning with cooler crust farther from the spreading center. This model is supported by the composition of amphibolites within the metamorphic soles, which are always MORB-like in major and trace element chemistry and geochemically distinct from magma suites of the overlying ophiolite [e.g., *Alabaster et al.*, 1982; *Lippard et al.*, 1986; *Moore*, 1984]. The model is also supported by the short time spans that typically elapse between formation of the ophiolite and formation of the high-grade metamorphic sole [e.g., *Hacker and Gnos*, 1997; *Hacker et al.*, 1996; *Searle and Malpas*, 1980].

[30] Thrusting of the ophiolite over an active ridge crest as the ridge crest enters the subduction zone terminates formation of arc-like ophiolite magmas because the angle of subduction becomes too low to generate partial melts in the subophiolitic asthenosphere. The angle of subduction is constrained to be low by the buoyancy of the lithosphere and asthenosphere near the ridge crest, both of which are hot and partially molten [*Cloos*, 1993]. Cloos calculates that oceanic crust younger than ≈10 million years is too buoyant to subduct and must thrust under the subduction zone at a shallow angle. The extent of this buoyant lithosphere will be a function of the spreading rate at the ridge crest; for a spreading half rate of 50 mm/yr, the width of buoyant crust could range up to 1000 km. This shallow thrusting will persist until enough lithosphere has been subducted to place older, less buoyant oceanic lithosphere in the subduction zone. If relative convergence persists long enough after subduction of the ridge crest, this older lithosphere will begin to sink and the ophiolite could be reborn or overprinted by renewed subduction-related magmatism. This will only occur if the ocean basin is sufficiently large so that obduction does not occur first.

[31] Previous authors have proposed that subduction actually initiates at the ridge as an ‘‘intraoceanic thrust’’ [e.g., *Boudier et al.*, 1982; *Coleman*, 1981; *Dilek et al.*, 1999; *Hacker and Gnos*, 1997]. While this model is frequently invoked, it has some distinct dynamical problems related to buoyancy. *Cloos* [1993] has shown that oceanic lithosphere on both sides of the spreading center is positively buoyant and cannot subduct (sink) for ≈10 million years after forming. This makes it unlikely for subduction to initiate at the ridge crest, and if it did, formation of the metamorphic sole would predate formation of the ophiolite.

[32] The most important distinction between the model presented here for ridge collision/subduction and that proposed by previous authors is that in my model ridge collision/subduction ends ophiolite formation and results in a new convergence regime of shallow underthrusting prior to obduction. In the older models, ridge-centered thrusting was called upon to initiate subduction, trapping ‘‘axis series’’ oceanic crust (formed at a true

MOR spreading center) above the new subduction zone, where it is overprinted with a subduction component [*Boudier et al.*, 1982; *Dilek et al.*, 1999; *Hacker and Gnos*, 1997]. A critique of the "ridge-initiated intraoceanic thrusting" model in Oman is presented by *Searle and Cox* [1999].

3.5. Stage 5: Resurrection

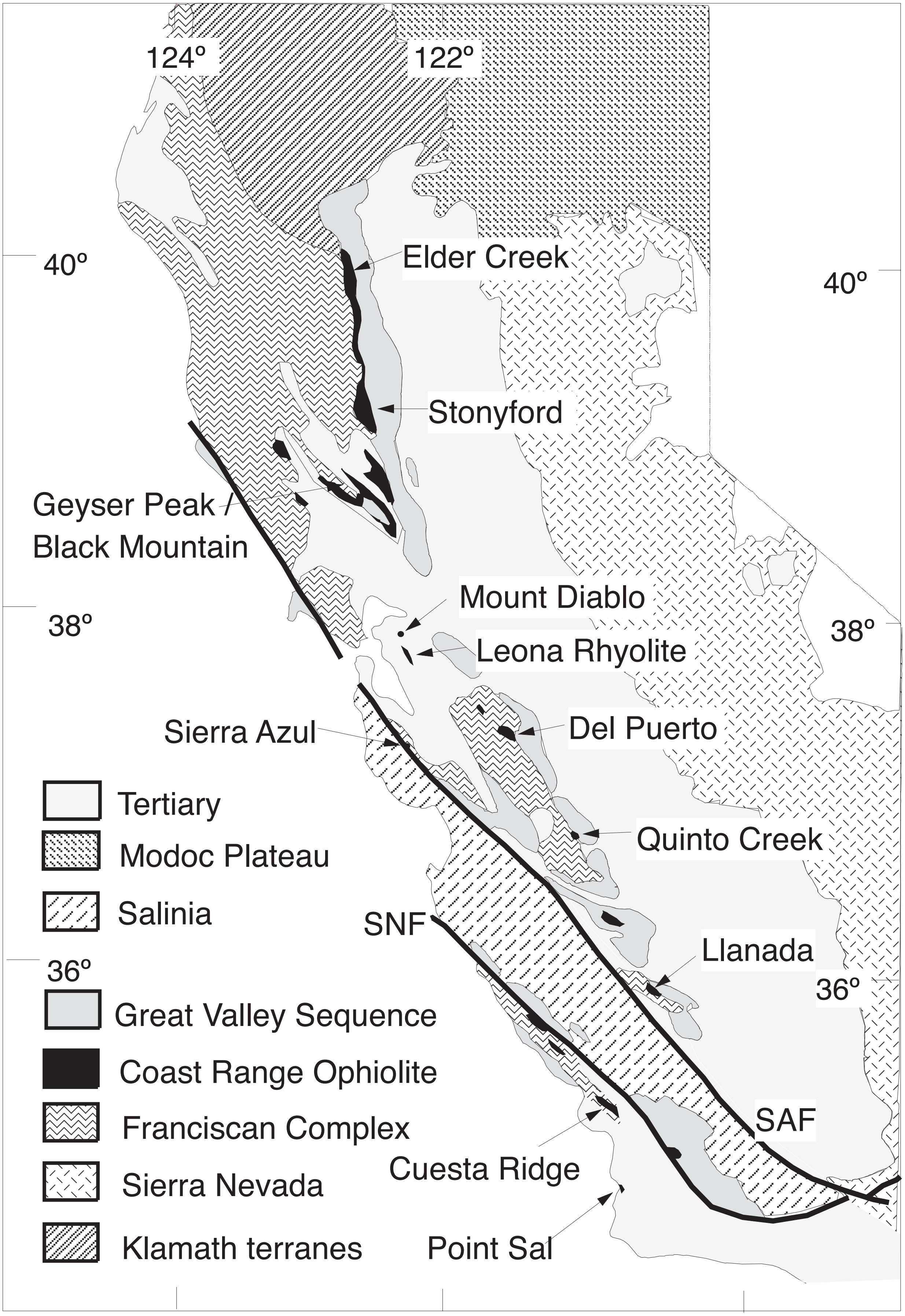
[33] Resurrection occurs when the ophiolite becomes emplaced on a passive continental margin (Tethyan subtype) or is otherwise uplifted and exposed (Table 1; Figure 2e). Resurrection typically occurs in stages over long periods of time, and some Mesozoic ophiolites (e.g., Troodos) are still in the process of being uplifted and exposed.

[34] The most common and easily recognized mode of emplacement is obduction onto a passive continental margin [*Coleman*, 1971]. Maneuvering a true mid-ocean ridge spreading center onto a passive continental margin always presented a problem for early ophiolite obduction models, but that problem has been solved by the recognition that most ophiolites form and evolve in the upper plate of a subduction zone, as outlined above. Obduction now becomes a simple process whereby the ophiolite attempts to subduct the passive margin continental lithosphere and is partially successful until the buoyant continental material rises back to the surface carrying the leading edge of the ophiolite on its back [e.g., *Gealey*, 1977]. Since subduction is largely driven by slab pull, the oceanic lithosphere attached to the continental margin will try to pull the continental lithosphere along with it as it descends into the subduction zone. Detachment of the subducting oceanic lithosphere from the continental passive margin lithosphere (in response to buoyancy of the continental lithosphere) will allow the continental lithosphere to return rapidly to the surface.

[35] This process is inevitable whenever a passive margin continent-ocean boundary is subducted at its other, oceanic-only end. The time elapsed between stage 4 ("death") and stage 5 ("resurrection") depends on the size of the remaining ocean basin. If the ocean basin is small, then the initial stages of resurrection will begin shortly after death (e.g., pericollisional ophiolites such as Bay of Islands). If, on the other hand, a fairly large ocean basin remains, magmatic activity in the ophiolite could stop long before obduction began. This seems to be the case in the Semail ophiolite, where formation of the metamorphic sole occurred around 94 Ma (shortly after formation of the stage 3 quartz diorites), but obduction onto the passive margin of Arabia did not occur until the mid-Campanian ($\approx$75 Ma), as attested by deposition of the Juweiza Formation, a syn-emplacement flysch deposit [*Glennie et al.*, 1974; *Searle and Cox*, 1999; *Yanai et al.*, 1990]. This implies $\sim$20 million years of convergence before emplacement of the Semail nappe.

4. Discussion

[36] The previous sections present a comprehensive new model for the formation of SSZ ophiolites and their evolution within the suprasubduction zone setting. It assumes that magmas that exhibit geochemical systematics that are characteristic of suprasubduction zone magmas did in fact form above a subduction zone and that the consistent progression of magma series within all SSZ ophiolites worldwide results from the operation of a physical process that progresses in a consistent and repeatable fashion. It should be noted, however, that these stages may overlap in time and may even be diachronous; that is, the initiation of each proposed stage may migrate toward the trench in concert with hinge rollback, and several stages may be active simultaneously at different



124º
122º
40º
40º
Elder Creek
Stonyford
Geyser Peak /
Black Mountain
Mount Diablo
38º
38º
Leona Rhyolite
Sierra Azul
Del Puerto
Tertiary
Modoc Plateau
Salinia
Quinto Creek
SNF
Llanada
36º
36º
Great Valley Sequence
Coast Range Ophiolite
Franciscan Complex
Sierra Nevada
Klamath terranes
SAF
Cuesta Ridge
Point Sal

locations within the SSZ environment. In this case, it is expected that stage 1 would be active closest to the trench, while stages 2 and 3 could be active simultaneously at locations progressively farther from the trench.

[37] In this section, I will present two case studies that apply the interpretive fabric of this model to two well-studied ophiolites: the Semail ophiolite of Oman and the Coast Range ophiolite of California. The Semail ophiolite is a typical Tethyan ophiolite that was obducted onto a passive continental margin in the late Cretaceous; the CRO of California is a typical Cordilleran ophiolite that was never obducted but was uplifted and exposed by the process of accretionary uplift. Both ophiolites exhibit remarkable similarities throughout most of their formation and differ largely in their mode of emplacement.

4.1. Case History 1: Coast Range Ophiolite of California

[38] The Coast Range ophiolite of California is a typical Cordilleran ophiolite of *Moore* [1984]. It differs from Tethyan ophiolites like Troodos and Semail largely in its mode of emplacement, or rather, its lack of emplacement. Unlike typical Tethyan ophiolites, which are obducted onto passive continental margins, the Coast Range ophiolite (CRO) of California has been emplaced by what I define here as accretionary uplift: the underplating of material in the accretionary prism, which gradually lifts the overlying ophiolite until collapse of the accretionary wedge preserves the ophiolite along low-angle normal faults [*Platt*, 1986]. The life cycle of the CRO is illustrated in Figure 2; locations discussed in the text are shown in Figure 3.

[39] New and revised U/Pb zircon dates for evolved rocks in the CRO show that it is somewhat older than recognized previously, with most dates falling in the range ≈172 to 165 Ma, although younger dates (to ≈145 Ma) are still found in the Del Puerto Canyon ophiolite remnant [*Evarts et al.*, 1999; *Mattinson and Hopson*, 1992]. This corresponds to the middle Jurassic (Bajocian-Callovian), shortly after the initiation of spreading in the North Atlantic and during a time of plate reorganization along the Cordilleran margin [*Ward*, 1995].

4.1.1. Stage 1: Birth

[40] Stage 1 in the Coast Range ophiolite of California (Figure 2a) is represented in most locales by layered gabbros and dunites of the plutonic complex, much of the sheeted dike/sill complex, and by plagioclase-clinopyroxene phyric arc tholeiite lavas that form the so-called "lower volcanic series" [*Evarts*, 1977; *Giaramita et al.*, 1998; *Hopson et al.*, 1981b; *Shervais*, 1990; *Shervais and Kimbrough*, 1985a, 1985b]. The layered gabbros often display solid-state deformation fabrics, including foliation, deformed layering, and high-temperature normal faults (Figure 4a). The dunites and layered gabbros are typically crosscut by later stage 2 intrusions or form xenoliths within the later stage 2 intrusions (Figure 4b). Calculated parent magmas of the layered gabbros are LREE-depleted arc tholeiite magmas similar to the plagioclase-clinopyroxene phyric lower volcanic series.

[41] Initiation of stage 1 subduction probably occurred in response to an increase in convergence rate in the middle Jurassic that was associated with opening of the central North Atlantic circa 175 Ma and collapse of a

Figure 3. Location map for the Coast Range ophiolite of California, showing locations discussed in the text. SAF, San Andreas fault; SNF, Sur Nacimiento fault.

Figure 4. Coast Range ophiolite of California: (a) ductilely deformed layered gabbro; note foliation and discontinuity of the thin mafic, felsic layers, Point Sal; (b) stage 1 layered gabbro crosscut by dike (vertical) and sill (lower left of photo) of stage 2 wehrlite-pyroxenite; Point Sal. Photos by author.

fringing arc against the western margin of North America [e.g., *Edelman and Sharp*, 1989; *Girty et al.*, 1995; *Saleeby*, 1990; *Tobisch et al.*, 1989; *Ward*, 1995; *Wright and Fahan*, 1988].

4.1.2. Stage 2: Youth

[42] Stage 2 in the Coast Range ophiolite (Figure 2b) is represented by an ultramafic-mafic intrusive complex and by an upper volcanic series of which includes olivine-clinopyroxene phyric basalts, high-Mg andesites, and boninites [*Evarts et al.*, 1999; *Evarts et al.*, 1992; *Hopson et al.*, 1981b; *Shervais*, 1990; *Stern and Bloomer*, 1992]. The ultramafic intrusive complex includes wehrlites, clinopyroxenites, harzburgites, primitive pegmatoidal gabbros, and some isotropic gabbros and diorites (Figure 4b) [*Evarts*, 1978; *Evarts et al.*, 1999; *Hopson et al.*, 1981b]. These rocks crosscut layering and foliation in the older layered gabbro series and include xenoliths of dunite and layered gabbro. They commonly form sill-like complexes at the base of the layered gabbros as well as stocks which intrude to higher levels in the ophiolite; these high-level intrusions may form igneous megabreccias as they subsume the older layered gabbros. The stage 2 volcanic rocks include olivine-pyroxene basalts and basaltic andesites, often with Cr-

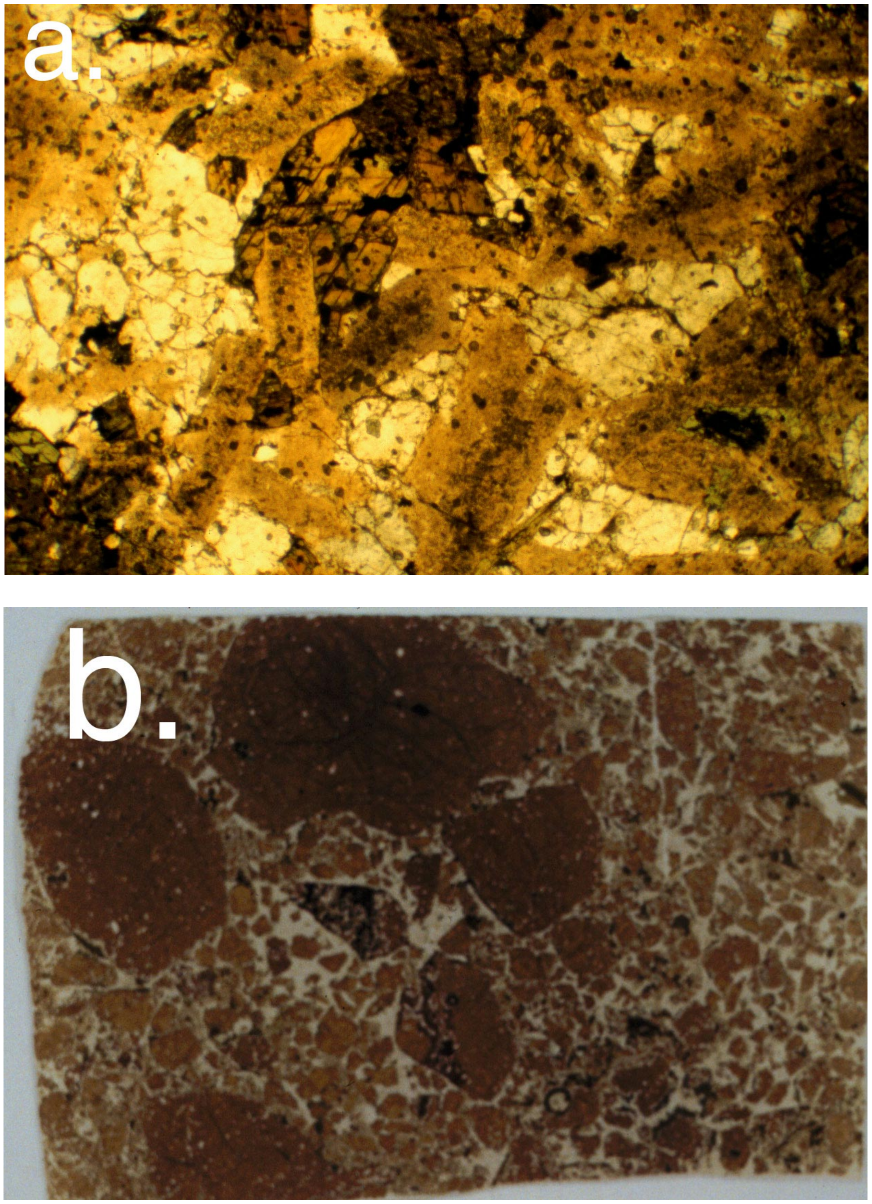


Figure 5. Photomicrographs (a) Elder Creek: Stage 3 intrusions, thin section UPL of hornblende-quartz diorite from large tabular pluton: clear, quartz; pale brown, altered feldspar, darker brown, hornblende; field of view is 5 mm, (b) Stonyford: low-power photomicrograph of volcanic glass breccia from Stonyford volcanic complex. Glass has alkali basalt composition and is younger than other volcanic rocks in the Coast Range ophiolite (stage 4); field of view is 5.5 cm.

rich spinel microphenocrysts [*Evarts et al.*, 1999; *Evarts et al.*, 1992; *Shervais*, 1990; *Stern and Bloomer*, 1992]. Rocks with boninitic affinities (high Si, Mg, Cr, Ni) occur at several Coast Range ophiolite localities, including Cuesta Ridge and Del Puerto Canyon [*Evarts et al.*, 1999; *Evarts et al.*, 1992; *Shervais*, 1990].

4.1.3. Stage 3: Maturity

[43] Stage 3 in the Coast Range ophiolite (Figure 2c) is clearly expressed in localities east of the San Andreas fault (Sacramento Valley, Diablo Range) by extensive plutons and sills of diorite, quartz diorite, and hornblende quartz diorites, and by lavas and tuffs of andesite, dacite, and rhyolite [*Evarts*, 1978; *Evarts et al.*, 1999; *Evarts et al.*, 1992; *Giaramita et al.*, 1998; *Hopson et al.*, 1981b; *Shervais*, 1990; *Shervais and Beaman*, 1991]. The intrusive rocks include sills of quartz keratophyre (andesite) up to 30 m thick, as well as stocks and sill-like plutons of hornblende diorite and hornblende quartz diorite up to 1 km thick and several kilometers long (Figure 5a). The late intrusives commonly form igneous breccias ("agmatites") in which xenoliths of dike complex, volcanic rock, older isotropic gabbro, and layered gabbro are submerged in a matrix of diorite or quartz diorite [*Shervais and Beaman*, 1991].

[44] Stage 3 volcanic rocks include pillows, flows, and volcaniclastic breccias of basaltic andesite, andesite, dacite, and rhyolite, and extensive tuffs of dacitic to rhyolitic compositions. These volcanic rocks form horizons up to 1.2 km thick on top of the older volcanic rocks [*Evarts*, 1977; *Evarts*, 1978; *Hopson et al.*, 1981b; *Lagabrielle et al.*, 1986; *Robertson*, 1989; *Robertson*, 1990]. In places (e.g., the "Crowfoot Point Breccia" at the Elder Creek ophiolite) these volcanic flows and breccias were reworked to form sedimentary breccias that unconformably overlie the ophiolite igneous complex but that are conformable with the overlying Great Valley Series [e.g., *Bailey et al.*, 1970; *Blake et al.*, 1987; *Lagabrielle et al.*, 1986; *Robertson*, 1990]. These sedimentary breccias record tectonic disruption of the ophiolite shortly after its formation and prior to deposition of sediments in the forearc basin beginning in the late Tithonian [*Bailey et al.*, 1970; *Blake et al.*, 1987; *Hopson et al.*, 1981b; *Lagabrielle et al.*, 1986; *Robertson*, 1990].

[45] West of the San Andreas and Sur Nacamiento faults, stage 3 is represented by less extensive intrusions of plagiogranite in the upper plutonic complex, by sill complexes of keratophyre, quartz gabbro, hornblende diorite, and quartz diorite, and by volcanic ash-rich radiolarian cherts that overlie the upper volcanic series [*Hopson and Frano*, 1977; *Hopson et al.*, 1981b; *Page*, 1972].

4.1.4. Stage 4: Death

[46] Death came to the Coast Range ophiolite in the northern Coast Ranges with subduction of an active spreading ridge in the earliest late Jurassic (Figure 2d). Evidence for a ridge subduction event is found throughout the Sacramento Valley and Diablo Range; this evidence includes the following: (1) late dikes with MORB geochemistry that crosscut plutonic and volcanic rocks of the earlier (stage 1 through stage 3) igneous cycles at Elder Creek [*Shervais*, 1993], Del Puerto Canyon [*Evarts et al.*, 1992], Mount Diablo [*Williams*, 1983a; *Williams*, 1983b; *Williams*, 1984], Sierra Azul [*McLaughlin et al.*, 1991; *McLaughlin et al.*, 1992], and Cuesta Ridge [*Pike*, 1974]; (2) pillow lavas with MORB geochemistry that are intercalated with sedimentary breccias above the Elder Creek ophiolite [*Shervais*, 1993]; (3) pillow lavas that compose the upper part of the ophiolite sequence at Black Mountain [*Giaramita et al.*, 1998] and Mount Diablo [*Williams*, 1983a]; (4) the Stonyford Volcanic Complex, a seamount that preserves fresh volcanic glass (Figure 5b) with ocean island basalts (OIB), alkali basalt, and high-Al basalt intercalated with radiolarian cherts similar to those that cap the Coast Range ophiolite elsewhere

[*Shervais and Hanan*, 1989]; and (5) globules of immiscible MORB-composition basaltic glass within the Leona rhyolite [*Shervais*, 1993]. This event has been dated at ≈163 Ma in the northern Coast Ranges, based on $^{39}Ar/^{40}Ar$ dates of volcanic glass from the Stonyford Volcanic Complex [*Hanan et al.*, 1991], but may be younger in the Diablo Range, where stage 3 volcanism persisted until ≈150 Ma [*Evarts et al.*, 1999].

4.1.5. Stage 5: Resurrection

[47] Resurrection of the Coast Range ophiolite did not occur by obduction, since the western margin of North America has not collided with a passive continental margin since the CRO formed. It was emplaced instead by the process I define as accretionary uplift (Figure 2e). Accretionary uplift is the process through which uplift of the ophiolite occurs in response to progressive underplating of the ophiolite by the accretionary prism in the subduction zone, represented here by the Franciscan Complex. As material is accreted in the subduction zone beneath the ophiolite, the ophiolite is progressively elevated, until gravitational collapse of the accretionary complex begins to exhume rocks subjected to high-pressure metamorphism, while the overlying ophiolite is preserved in the upper plate of the detachment fault. The implications of this process have been discussed by *Platt* [1986], *Jayko et al.* [1987], and *Harms et al.* [1992]. During its life overlying the accretionary complex the ophiolite may undergo several cycles of uplift and detachment faulting. Changes in relative plate motions and convergence directions can even lead to the onset of compressional tectonics and the juxtapositioning of units along thrust faults; these changes are characteristic of CRO tectonics during the Neogene [*Glen*, 1990; *Unruh et al.*, 1995; *Wentworth et al.*, 1984].

4.2. Case History 2: Semail Ophiolite, Oman

[48] The Semail ophiolite of Oman (Figure 6) is the second most intensely studied ophiolite in the world after the Troodos ophiolite of Cyprus [e.g., *Searle and Cox*, 1999]. It is a classic example of a Tethyan ophiolite that was obducted onto a passive continental margin and exposed by isostatic rebound of the continental margin beneath the ophiolite. Rifting of the northeastern margin of Gondwana to form the Neo-Tethys ocean began in the middle Permian and was completed by the early to mid Triassic (Figure 7a), as shown by the occurrence of Triassic seamounts of ocean island basalt with limestone carapaces in the Haybi melange [*Robertson*, 1986; *Searle and Graham*, 1982; *Searle et al.*, 1980]. This Neo-Tethyan ocean basin persisted for at least 125 million years and was not completely consumed until the late Cretaceous [*Glennie et al.*, 1974; *Lippard et al.*, 1986; *Searle and Cox*, 1999; *Yanai et al.*, 1990].

[49] Formation of the ophiolite above a NE dipping subduction zone began sometime in the pre-mid Cretaceous, as shown by U/Pb zircon dates of stage 3 quartz diorites [98-93 Ma: *Tilton et al.*, 1981]. A minimum age of 112–131 Ma for the initiation of subduction is implied by K-Ar and Ar-Ar ages of blueschist and eclogite facies rocks exposed through windows beneath the ophiolite nappe [*El-Shazly et al.*, 1990; *El-Shazly and Lanphere*, 1992; *Lippard et al.*, 1986; *Montigny et al.*, 1988; *Searle and Cox*, 1999]. Development of the Semail ophiolite is shown schematically in Figure 7a–7g.

4.2.1. Stage 1: Birth

[50] Stage 1 crust of the Semail ophiolite (Figure 7b) is represented by arc tholeiites of the Geotimes volcanic unit [*Alabaster et al.*, 1982; unit V1 of *Ernewein et al.*, 1988; *Pearce et al.*, 1981; *Umino et al.*, 1990] and by layered

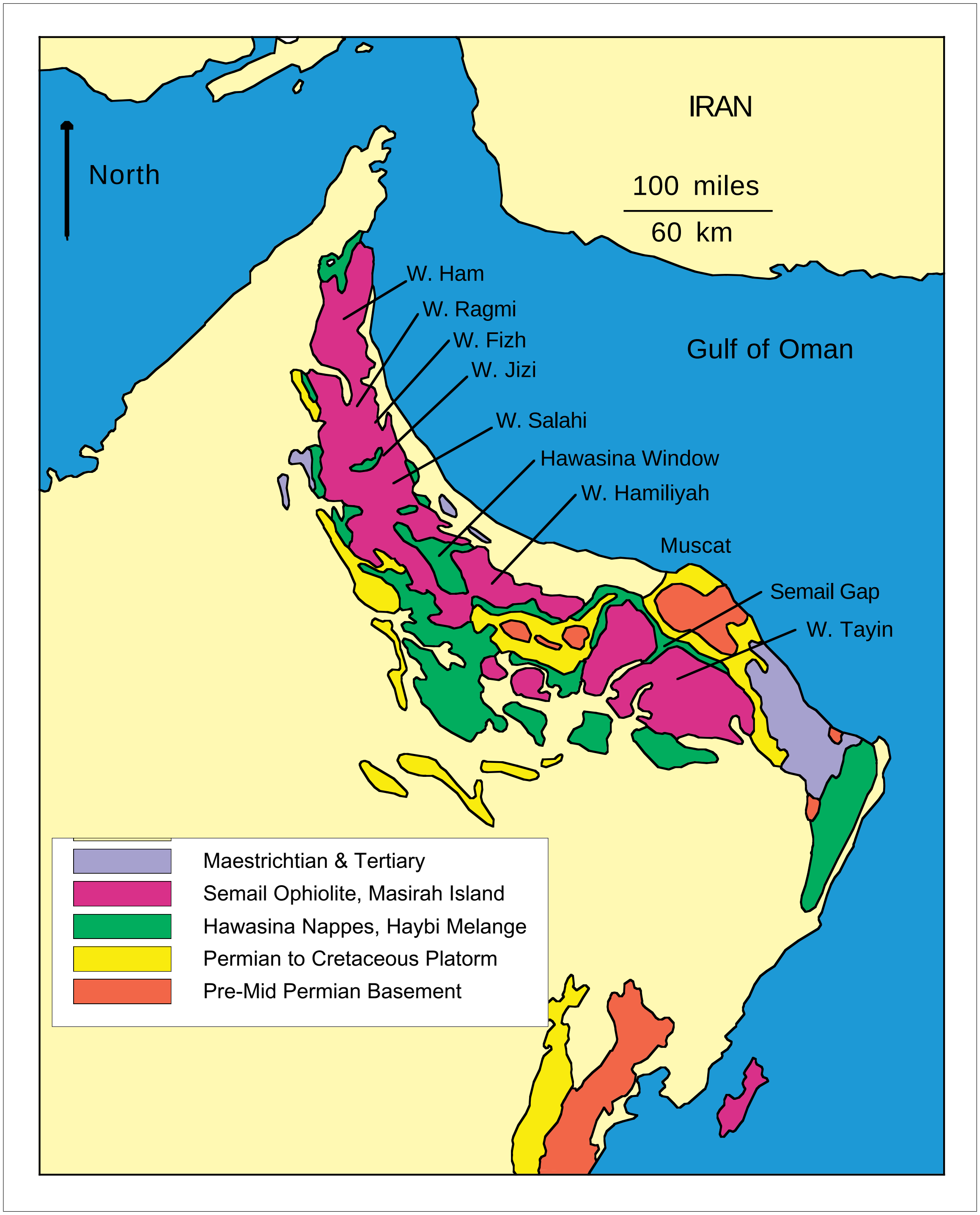


Figure 6. Location map for the Oman ophiolite, showing units and locations discussed in the text.

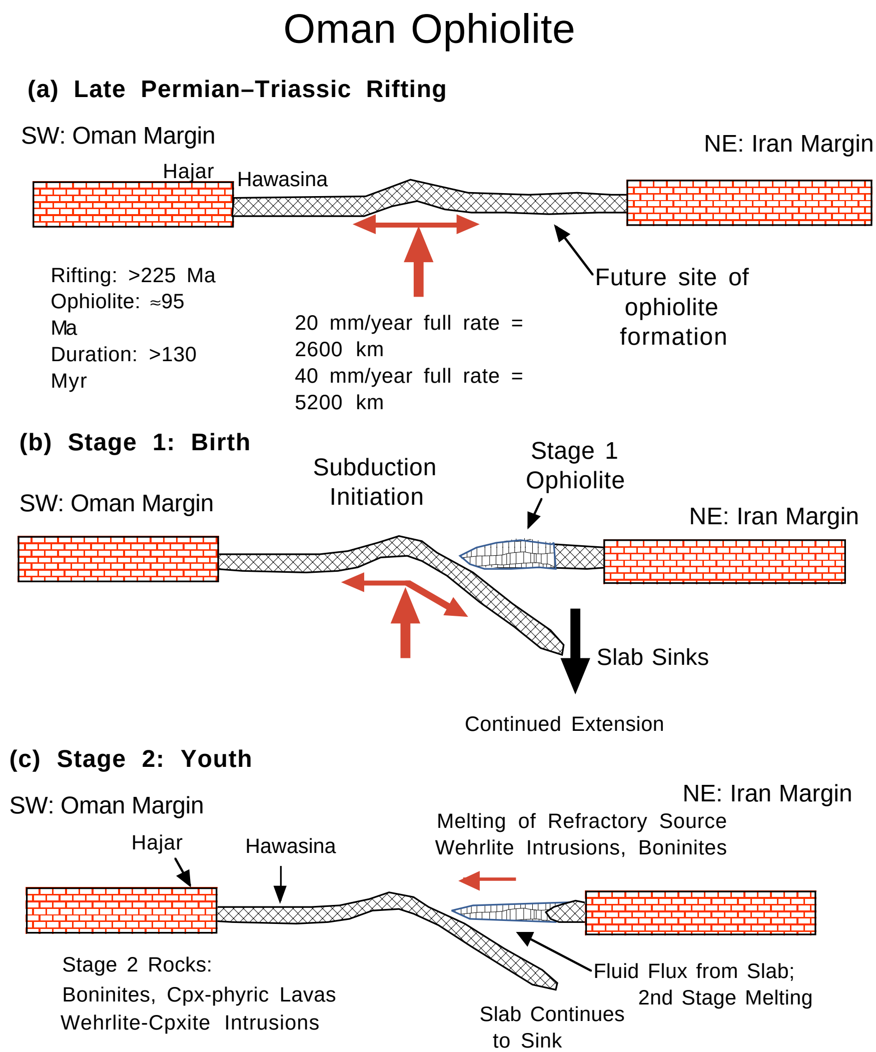


Figure 7. Generalized model for the formation and evolution of the Semail ophiolite, Oman. (a) Rifting begins in the late Permian, with formation of oceanic crust by the early to mid Triassic; duration of spreading implies ocean basin 2600–6000 km wide prior to ophiolite formation; Hajar indicates passive margin sediments deposited on stable Arabian craton, Hawasina indicates sediments deposited on the continental slope/rise and onto the abyssal plain; (b) Stage 1: birth, which is the initiation of NE dipping subduction along the Iranian margin; formation of Geotimes arc tholeiites, layered gabbros, and sheeted dikes; (c) Stage 2: youth, which is the continued extension of upper plate over sinking slab, coupled with fluid flux from slab, results in refractory melts formed by second stage melting of previously depleted asthenosphere; boninites, cpx-phyric lavas, wehrlite-pyroxenite, gabbronorite intrusions; (d) Stage 3: maturity, which is where subduction stabilizes, formation of calc-alkaline series lavas (andesites, dacites, and rhyolites of Alley volcanics; quartz diorite intrusions); (e) Stage 4: death, which is the collision of subduction zone with ridge crest; formation of high-T metamorphic sole overlaps in part with last stages of calc-alkaline magmatism; (f) Stage 4: death (continued), which shows leak-through of small volume melts of MORB asthenosphere to form Salahi volcanics, initiation of shallow overthrust of ophiolite across young, buoyant oceanic lithosphere; Stage 5: resurrection, which is the obduction of the "dead" ophiolite onto the passive continental margin; partial subduction of continental crust caused by pull of subducting oceanic lithosphere; delamination of this lithosphere results in rapid re-emergence of continental crust, with sliding of ophiolite toward the foreland.

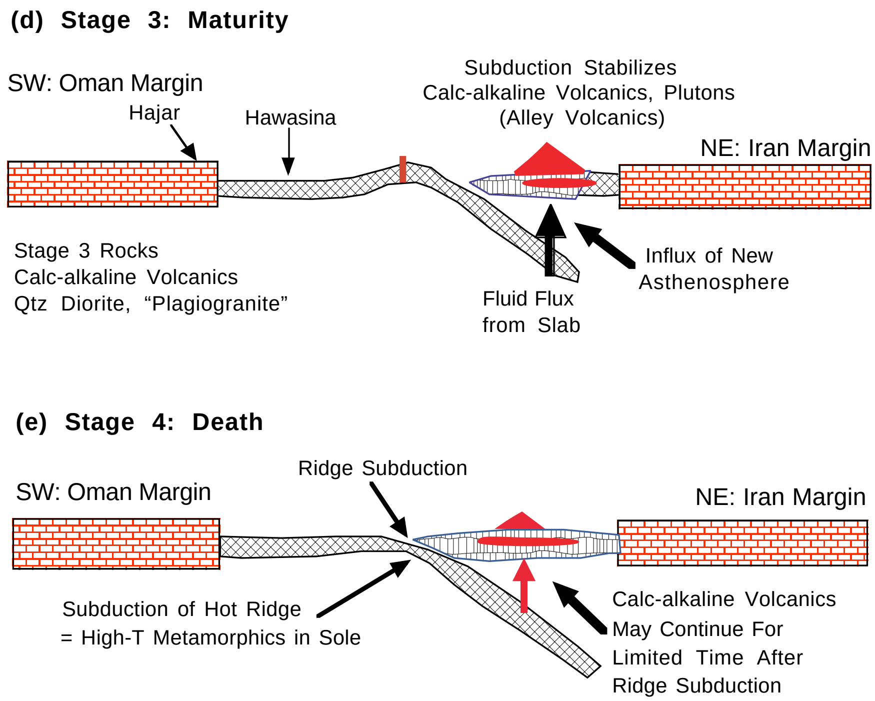


Figure 7 (continued)

gabbros of the lower crust [*Benn et al.*, 1988; *Browning*, 1984; *Ernewein et al.*, 1988; *Juteau et al.*, 1988a; *Juteau et al.*, 1988b; *Pallister and Hopson*, 1981]. Basalts of the Geotimes unit (along with most of the sheeted dike complex) have been attributed to formation at a mid-ocean spreading axis because they resemble MORB in many aspects of their geochemistry. In particular, the Geotimes basalts are low-K tholeiites with slightly depleted chondrite-normalized REE patterns, and they are associated with the sheeted dike complex [*Alabaster et al.*, 1982; *Lippard et al.*, 1986; *Pallister and Knight*, 1981; *Pearce et al.*, 1981; *Umino*, 1995; *Umino et al.*, 1990]. These basalts are enriched in the LILE Sr, Rb, K, Ba, and Th and depleted in the compatible elements Sc and Cr [*Alabaster et al.*, 1982; *Umino et al.*, 1990]. The high field strength elements Ta, Nb, Zr, Hf, Ti, Y, and Yb have flat MORB-normalized concentrations at $1\times$ to $2\times$ normal MORB [*Alabaster et al.*, 1982; *Umino et al.*, 1990]. Because the LILE (except Th) may be mobile during seafloor alteration, these elements cannot be used to establish the provenance of the Geotimes lavas. Ratio plots like Th/Yb versus Ta/Yb show that the Geotimes basalts fall in the arc tholeiite field, along with most other volcanic units of the Semail ophiolite [*Alabaster et al.*, 1982; *Pearce et al.*, 1981], although this interpretation may be questioned (P. Kelemen, personal communication, 2000).

[51] Gabbros associated with the stage 1 magma series are layered on a centimetric to decimetric scale, with layers defined by variations in modal proportions, by the appearance or loss of a cumulus phase, and by variations in the size of the cumulus crystals [*Benn et al.*, 1988; *Browning*, 1984; *Ernewein et al.*, 1988; *Juteau et al.*, 1988a, 1988b; *Pallister and Hopson*, 1981]. The layers display both synmagmatic deformation textures (e.g., ductile

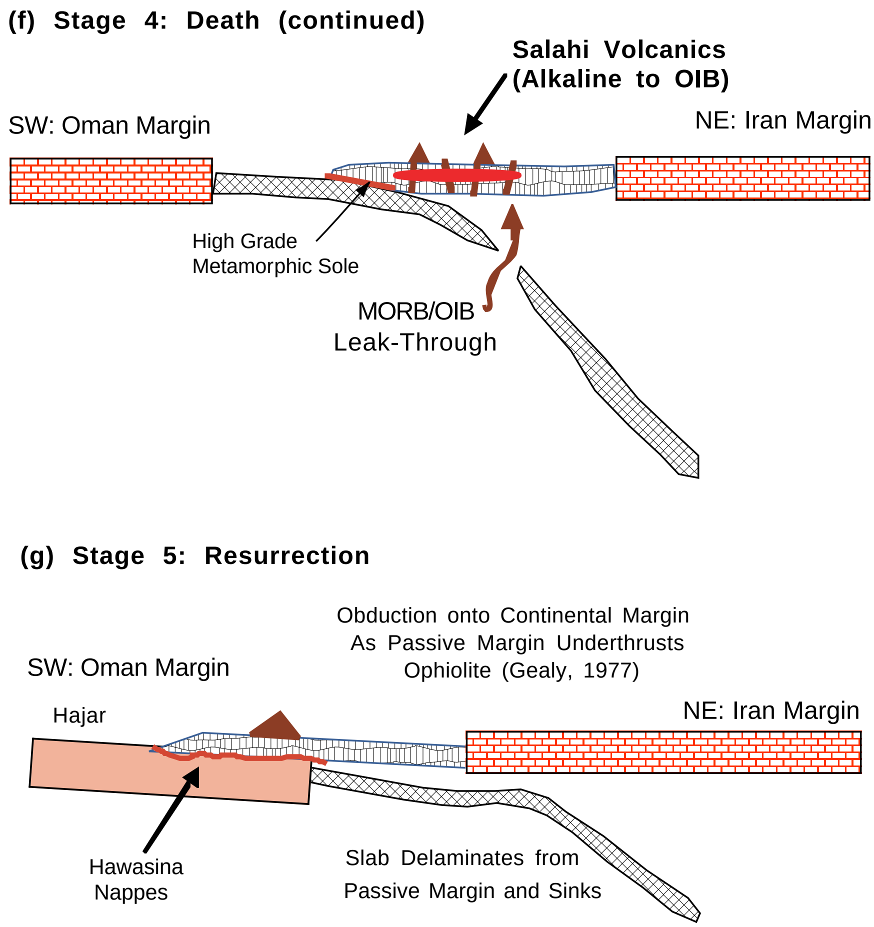


Figure 7 (continued)

normal faults with amphibolite facies minerals [*Ernewein et al.*, 1988; *Lippard et al.*, 1986]) and subsolidus ductile deformation (Figure 8a). Most of the layered gabbroic cumulates display mineral assemblages consistent with formation from a magma similar to the Geotimes volcanic unit, but some sections (e.g., Wadi Ragmi) contain common cumulus orthopyroxene and appear to be related to the later stage 2 magma series [*Lippard et al.*, 1986].

4.2.2. Stage 2: Youth

[52] Stage 2 in the Semail ophiolite (Figure 7c) is represented most clearly by ultramafic rocks that intrude the earlier layered gabbros [*Benn et al.*, 1988; *Ernewein et al.*, 1988; *Juteau et al.*, 1988a; *Juteau et al.*, 1988b; *Lippard et al.*, 1986] and by olivine-clinopyroxene-phyric and orthopyroxene-phyric volcanic rocks that overlie the Geotimes volcanic unit [*Alabaster et al.*, 1982; *Ernewein et al.*, 1988; *Umino et al.*, 1990].

[53] Plutonic rocks associated with stage 2 include dunite, wehrlite, pyroxenite, troctolite, and pegmatoidal gabbro [*Benn et al.*, 1988; *Juteau et al.*, 1988a; *Juteau et al.*, 1988b; *Lippard et al.*, 1986; *Pallister and Hopson*, 1981]. The ultramafic intrusive series forms stocks and dikes that crosscut the layered series at high angles (Figure 8b), as well as

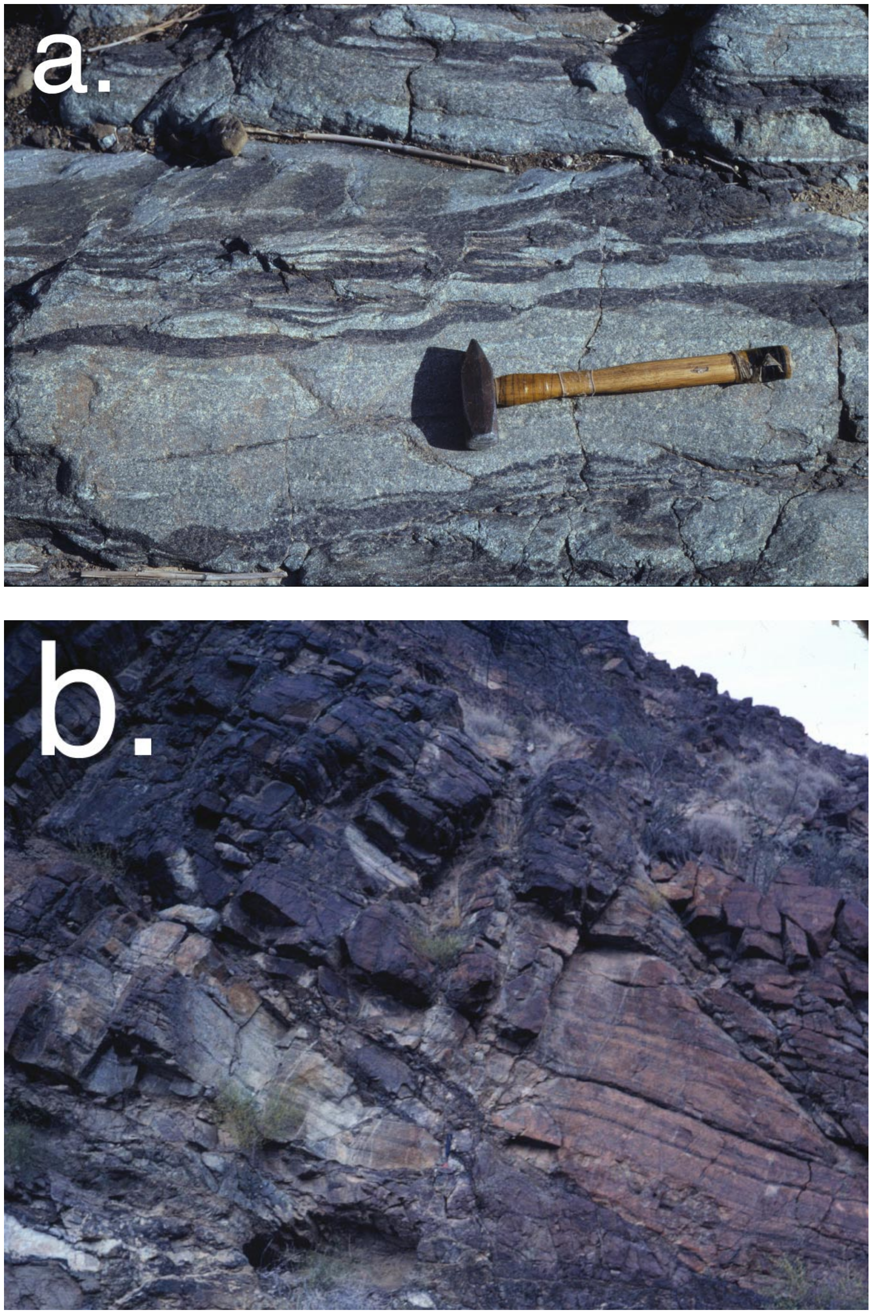


Figure 8. Field photos of the Semail ophiolite, Oman. (a) Deformed gabbro with boudinaged mafic layers, Wadi Aysaybah; (b) wehrlite-dunite intrusion into layered gabbro, megabreccia, Wadi Aysaybah; (c) oblique air photo of large plagiogranite intrusion, near Murayat; (d) Plagiogranite agmatite, with xenoliths of diabase. (All photos by John Pallister, ©John Pallister, 1978–1981).

sills that penetrate parallel to the older layering and mimic cumulus layering [*Benn et al.*, 1988; *Juteau et al.*, 1988a; *Juteau et al.*, 1988b; *Umino et al.*, 1990]. Early studies

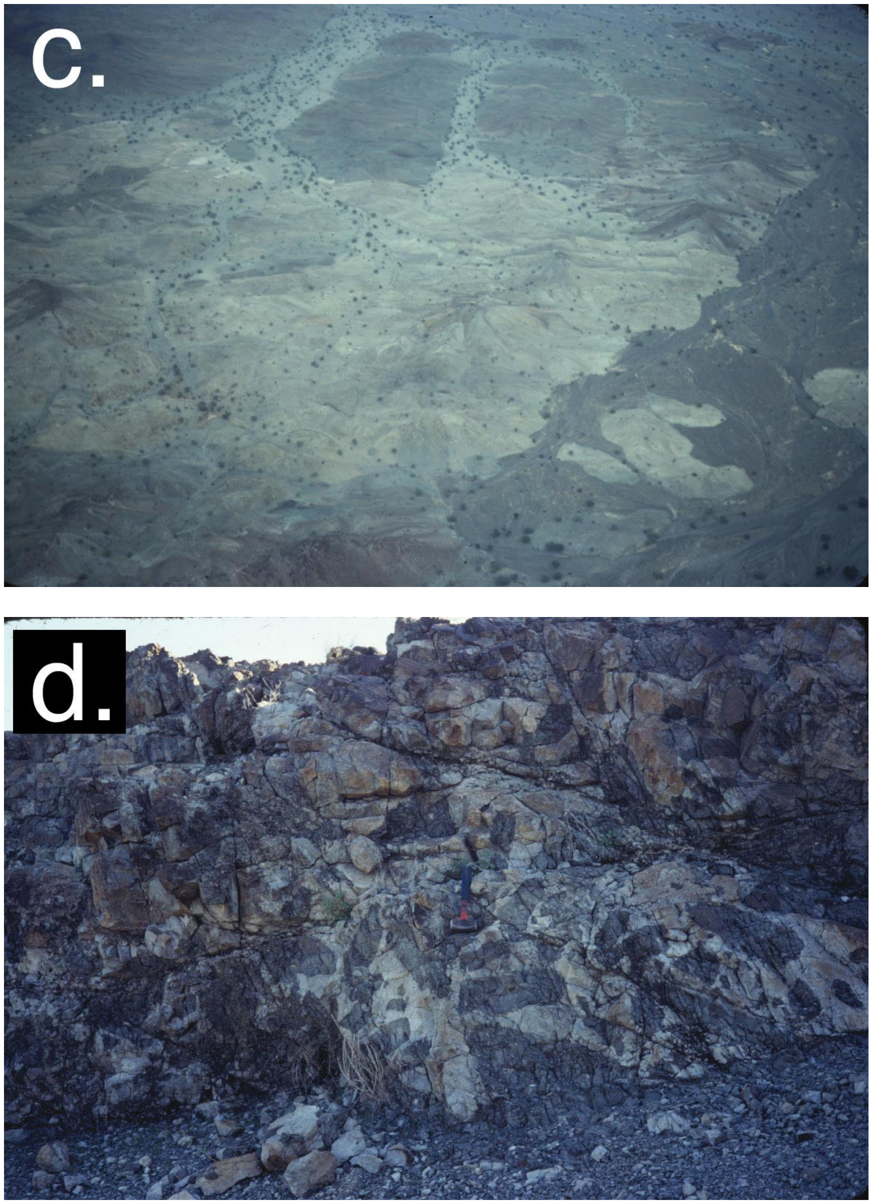


Figure 8 (continued)

recognized the intrusive nature of these ultramafic rocks but attributed their origin to "mobilization" of preexisting cumulates, rich in interstitial magma, that formed at deeper levels in the cumulate pile [e.g., *Lippard et al.*, 1986; *Pallister and Hopson*, 1981]. Later work has shown that these rocks represent a separate, refractory magma series characterized by more magnesian mafic minerals and more calcic feldspars than the earlier layered cumulates [*Benn et al.*, 1988; *Juteau et al.*, 1988a; *Juteau et al.*, 1988b; *Umino et al.*, 1990]. *Umino et al.* [1990] recognize two suites within the stage 2 intrusives, a Cpx series of dunite, wehrlite, olivine gabbro, and diorite, and an Opx series of lherzolites, gabbronorites, two-pyroxene diorites, and trondhjemites; they correlate both suites with subduction-related magmatism and with volcanic rocks of the Alley unit. In contrast to rocks of the layered series (stage 1), the stage 2 intrusives are generally undeformed.

[54] Volcanic rocks that correlate with the stage 2 intrusive series include olivine + clinopyroxene phyric lavas of the Lasail and Cpx-phyric units of *Alabaster et al.* [1982], and orthopyroxene-phyric lavas of the Alley unit, as described by *Umino et al.* [1990]. *Umino et al.* [1990] suggest that these lavas generally underlie the more evolved dacite-rhyolite lavas of the Alley unit (which I associate with stage 3), but it is also clear that there is considerable intercalation of the more evolved lavas with the more refractory lavas and that the change from stage 2 "refractory magmas" to stage 3 "evolved magmas" was not a discrete event, but a gradual change from one magma series to another.

4.2.3. Stage 3: Maturity

[55] Stage 3 in the Semail ophiolite (Figure 7d) is represented by late intrusive complexes of gabbro-diorite-trondhjemite and by andesite-dacite-rhyolite of the Alley volcanic suite, as defined by *Umino et al.* [1990]. The late intrusive complexes of stage 3 are linked in places to stocks of more refractory magmas but in others are restricted to more evolved rocks. These rocks (which include the plagiogranite of many authors) intrude the older cumulate section and the overlying sheeted dike complex. They form both stocks and dikes but are most common as large, tabular, sill-like intrusions oriented perpendicular to the overlying sheeted complex that range up to 10 km^2 in size (Figure 8c) [*Lippard et al.*, 1986]. The late intrusions commonly form igneous breccias with xenoliths of isotropic gabbro and dike complex submersed in a matrix of diorite or quartz diorite (Figure 8d).

[56] Intermediate to evolved volcanic rocks are common in the Alley unit and appear to dominate the upper stratigraphic levels of this unit [e.g., *Alabaster et al.*, 1982; *Ernewein et al.*, 1988; *Pearce et al.*, 1981; *Umino et al.*, 1990]. These rocks (andesite, dacite, and rhyolite) are the extrusive equivalents of the late intrusive complexes (diorite, quartz diorite, trondhjemite). Plutonic rocks of stage 3 exhibit calc-alkaline differentiation trends and isotopic characteristics [*Briqueu et al.*, 1991; *Lachize et al.*, 1996; *Lippard et al.*, 1986]. The onset of calc-alkaline magmatism following the prolonged intrusion and eruption of stage 2 refractory magmas indicates a transition to a more normal "subduction" regime and the establishment of an incipient volcanic arc complex.

4.2.4. Stage 4: Death

[57] Active formation of the Semail ophiolite ended when it consumed the Neo-Tethys spreading center (Figure 7e). This event is represented by a metamorphic sole that underlies the Semail nappe and separates peridotites of the ophiolite (above) from accretionary rocks of the underlying Haybi melange [*Glennie et al.*, 1974]. Metamorphic rocks of the sole include amphibolites, garnet amphibolites, and granulites, all of which formed on high geothermal gradients with counterclockwise pressure-temperature-time (*P-T-t*) paths [*Ghent and Stout*, 1981; *Gnos*, 1998; *Hacker and Mosenfelder*, 1996; *Searle and Cox*, 1999; *Searle and Malpas*, 1980]. These rocks formed $\approx$94±2 Ma [*Hacker et al.*, 1996; *Lanphere*, 1981]; that is, only slightly after the intrusion of some stage 3 quartz diorites [*Tilton et al.*, 1981].

[58] *Hacker et al.* [1996] show that formation of the metamorphic sole must coincide with ridge subduction because the geothermal gradient

expressed by the amphibolites, garnet amphibolites, and granulites of the metamorphic sole is too high to represent an established subduction zone. As discussed earlier, several authors have suggested that subduction was initiated by thrusting that nucleated at the spreading center [*Boudier et al.*, 1988; *Coleman*, 1981; *Hacker and Gnos*, 1997; *Hacker et al.*, 1996]. It has been shown that this is physically unlikely, if not impossible, because oceanic lithosphere <10 m.y. in age is too buoyant to subduct [*Cloos*, 1993].

[59] There are additional arguments against this model in Oman, as discussed recently by *Searle and Cox* [1999]. If the ophiolite formed by this "ridge-centered thrust" model, then amphibolites of the metamorphic sole and volcanics of the ophiolite should have the same geochemical characteristics and form in the same tectonic setting (at a mid-ocean ridge spreading center). In Oman this is clearly not the case; abundant data show that amphibolites of the metamorphic sole are oceanic tholeiites, similar to Jurassic tholeiitic basalts in the Haybi melange, and are unrelated to rocks of the ophiolite [e.g., *Alabaster et al.*, 1982; *Lippard et al.*, 1986; *Searle and Cox*, 1999; *Searle and Graham*, 1982; *Searle et al.*, 1980].

[60] Ridge collision/subduction ended the formation of normal ophiolitic crust (layered gabbros, ultramafic intrusives, calc-alkaline stocks and sills) because the young oceanic lithosphere was too buoyant to subduct, and the ophiolite entered a phase of intraoceanic thrusting along a shallow decollement (Figure 7f). Generation of suprasubduction zone magmas ceased because there was no mantle wedge in which to generate the melts. Instead, leakage of off-axis, intraplate oceanic basalts and alkali basalts from the lower plate asthenosphere into the upper plate lithosphere fed eruptions of the Salahi volcanics during the Coniacian, between ≈85 and 90 Ma (Figure 7f) [*Alabaster et al.*, 1982; *Lippard et al.*, 1986; *Umino et al.*, 1990]. These lavas are separated from the underlying Alley volcanics by ≈15 m of pelagic sediment [*Ernewein et al.*, 1988], consistent with a time gap of several million years between the last SSZ ophiolite lavas and eruption of the Salahi basalts. This time gap may represent shallow underthrusting of the NE flank of the spreading axis, prior to its subduction beneath the ophiolite.

[61] Petrologic and geochemical studies show that the Salahi volcanics are intraplate ocean island basalts and transitional alkali basalts that formed off-axis, after the formation of the underlying ophiolite assemblage [*Alabaster et al.*, 1982; *Ernewein et al.*, 1988; *Lippard et al.*, 1986; *Umino et al.*, 1990]. Diabase dikes with similar compositions cut amphibolites of the metamorphic sole but do not extend down into the underlying sediments and sedimentary melange [*Ghent and Stout*, 1981], showing that "leak-through" of oceanic magmas continued even after shallow thrusting across the spreading center commenced [*Ernewein et al.*, 1988].

4.2.5. Stage 5: Resurrection

[62] Resurrection of the Semail ophiolite by obduction onto the passive margin of the Arabian platform (Figure 7g) did not occur until the mid-Campanian (≈75 Ma), as attested by deposition of the Juweiza Formation, a *syn*-emplacement flysch deposit [*Glennie et al.*, 1974; *Lippard et al.*, 1986; *Yanai et al.*, 1990]. This implies ≈20 million years of convergence after ridge subduction before emplacement of the Semail nappe. If the thrusting associated with this convergence was relatively rapid (≈50 mm/yr), then ≈1000 km of oceanic lithosphere must have been subducted during this time. Since ridge subduction coincided with opening of the South Atlantic around 100 Ma [*Fairhead and Binks*, 1991; *Schult*

and Guerreiro, 1980; *Turner et al.*, 1994], the actual rate of convergence may have been much higher.

[63] The extent of Neotethyan oceanic crust prior to its subduction is an issue of some controversy. It has been suggested that Tethyan ophiolites like Oman formed in small, intracontinental ocean basins only a few hundred kilometers wide (“Red Sea model [*Lippard et al.*, 1986; *Robertson*, 1986]). In the case of Oman this claim is difficult to understand in light of the fact that the Haybi melange contains late Permian to Triassic oceanic basalts and that sedimentary records show that rifting to form the passive margin of Oman began in the middle Permian and was complete by the early Triassic at the latest [*Gealey*, 1977; *Glennie et al.*, 1974; *Searle and Graham*, 1982; *Searle et al.*, 1980; *Searle and Malpas*, 1980]. If we assume a modest spreading rate of only 40 mm/yr (20 mm/yr half rate) and 150 million years of spreading (early Triassic at 245 Ma to late Cretaceous at 95 Ma), then at least 6000 km of ocean basin must have existed! Even assuming a more conservative 100 million years of spreading (late Triassic at 195 Ma to late Cretaceous at 95 Ma), then at least 4000 km of ocean basin must have existed. Both of these estimates would be much higher if faster rates were assumed. There is no evidence for pre-Cretaceous arc terranes that could have consumed this Neo-Tethys crust along its northern margin, so it is suggested that an ocean basin at least 4000 km wide and probably much wider must have existed prior to formation of the Semail ophiolite.

[64] The final stage of ophiolite emplacement in Oman is well documented and will not be discussed here in detail [*Bechennec et al.*, 1988; *Cawood*, 1991; *Gealey*, 1977; *Gregory et al.*, 1998; *Miller et al.*, 1998; *Searle and Cox*, 1999]. It is noted, however, that rapid emergence of the Semail ophiolite after subduction of the Arabian continental margin is consistent with detachment of the oceanic portion of the passive plate margin (Figure 7g). Detachment of the oceanic lithosphere from the descending slab releases the slab pull that drives subduction and will result in a rapid change from negative to positive buoyancy in the partially subducted continental crust.

5. Conclusions

[65] It is clear from the preceding discussion that suprasubduction zone ophiolites display a consistent sequence of events during their formation and evolution that demonstrates that they must form in response to processes that are common to all such ophiolites and that are characteristic of their mode of formation. These processes represent the normal response of oceanic lithosphere to the initiation of new or renewed subduction, as outlined by *Stern and Bloomer* [1992] and here. It should be stressed that this model does not apply to all ophiolitic rocks: many ophiolitic assemblages clearly formed at mid-ocean ridges or seamounts and were accreted within a subduction complex (e.g., the Ligurian ophiolites and ophiolitic rocks within the Franciscan Complex, California), while others represent true “back arc basin” assemblages. This model applies only to “true” SSZ ophiolites that form by forearc or intra-arc extension.

[66] A corollary of this theorem is that SSZ ophiolite formation is not a stochastic event but the predictable consequence of processes and tectonic events that recur systematically. This implies that ad hoc models of ophiolite formation that rely on special circumstances or nonuniformitarian processes are suspect. In particular, there is no reason to believe that the situations in which oceanic basalts erupted at a mid-ocean ridge may have some geochemical characteristics of arc tholeiites (e.g., South Chile Rise [*Klein and Karsten*, 1995])

will result in oceanic crust that is any more likely to be preserved than ordinary, "true MORB" oceanic crust. Some SSZ ophiolites may form by "trapping" oceanic crust behind nascent subduction zones, but it seems highly unlikely that the only crust that would be trapped is anomalous, arc-contaminated crust formed during some earlier cycle of subduction. Further, such a mechanism offers no explanation of the other magma series associated with all SSZ ophiolites (boninites, calc-alkaline lavas) nor the consistent sequence in which these suites form.

[67] Another implication of this model is the need for combined field, petrologic, and geochemical studies of many ophiolites to elucidate the similarities and differences among them. Detailed studies that focus on a single, well-exposed example can tell us about small-scale processes that may be active in many different tectonic environments but do not address fundamental questions about larger-scale tectonic processes. We should also note here that evidence for several of the stages proposed here is based largely on field relations, such as crosscutting intrusive relationships or stratigraphic superposition, and is not dependent on geochemical discrimination diagrams. This reaffirms the importance of well-defined field relationships as the basis for any geochemical study.

[68] Finally, we should not overlook the most important corollary of this model: that many "ophiolites" will eventually evolve into mature island arc systems, in which the older "ophiolitic" stage of formation is obscured by a carapace of younger volcanics, and by intrusion of later plutons. This seems to be the case in the western Pacific, where ophiolitic rocks are now found in the forearc basement of the Marianas and Bonin arcs. Ophiolites will only be preserved where the "ophiolite-forming" process is arrested during its early stages of development ("death"), and where later tectonic events lead to exposure through obduction or accretionary uplift ("resurrection"). These conditions are typically found in collisional environments, which are thus the most common setting for preservation of ophiolites.

Acknowledgments

[69] This paper is an outgrowth of ideas conceived during the 1998 Penrose Workshop "Ophiolites and Oceanic Lithosphere: New Insights From Field Studies and Ocean Drilling Program", convened by Yildrim Dilek, Eldridge Moores, Don Elthon, and Adolphe Nicolas. It was at this meeting that I was struck by the consistent petrologic succession found among all SSZ ophiolites throughout the world and by the implications of this consistent succession for the nonstochastic formation of ophiolites. I would also like to thank Hubert Staudigel for inviting me to participate in the GERM symposium at Fall AGU in 1999 and to present these ideas here. I am also grateful to John Pallister, USGS, for permission to use unpublished field photographs of the Semail ophiolite from his personal collection. This work was funded in part by NSF grants EAR8816398 and EAR9018721.

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