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Continental growth at convergent margins facing large ocean basins: a case study from Mesozoic convergent-margin basins of Baja California, Mexico

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Abstract

Mesozoic rocks of the Baja California Peninsula form one of the most areally extensive, best-exposed, longest-lived (160 my), least-tectonized and least-metamorphosed convergent-margin basin complexes in the world. This convergent margin shows an evolutionary trend that may be typical of arc systems facing large ocean basins: a progression from highly extensional (phase 1) through mildly extensional (phase 2) to compressional (phase 3) strain regimes. This trend is largely due to the progressively decreasing age of lithosphere that is subducted, which causes a gradual decrease in slab dip angle (and concomitant increase in coupling between lower and upper plates), as well as progressive inboard migration of the arc axis.

This paper emphasizes the usefulness of sedimentary and volcanic basin analysis for reconstructing the tectonic evolution of a convergent continental margin. Phase 1 consists of Late Triassic to Late Jurassic oceanic intra-arc to backarc basins that were isolated from continental sediment sources. New, progressively widening basins were created by arc rifting and sea floor spreading, and these were largely filled with progradational backarc arc-apron deposits that record the growth of adjacent volcanoes up to and above sea level. Inboard migration of the backarc spreading center ultimately results in renewed arc rifting, producing an influx of silicic pyroclastics to the backarc basin. Rifting succeeds in conversion of the active backarc basin into a remnant backarc basin, which is blanketed by epiclastic sands.

Phase 1 oceanic arc–backarc terranes were amalgamated by Late Jurassic sinistral strike slip faults. They form the forearc substrate for phase 2, indicating inboard migration of the arc axis due to decrease in slab dip. Phase 2 consists of Early Cretaceous extensional fringing arc basins adjacent to a continent. Phase 2 forearc basins consist of grabens that stepped downward toward the trench, filled with coarse-grained slope apron deposits. Phase 2 intra-arc basins show a cycle of (1) arc extension, characterized by intermediate to silicic explosive and effusive volcanism, culminating in caldera-forming silicic ignimbrite eruptions, followed by (2) arc rifting, characterized by widespread dike swarms and extensive mafic lavas and hyaloclastites. This extensional-rifting cycle was followed by mid-Cretaceous backarc basin closure and thrusting of the fringing arc beneath the edge of the continent, caused by a decrease in slab dip as well as a possible increase in convergence rate.

Phase 2 fringing arc terranes form the substrate for phase 3, which consists of a Late Cretaceous high-standing, compressional continental arc that migrated inboard with time. Strongly coupled subduction resulted in accretion of blueschist metamorphic rocks, with development of a broad residual forearc basin behind the growing accretionary wedge, and development of extensional forearc (trench–slope) basins atop the gravitationally collapsing accretionary wedge. Inboard of

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this, ongoing phase 3 strongly coupled subduction, together with oblique convergence, resulted in development of forearc strike-slip basins upon arc basement.

The modern Earth is strongly biased toward long-lived arc–trench systems, which are compressional; therefore, evolutionary models for convergent margins must be constructed from well-preserved ancient examples like Baja California. This convergent margin is typical of many others, where the early to middle stages of convergence (phases 1 and 2) create nonsubductable arc–ophiolite terranes (and their basin fills) in the upper plate. These become accreted to the continental margin in the late stage of convergence (phase 3), resulting in significant continental growth.

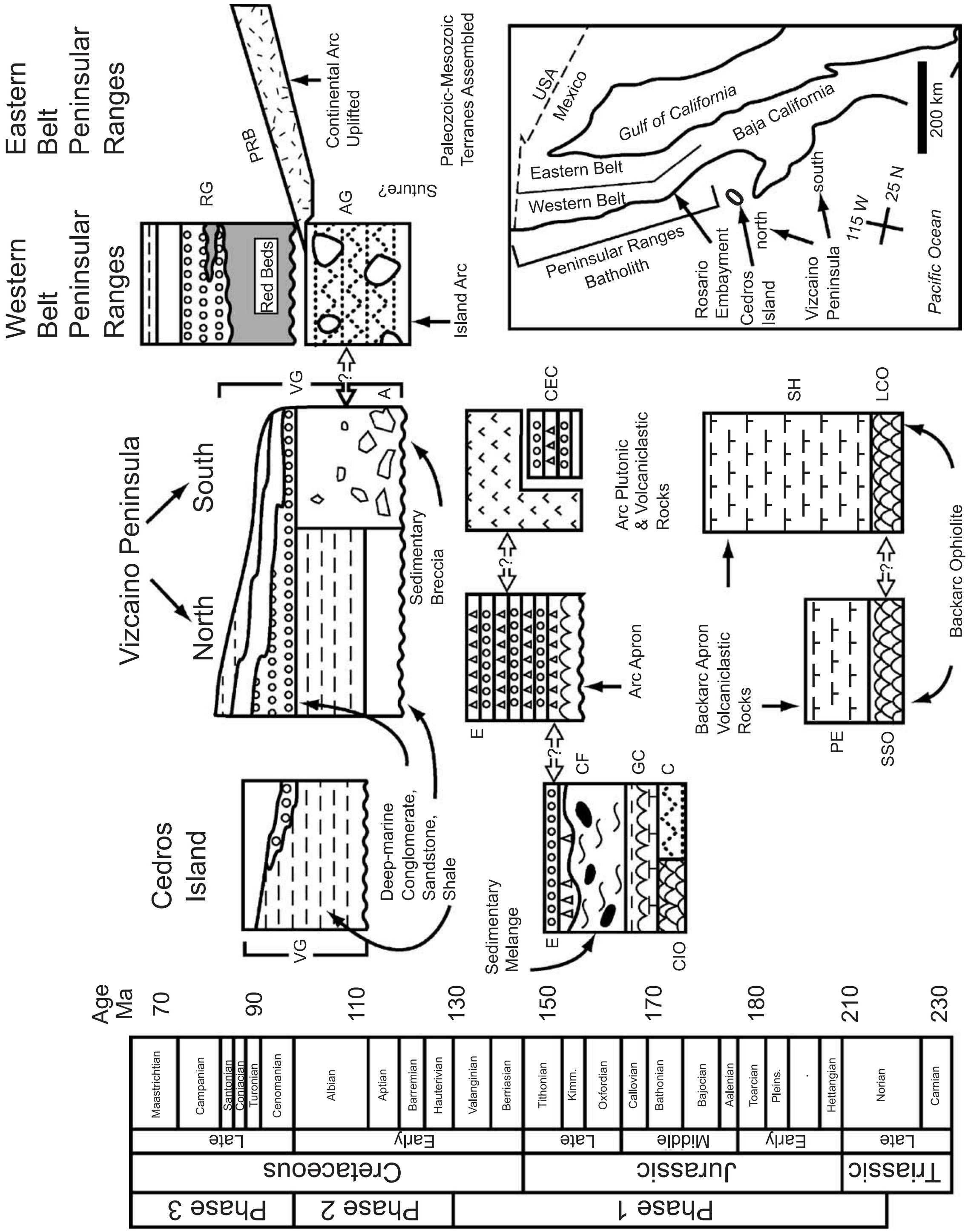
Keywords: Convergent margin; Intra-arc basins; Backarc basins; Forearc basins; Mexico

1. Introduction

Mesozoic rocks of the Baja California Peninsula (Fig. 1) form one of the best-exposed and longest-lived convergent-margin basin complexes in the world. These rocks are well exposed in the Peninsular ranges of northern Baja California and in the Vizcaino Peninsula–Cedros Island region of western Baja California, whereas to the south and southeast, they are largely buried under volcanic and sedimentary rocks of the Late Cenozoic Gulf Extensional Province. Mesozoic arc magmatism shifted inboard (toward the continent) with time (Fig. 1), from the Vizcaino–Cedros region (Late Triassic to Jurassic intra-oceanic, arc–ophiolite assemblages) to the western Peninsular Ranges (Early Cretaceous fringing oceanic arc) to the eastern Peninsular Ranges and adjacent mainland Mexico (Late Cretaceous continental arc).

This convergent margin shows an overall evolutionary trend that I suggest may be typical of arc systems facing large ocean basins: a progression from highly extensional through mildly extensional to compressional strain regimes (Fig. 2). In this evolutionary model, subduction is initiated by rapid sinking of very old, cold oceanic lithosphere, but over many tens of millions of years, the age of the lithosphere that reaches the trench and is subducted becomes progressively younger (Busby et al., 1998). This results in a gradual decrease in slab dip angle, causing an increase in coupling between the lower and upper plates, as well as progressive inboard

Fig. 1. Tectonstratigraphic chart and locations of Mesozoic convergent-margin assemblages in Baja California, Mexico, grouped by evolutionary phases 1, 2, and 3 (see Fig. 2). Generalized geologic map of part of the Peninsular Ranges, of the Vizcaino Peninsula and of Cedros Island are presented in Figs. 3, 4 and 5. Phase 1—*South Vizcaino Peninsula*: La Costa ophiolite (LCO) and San Hipolito Formation (SH) (Whalen and Pessagno, 1984; Moore, 1985); volcaniclastic strata of Cerro El Calvario (CEC) (Moore, 1984, 1985). *North Vizcaino Peninsula*: Sierra San Andres ophiolite (SSO) and overlying Puerto Escondido tuff (PE) (Barnes, 1984; Moore, 1985); Eugenia Formation (E) (Hickey, 1984; Kimbrough et al., 1987). *Cedros Island*: Cedros Island ophiolite (CIO) and Choyal oceanic arc assemblage (C), overlapped by Gran Canon Formation backarc apron volcaniclastic rocks (GC) (Kimbrough, 1984, 1985; Busby-Spera, 1987, 1988a,b; Critelli et al., 2002); Coloradito Formation (CF) (Kilmer, 1984; Boles and Landis, 1984); Eugenia Formation (EF) (Kilmer, 1984). Phase 2—*Vizcaino Peninsula and Cedros Island*: forearc strata, including Asuncion Formation (A) and lower Valle Group (VG) (Barnes, 1984; Moore, 1984, 1985; Patterson, 1984; Busby-Spera and Boles, 1986; Smith et al., 1991; Smith and Busby, 1993). *Western Belt Peninsular Ranges*: Alisitos Group (AG) oceanic arc assemblage (Allison, 1955; Silver et al., 1963, 1979; Gastil et al., 1975; Beggs, 1984; Gastil, 1985; Silver, 1986; Busby-Spera and White, 1987; White and Busby-Spera, 1987; Adams and Busby, 1994; Fackler Adams and Busby, 1998); associated backarc strata on continental-margin substrate (Paleozoic–Mesozoic terranes of eastern belt of Peninsular Ranges) (Gastil, 1985; Griffith, 1987; Goetz et al., 1988; Goetz, 1989). Phase 3—*Vizcaino Peninsula and Cedros Island*: forearc strata, upper Valle Group (VG) (Barnes, 1984; Patterson, 1984; Morris et al., 1989; Smith and Busby, 1994), underthrust by blueschist-grade subduction complexes (Figs. 4 and 5; Sedlock, 1988; Smith et al., 1991; Sedlock et al., 1993). *Western Peninsular Belt Ranges*: forearc strata, Rosario Group (RG) (Kilmer, 1963; Gastil and Allison, 1966; Nilsen and Abbott, 1981; Bottjer and Link, 1984; Yeo, 1984; Boehlke and Abbott, 1986; Cunningham and Abbott, 1986; Morris and Busby-Spera, 1988, 1990; Morris et al., 1989; Filmer and Kirschvinck, 1989; Morris, 1992; Abbott et al., 1993; Fulford and Busby-Spera, 1993; Morris and Busby, 1996; Busby-Spera et al., 1988). *Eastern Belt Peninsular Ranges*: continental-arc plutons record an eastward-migrating linear locus of magmatism from ca. 100 to 75 Ma (Krummenacher et al., 1975; Silver, 1986; Silver and Chappal, 1988), with a major pulse of “La Posta type” plutons (Walawander et al., 1990) at about 98 Ma (Kimbrough et al., 2001); intrudes Paleozoic to Mesozoic continental margin terranes (Gastil and Miller, 1981; Gastil, 1985).


Phase 1
Phase 2
Phase 3
Triassic
Jurassic
Cretaceous
Late
Early
Middle
Carnian
Norian
Hettangian
Pleins.
Toarcian
Aalenian
Bajocian
Bathonian
Callovian
Oxfordian
Kimm.
Tithonian
Berriasian
Valanginian
Hauterivian
Barremian
Aptian
Albian
Cenomanian
Turonian
Coniacian
Santonian
Campanian
Maastrichtian
Age Ma
70
90
110
130
150
170
180
210
230
Cedros Island
VG
Deep-marine Conglomerate, Sandstone, Shale
Sedimentary Melange
E
CF
GC
C
CIO
Vizcaino Peninsula
North
South
Sedimentary Breccia
A
E
Arc Apron
CEC
Arc Plutonic & Volcaniclastic Rocks
Backarc Apron Volcaniclastic Rocks
PE
SSO
Backarc Ophiolite
SH
LCO
Western Belt Peninsular Ranges
Eastern Belt Peninsular Ranges
RG
Red Beds
AG
Island Arc
PRB
Continental Arc Uplifted
Suture?
Paleozoic-Mesozoic Terranes Assembled
USA
Mexico
Gulf of California
Baja California
Eastern Belt
Western Belt
Peninsular Ranges Batholith
Rosario Embayment
Cedros Island
north
Vizcaino Peninsula
south
115 W
25 N
200 km
Pacific Ocean

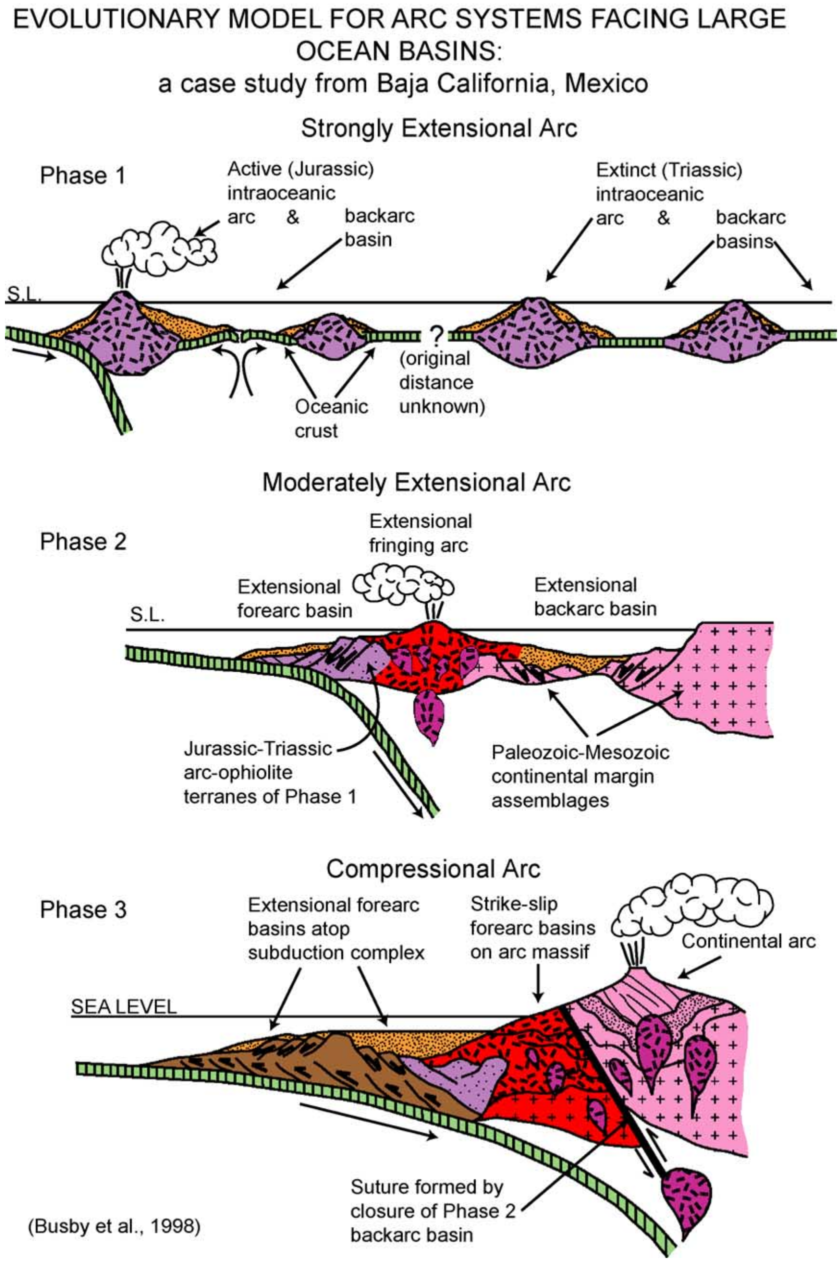


Fig. 2. Evolutionary model for arc systems facing large ocean basins, using the Mesozoic of Baja California, Mexico, as an example. Three main phases are recognized (see Fig. 1 and text).

migration of the arc axis. Subduction of old slabs causes trench rollback to be faster than trenchward migration of the upper plate, producing an extensional arc system, whereas compressional systems occur where an overiding plate advances trenchward faster than trench rollback, largely due to young age of the subducting slab (Dewey, 1980). Jarrard (1986), in an analysis of the dynamic controls on the tectonics of modern arc–trench systems, documented a strong positive correlation between the age of a subduction zone and the amount of compressional strain in the overriding plate. Dewey (1980) also predicted that convergent margins should gradually evolve from extensional to compressional. Our work in Baja California confirms this hypothesis, and provides more detailed conceptual models for

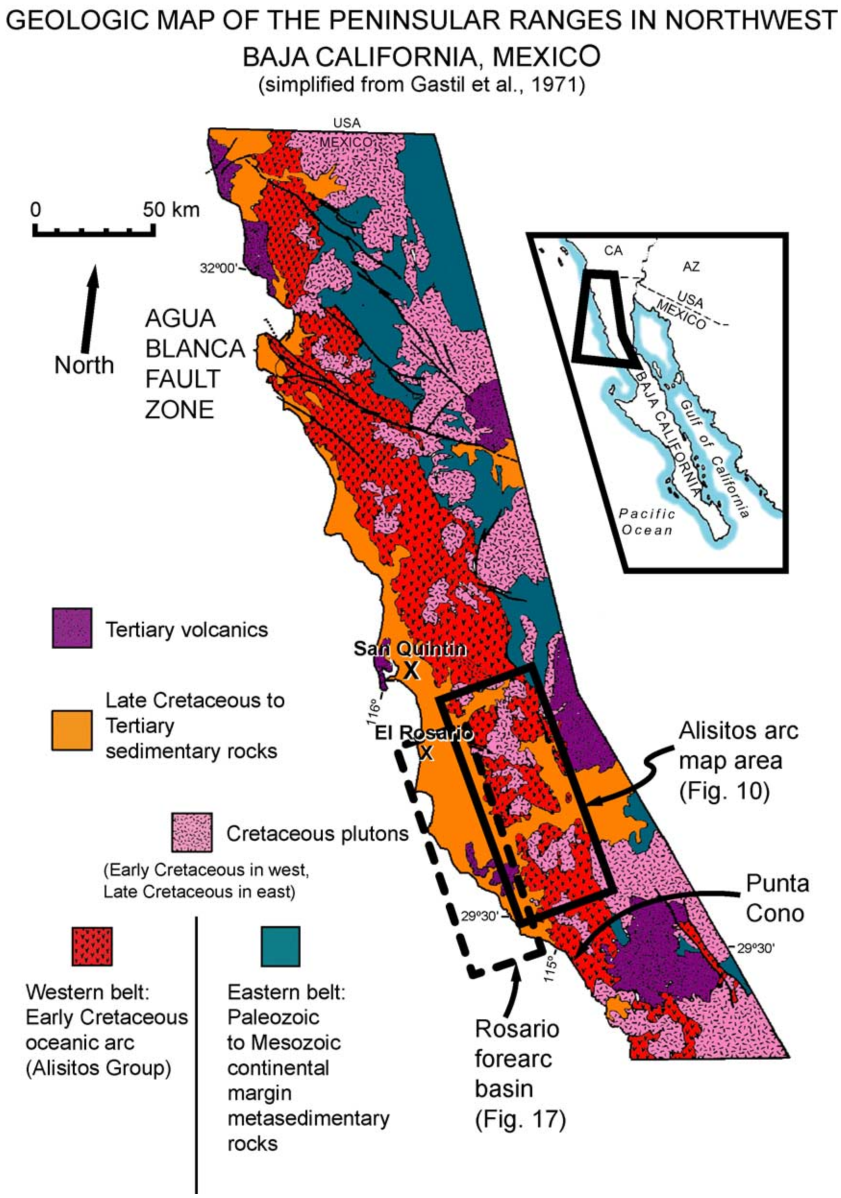


Fig. 3. Geologic map of the Peninsular ranges in northwest Baja California, showing lithostratigraphic units of the western and eastern Peninsular Ranges depicted in Fig. 1. Continental margin rocks are restricted to Paleozoic–Mesozoic metasedimentary rocks of the eastern Peninsular Ranges; the rest of the rocks on this map, and all of the rocks outboard of this (Figs. 1, 4 and 5) represent Mesozoic additions to the continental margin. A down-dip view of the Alisitos arc map area (AG of Fig. 1), interpreted as the extensional fringing arc of phase 2 (Figs. 2 and 11), is presented in Fig. 10. A geologic map and sequence stratigraphic interpretation of the Rosario forearc basin (RG of Fig. 1) is presented in Fig. 17, and interpreted as a phase 3 forearc strike slip basin in Fig. 2.

basin development along convergent margins facing large ocean basins.

Most of the subduction zones in the world have been running for a long time, so there are few examples of early, extensional arc systems relative to later, neutral or compressional arc systems (Jarrard, 1986). For this reason, we need to study the geologic record to learn more about the growth of continents along convergent margins. I present a case study of a long-lived convergent margin, synthesizing published data and interpretations (Figs. 1 and 2; Busby et al., 1998) with new data (Figs. 3, 4 and 5). I recognize three main tectonic phases in the Mesozoic evolution of Baja California. These are illustrated and described in Fig. 2, and include phase 1, highly extensional intraoceanic arc systems; phase 2, mildly extensional fringing-arc system; and phase 3, compressional continental–arc system. In my tectonic model, all of the elements that led to Mesozoic growth of the Mexican margin (except for oceanic rocks of the subduction

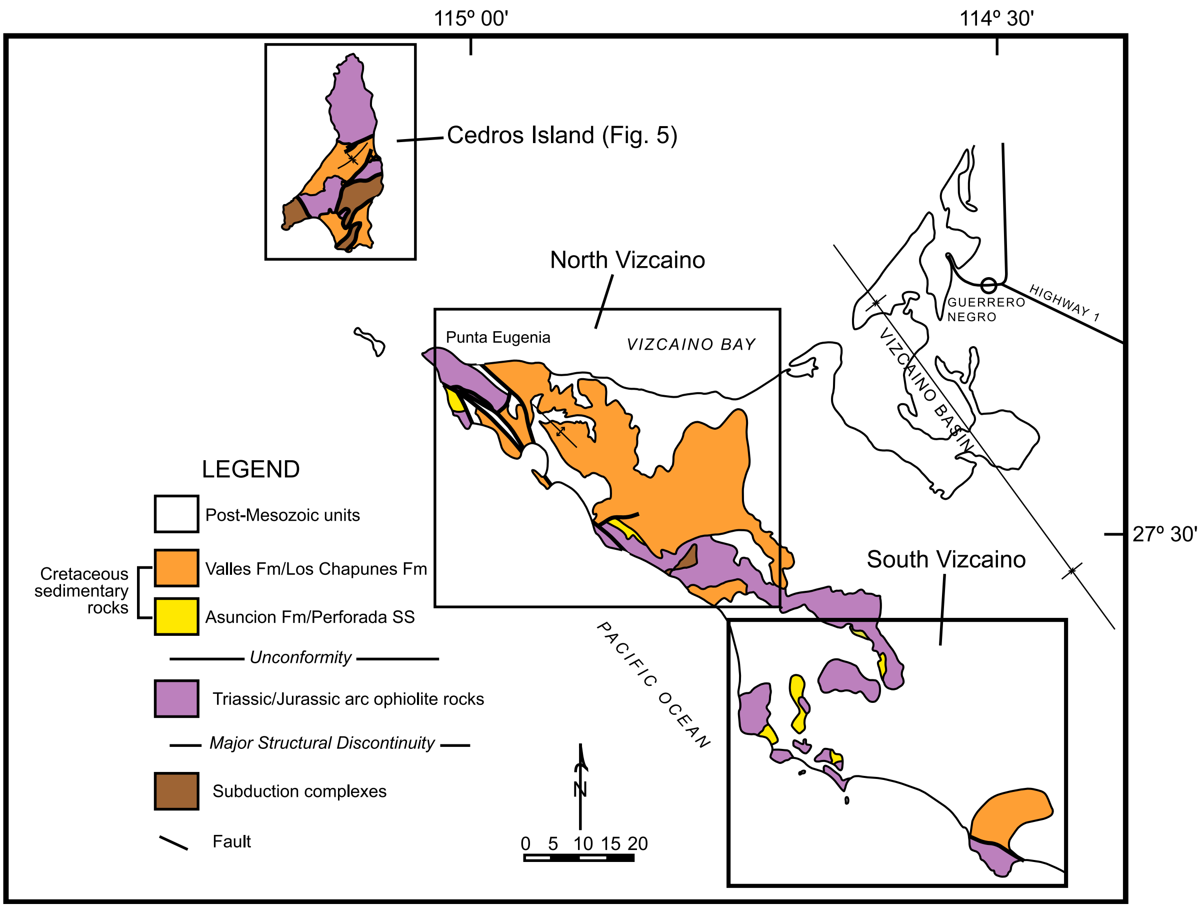


Fig. 4. Generalized geologic map of the Vizcaino Peninsula and Cedros Island, showing the distribution of Mesozoic lithostratigraphic units depicted in Fig. 1 (Cedros Island, north Vizcaino and south Vizcaino). The oldest rocks are Late Triassic to Late Jurassic arc–ophiolite assemblages of phase 1 (Fig. 2; LCO, SH, SSO, PE, CIO, C, GC, E and CEC of Fig. 1). These subterranes were amalgamated and unconformably overlapped by Early Cretaceous forearc sedimentary rocks (Asuncion Formation (A) and lower Valle Group (VG) of Fig. 1), interpreted to represent the fill of extensional forearc basins formed during phase 2 (Figs. 2, 11 and 12). These in turn are overlain by Late Cretaceous forearc sedimentary rocks of phase 3 (upper Valle Group (VG) of Fig. 1), and interpreted to represent the fill of extensional basins formed by gravitational collapse at the top of an overthickened accretionary wedge (Fig. 2), shown as subduction complexes on this map and on the map of Cedros Island (Fig. 5).

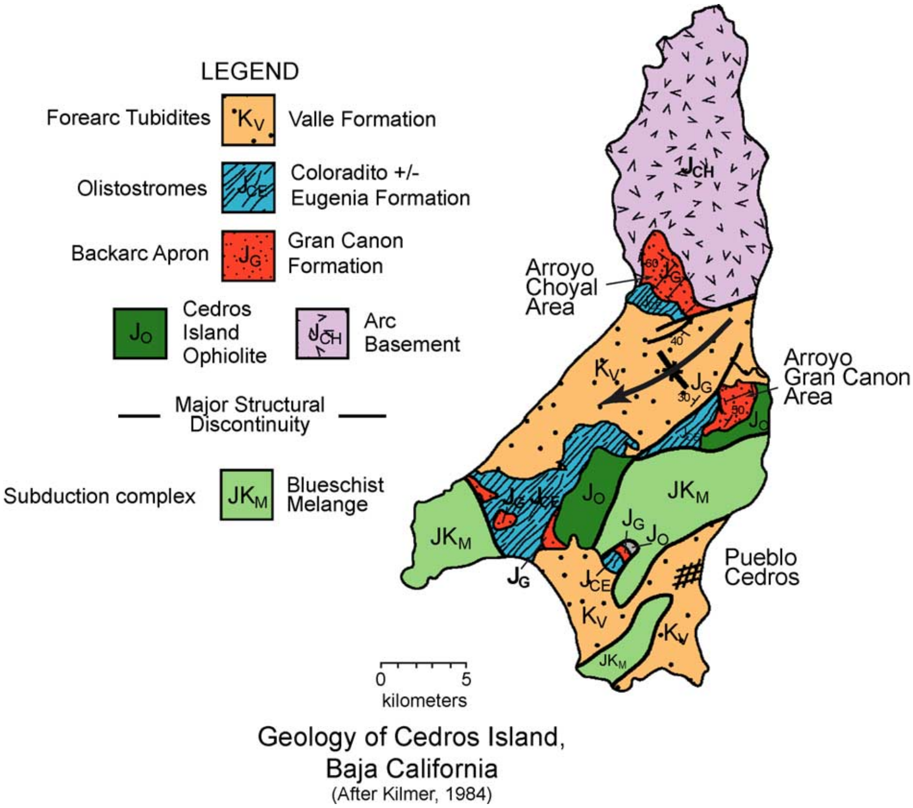


Fig. 5. Generalized geologic map of Cedros Island (for location, see Fig. 4). The oldest rocks consist of a rifted arc–ophiolite assemblage (Cedros Island ophiolite and arc basement, or CIO and C of Fig. 1). These are overlain by the Middle Jurassic Gran Canon Formation (GC of Fig. 1), interpreted as a progradational backarc apron that built southward from the rifted arc onto the spreading sea floor (Figs. 6 and 7). This in turn is overlain by sedimentary melange of the Coloradito Formation and arc volcaniclastic rocks of the Eugenia Formation (CF and EF, Fig. 1). Cretaceous forearc turbidites in the core of the Pinos syncline were deposited in normal-faulted basins during phase 3 (Figs. 2 and 15), contemporaneous with unroofing of bluechists along detachment faults at deeper structural levels (Sedlock, 1988; Smith et al., 1991).

complex) were formed and accreted in the upper plate of the margin.

2. Phase one: strongly extensional arc–ophiolite systems

The earliest stages of subduction in Baja California (ca. 220–130 Ma) are represented by intraoceanic arc–ophiolite systems of west-central Baja California (Fig. 1). Ophiolites include the Late Triassic Sierra de San Andres ophiolite in the northern Vizcaino peninsula (Moore, 1985; Sedlock, in press), the Late Triassic La Costa ophiolite in the southern Vizcaino Peninsula (Moore, 1985), and the Cedros Island ophiolite (Kimbrough, 1984). These are all interpreted to be suprasubduction zone ophiolites formed in forearc, intra-arc or backarc environments, because they have arc geochemical signatures (Moore, 1985; Kimbrough, 1984) and are directly overlain by arc volcanic–volcaniclastic rocks (Fig. 1). Early detachment faults are described from the Sierra de San Andres ophiolite (Sedlock, in press). Normal faults at the arc–ophiolite rift boundary on Cedros Island were reactivated in Late Cretaceous time, obscuring their primary features (Smith and Busby, 1993).

Volcaniclastic cover on the Sierra de San Andres ophiolite is referred to as the Puerto Escondido tuff (Barnes, 1984); similar strata resting on the La Costa ophiolite (San Hipolito Formation) contain Late Triassic fossils (Whalen and Pessagno, 1984; Barnes, 1984; Moore, 1985). Volcaniclastic cover on the

Cedros Island ophiolite, referred to as the Gran Canon Formation (Kilmer, 1977, 1984; Kimbrough, 1984, 1985; Busby-Spera, 1987, 1988a,b; Critelli et al., 2002), also overlaps rifted arc basement of the Choyal Formation, all of Late Middle Jurassic age (Fig. 1). The Gran Canon Formation is locally hydrothermally altered at its basal contact with the ophiolite indicating deposition on hot ocean floor (Busby-Spera, 1988a). I interpret the volcaniclastic cover of these ophiolites to represent arc aprons, consisting of pyroclastic and lesser epiclastic detritus deposited in small, steep-sided intra-arc and backarc basins.

The Gran Canon Formation on Cedros Island (Figs. 5–9a) offers what is arguably the least deformed, best-exposed outcrop view of a backarc apron studied to date (Marsaglia, 1995); therefore, its distinctive stratigraphy can be used to identify dismembered or poorly exposed backarc aprons elsewhere in the geologic record. The Gran Canon Formation is of further interest because it affords a more proximal

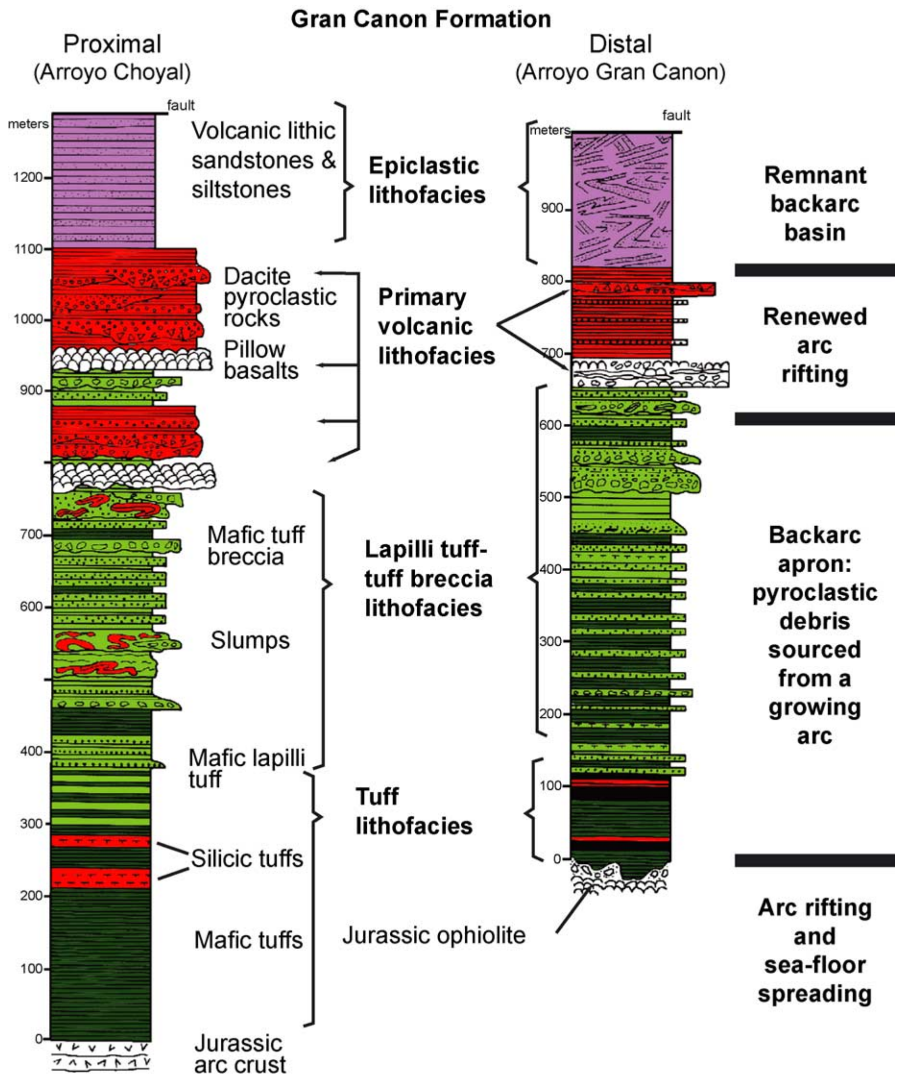


Fig. 6. Representative measured sections through proximal and distal parts of the Gran Canon Formation, modified from Busby-Spera (1988a,b) (localities in Fig. 5). Paleocurrent data indicate a northern source, and pyroclastic debris thickens from south to north in the outcrop areas across present-day Cedros Island (Fig. 5).

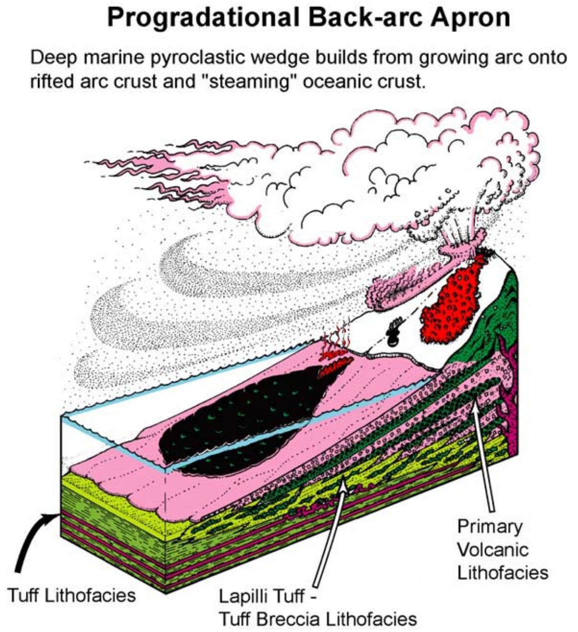


Fig. 7. Intraoceanic backarc-apron deposits typical of Phase 1, using the Gran Canon Formation on Cedros Island (Fig. 5) as an example. This drawing shows the backarc basin at the time of maximum progradation of the pyroclastic backarc apron, during renewed arc rifting (stage IV, Fig. 8).

view of a backarc apron than is available from most of the drill sites in the western Pacific, and this proximal view provides more detailed information on the volcanic and tectonic evolution of an intra-oceanic arc during backarc basin formation (Critelli et al., 2002). For these reasons, I present a summary of the stratigraphic evolution of the Gran Canon Formation, which shows a simple, uniform sedimentation pattern that may be typical of backarc basins isolated from terrigenous sediment influx (Busby-Spera, 1988a; Critelli et al., 2002; Figs. 6 and 7).

A rifted arc and ophiolte assemblage and overlying volcaniclastic rocks, all of late Middle Jurassic age, represent the arc side of a backarc basin on present-day Cedros Island (Kimbrough, 1984; Figs. 5–8). Backarc ophiolite generation was immediately followed by progradation of a deepwater pyroclastic apron into the backarc basin across rifted arc onto oceanic crust in a progressively widening backarc basin. Busby-Spera (1988a,b) divided this pyroclastic apron into *tuff, lapilli tuff–tuff breccia,* and *primary volcanic lithofacies* (Figs. 6 and 7). Pyroclastic tex-

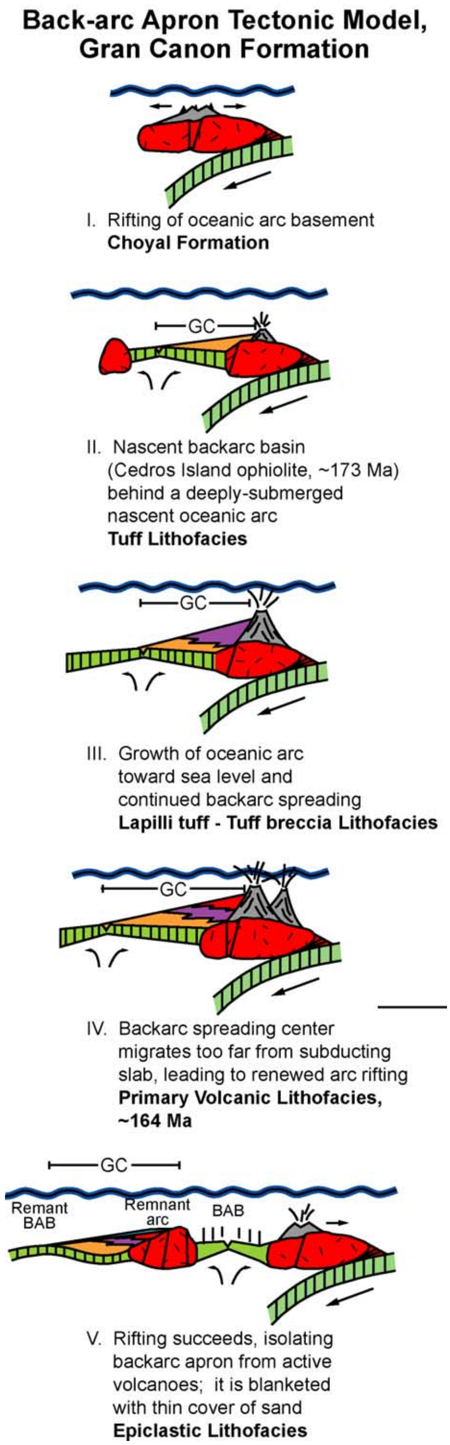


Fig. 8. Tectonic model for the Gran Canon Formation.

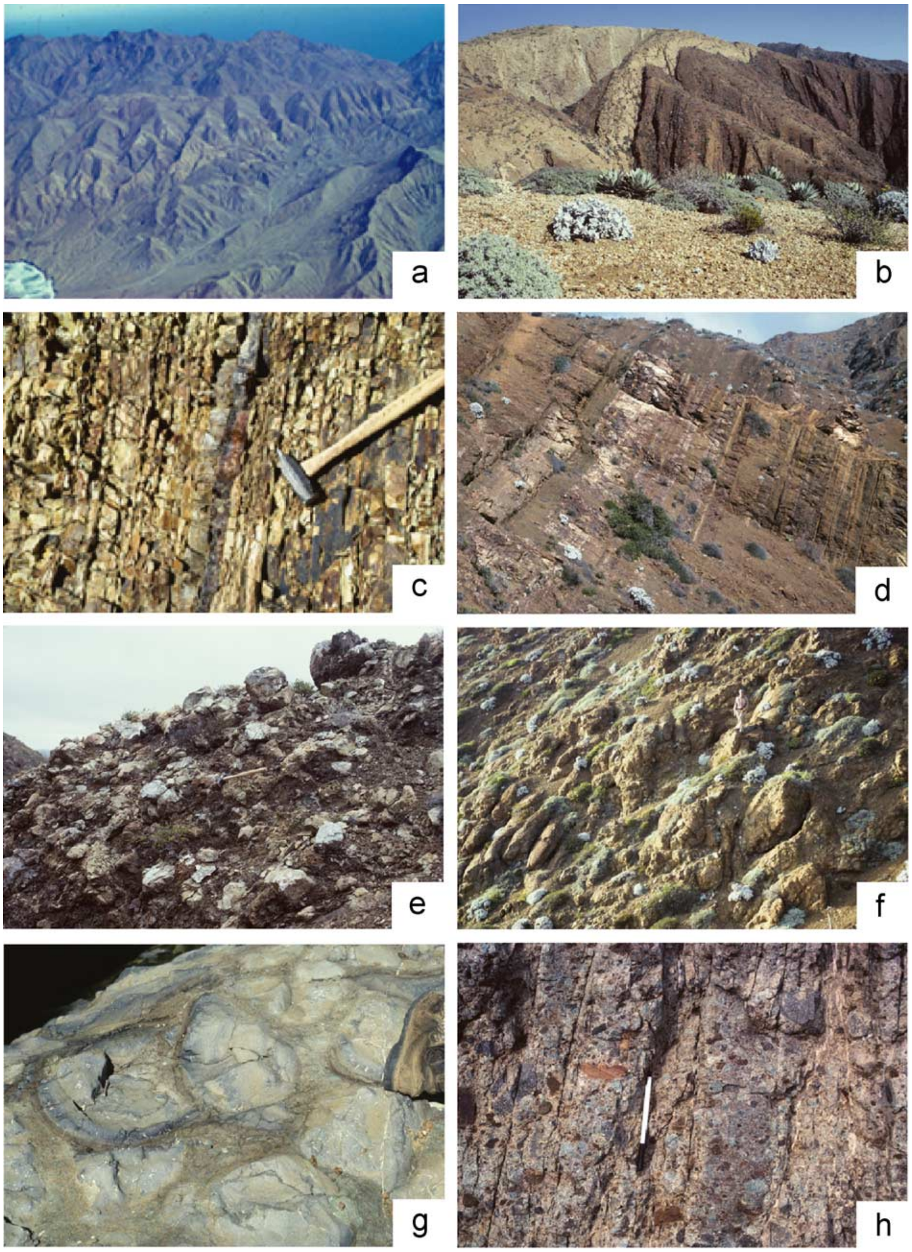


Fig. 9. Outcrop photos of the Gran Canon Formation on Cedros Island (Figs. 5–8). (a) Excellent exposure allow tracing of beds only a meter or two thick for distances of 10 km or more. (b) Early bimodal volcanism during the nascent arc–backarc phase produces mafic tuffs (brown) and lesser silicic tuffs (white) of the *tuff lithofacies* (Fig. 6). Rifted arc basement forms far ridge, visible on upper right. (c) Fine-grained, thin-bedded tuffs and minor porcellanites (*tuff lithofacies*) deposited from very dilute sediment gravity flows fed by deepwater eruptions. (d) *Lapilli tuff–tuff breccia lithofacies*, with medium- to very thick beds of lapilli tuff and tuff breccia (left) overlying *tuff lithofacies* (right); section viewed is 10 m thick. (e) Closeup of tuff breccia in proximal deepwater apron (hammer for scale), dominated by massive, unsorted, ungraded monolithic debris flow deposits. (f) Pillow lavas of *the primary volcanic lithofacies* (Fig. 6), erupted from fissures on the backarc apron produced by renewed arc rifting (author for scale). These yield excellent paleoslope data. (g) Pillow breccia of the *primary volcanic lithofacies*, with pie-shaped slices through chilled rinds to pillow interiors. Field boot for scale. (h) Dacitic subaqueous pyroclastic flow deposit of the *primary volcanic lithofacies*, with abundant pumices (brown) as well as copper sulfide altered dacite volcanic rock fragments (blue).

tures in this apron reflect the growth of the source arc terrane from a deeply submerged chain with limited production and dispersal of ash, to a shallow-marine edifice producing hyaloclastic and scoriaceous debris, to an emergent arc erupting pumiceous differentiated magmas (Figs. 6 and 9).

The *tuff lithofacies* records deposition from dilute sediment gravity flows generated by hydrostatically suppressed eruptions at a deeply submerged, nascent arc (Fig. 8, stage II). It consists of thin-bedded, well-sorted, laterally continuous tuffs that thicken into paleo-lows, indicating deposition from dilute sediment gravity flows. Its bimodal composition (Fig. 9b,c) is typical of magmatism in early rift and seafloor spreading phases, and a high ratio of unaltered green hornblende to total hornblende indicates a syneruptive, pyroclastic source (Critelli et al., 2002). It contains platy shards instead of bubble wall shards, suggestive of deepwater eruptions. The *lapilli tuff–tuff breccia lithofacies*, in contrast, shows an upsection increase in scorea fragments of explosive magmatic origin, recording a decrease in hydrostatic pressure as the summit of the volcanic source grew closer to sea level (Fig. 8, stage III). It consists of medium- to very thick-bedded lapilli tuffs and tuff breccias (Fig. 9d), deposited from debris flows on proximal parts of the apron (Fig. 9e) and from high-density turbidity currents that deposited graded megabeds on distal parts of the apron (Fig. 6). Rapid buildup of coarse detritus on proximal parts of the Gran Canon backarc apron resulted in development of numerous slumped horizons (Fig. 6). The *primary volcanic lithofacies* consists of basalt lava flows fed from fissures that extended down the backarc apron (Figs. 6, 7 and 9f), as well as monolithologic dacitic subaqueous pyroclastic flow deposits (Figs. 6 and 9h). Basalt lavas have the largest pillows (Fig. 9f) in proximal parts of the apron, and show a down-apron increase in pillow breccia (Fig. 9g), interpreted to represent flow-foot rubble that cascaded down the apron from flow fronts at upslope positions (Busby-Spera, 1987). The dacite pyroclastic flow deposits (Fig. 9h) form graded megabeds tens of meters thick (Fig. 6), inferred to be fed by caldera-forming explosive eruptions at the climax of renewed arc rifiting (Fig. 8, stage IV). Recent work in the Izu Bonin and Kurile arcs has shown that arc rifting produces silicic calderas (Gill et al., 1992; Iizasa et al., 1999; Fiske et al., 1995, 2001) where dacite pyroclastic debris dominates the stratigraphic record of the backarc basin (Nishimura et al., 1991, 1992). This second rifting event resulted in conversion of the active backarc basin into a remnant backarc basin (Fig. 8, stage V). Such an event typically occurs within 10–15 my after the birth of a backarc basin (Karig, 1983). The uppermost lithofacies, the *epiclastic lithofacies* (Fig. 6), is a fine-grained epiclastic sandstone that draped the backarc apron after rifting isolated it from the active arc (Fig. 8).

The phase 2 forearc basin forms an overlap assemblage upon amalgamated arc–ophiolite terranes of phase 1 (Figs. 1 and 2), indicating that they were amalgamated in Late Jurassic to Early Cretaceous time. The presence of continental-margin-derived slide sheets in the Late Jurassic Coloradito Formation, which overlies the Gran Canon Formation on Cedros Island (Boles and Landis, 1984), as well as granitic clasts with a strong inherited component of Precambrian lead in Early Cretaceous strata of the northern Vizcaino Peninsula (Kimbrough et al., 1987) indicates juxtaposition with a continent or continental fragment. This docking must have been gentle, however, because evidence for shortening in phase 1 terranes is generally lacking, except in the Sierra de San Andres ophiolite, and even there thrust faults and folds are not widely or penetratively developed (Sedlock, in press). Instead, distribution of key units there suggests sinistral slip in Late Jurassic or Early Cretaceous time, coeval with extension (Sedlock, in press). I suggest that terrane amalgamation was accomplished by sinistral transtension in the upper plate of the arc–trench system, with a possible local restraining bend causing minor shortening in the Sierra de San Andres ophiolite (discussed further below). A decrease in slab dip during ongoing subduction then resulted in establishment of a new fringing arc inboard of the arc–ophiolite terranes, in the present-day position of the western Peninsular Ranges (Figs. 1 and 3).

3. Phases two and three: western and eastern belts of the Peninsular Ranges batholith

The Mesozoic geology of the Peninsular Ranges spans phases two and three of this paper (Fig. 1).

We have proposed that in the second phase of subduction (Early Cretaceous time, ca. 140–100 Ma), an extensional volcanoplutonic arc developed in the present-day western Peninsular Ranges of Baja California (Alisitos Group, Fig. 1), and an extensional forearc basin complex developed outboard of it in the present-day Vizcaino Peninsula (Asuncion Formation and Lower Valle Group, Fig. 1). In our model, the Alisitos Group represents a fringing arc that lay near the edge of the continental margin, separated from it by a backarc basin that received both continental- and arc-derived sediment (Fig. 2).

In the third phase of subduction (Late Cretaceous to Early Paleocene time, ca. 100–50 Ma), a high-standing continental arc was established in the present-day eastern Peninsular Ranges (Busby et al., 1998; Figs. 1 and 2). Although the third phase was part of a gradual trend toward progressively more compressional stress, this trend accelerated in mid-Cretaceous time (ca. 105–95 Ma), perhaps due to an increase in plate convergence rate (Engebretson et al., 1985). We have proposed that this increased rate of convergence collapsed the fringing arc against the continent and caused reverse faulting and uplift (Busby et al., 1998; Figs. 1 and 2). In our model, the boundary between the western and eastern Peninsular Ranges represents the site of backarc basin closure.

In this section, I summarize previous work on the geology of the Peninsular Ranges batholith in Baja California, and discuss alternative models for the origin of the western and eastern belts, and the boundary between them. This is important to do because five different models have been published over the past 3 years, and ideas are evolving quickly (Busby et al., 1998; Johnson et al., 1999; Dickinson and Lawton, 2001; Wetmore et al., 2002; Umhoefer, in press).

3.1. Eastern and western belts of the Peninsular Ranges batholith

Three fourths of the Peninsular Ranges batholith lies south of the international border, and although it is much less well studied than rocks north of the border, existing data strongly suggest that the major lithologic belts and structural features persist for the length of the batholith (Todd et al., 1988; Lovera et al., 1997).

The 800-km-long batholith is divided axially into a western gabbro to monzogranite belt, and an eastern granodiorite–granite belt (Silver et al., 1979; Silver and Chappal, 1988; Walawander et al., 1990). Plutons of the western belt yield U–Pb zircon ages of 140 to 105 my, with no systematic geographic distribution ("static arc"), whereas plutons of the eastern belt record an eastward-migrating linear locus of magmatism from 105 to 80 my ("migrating arc" of Silver, 1986). The boundary between the western and eastern batholithic belts coincides with a step in δ^{18}O values (Taylor and Silver, 1978) and rare earth element abundances (Gromet and Silver, 1979), interpreted as the western margin of major continental crustal contribution to the batholith. This boundary also approximately coincides with a "remarkably regular contact, defined by gravity and magnetic data in southern California and northernmost Baja that dips about 45° to the east and extends to a minimum depth of 10–15 km" (Todd et al., 1988). This boundary also coincides with west-vergent high-angle thrusts that place high-grade eastern belt metamorphic rocks over lower-grade western belt metamorphic (shown on phase 3 in Fig. 2). These ductile, west-vergent high-angle thrusts have been documented in California (Cuyamaca–Laguna Mountains shear zone of Todd et al., 1988), in northern Baja California in the Sierra San Pedro Martir (Johnson et al., 1999) and as far south as the latitude of El Rosario (Fig. 3; Griffith, 1987; Goetz et al., 1988; Goetz, 1989). These workers report dates from deformed plutons that range in age from about 115 to 105 Ma, and dates on undeformed, cross-cutting plutons range that from 98 to 90 Ma.

Early publications interpreted the western Peninsular Ranges as an exotic oceanic arc terrane that was accreted to continental-margin rocks of the eastern Peninsular Ranges in mid-Cretaceous time (Gastil et al., 1978). Most workers since then have considered this boundary to be a suture between North America and a fringing arc that was originally separated from North America by a backarc basin (Fig. 3, Phase 2; Gastil et al., 1978,; Gastil and Miller, 1981; Rangin, 1978; Phillips, 1993; Saleeby and Busby-Spera, 1992; Thomson and Girty, 1994; Busby et al., 1998). The presence of the Julian Schist/Bedford Canyon Formation on either side of this boundary (interpreted as Jurassic forearc turbidites) supports the fringing arc (Fig. 2) rather than exotic arc interpretation (Saleeby and Busby-Spera, 1992). Mapping both north and south of the border has shown that the western belt Santiago Peak Volcanics overlies these turbidites in depositional contact (Kimbrough and Herzig, 1994; Sutherland et al., 2002).

An alternative interpretation of the boundary between the western and eastern Peninsular Ranges, published by Dickinson and Lawton (2001), is that: (1) this zone of deformation marks the site of closure of a large ocean basin, not a backarc basin, (2) this closure occurred before (not after) development of the Alisitos arc terrane, and (3) the Alisitos arc is a post-accretion continental margin arc. However, almost all of the evidence given for Dickinson and Lawton's hypothesis is taken from mainland Mexico where exposure is poor, rather than from Baja California, where exposure is excellent. The existing data from Baja California show that the age of the "suture" is younger (not older) than the Alisitos arc, and the geochemistry of the Alisitos arc precludes an origin as a continental margin arc. Furthermore, the "suture" lacks the accretionary wedge and forearc strata one would expect if it recorded closure of a major ocean basin (see models for remnant ocean basins published by Graham et al., 1975; Ingersoll, 1995). Last, Triassic to Jurassic continental margin arc rocks of southwestern North America (Busby-Spera, 1988a,b) and Permian to Cretaceous continental arc rocks of western South America (Burke, personal communication, 1988) show evidence for protracted extension and subsidence. Although there is no ocean floor record for most of the Mesozoic, I interpret this protracted and widespread extensional continental arc tectonic regime to record trench rollback during subduction of an old, cold and very large paleo-Pacific oceanic plate following the breakup of Pangaea. Subduction of the young, warm shrinking ocean basin shown by Dickinson and Lawton (2001) should have produced arc uplift and shortening (Jarrard, 1986).

Another alternative interpretation of the boundary between the western and eastern Peninsular Ranges, published by Johnson et al. (1999), is a two-subduction-zone model, in which the Early Cretaceous Alisitos oceanic arc originated above an east-dipping subduction zone outboard of another east-dipping subduction zone beneath a coeval hypothesized continental margin arc; in this model, subduction along the inboard continental arc drove convergence and "non-terminal suturing" of the two arcs by about 105 Ma. The SHRIMP U–Pb zircon evidence for the hypothesized coeval continental arc magmatism is, however, preliminary (Johnson et al., 1999), and the possibility remains that it is older or younger than the Alisitos arc, permitting the backarc basin closure model to remain an option (Fig. 3, Phases 2–3). We see no sedimentological evidence for uplift of the Peninsular Ranges in the forearc sedimentology before about 100 Ma (discussed below); therefore, we suggest that major uplift did not occur until the backarc basin closed completely, although the closure process may have begun along some segments of the arc by 115 Ma.

A third alternative interpretation of the boundary between the western and eastern Peninsular Ranges, published by Wetmore et al. (2002), is that western Peninsular Ranges rocks north of the present-day Agua Blanca fault zone of Fig. 3 (Santiago Peak Volcanics) represent an oceanic arc that fringed North America, whereas rocks south of the fault (Alisitos group) represent an exotic arc. Along-strike variations are common in modern volcanoplutonic arcs, however; a higher proportion of marine strata in the Alisitos arc (relative to the Santiago Peak volcanics) could reflect greater amounts of extension and subsidence in that segment of the convergent margin, and in any event, the proportion of nonmarine rocks in the Alisitos arc is substantial (Allison, 1955; Fackler Adams and Busby, 1998; Busby et al., 2003). The suggestion that the Santiago Peak Volcanics contain welded tuffs while the Alisitos ignimbrites are poorly welded (Wetmore et al., 2002) does not take into account the large volumes of welded and ultrawelded tuff mapped in the Alisitos arc (Beggs, 1984; Fackler Adams, 1997; Fackler Adams and Busby, 1998; Busby et al., 2003). The interpretation that zircon inheritance occurs in the Santiago Peak volcanics but not in the Alisitos arc (Wetmore et al., 2002) is based on an extremely limited published data set for the Alisitos arc, which in fact does show evidence of inheritence (Busby et al., 2003); similarly, the suggestion that the ages of the two arcs are different is based on a handful of published U–Pb zircon dates for the Alisitos arc. Last but not least, the exotic arc model for the Alisitos arc is that it requires removal of a significant proportion of intervening (North

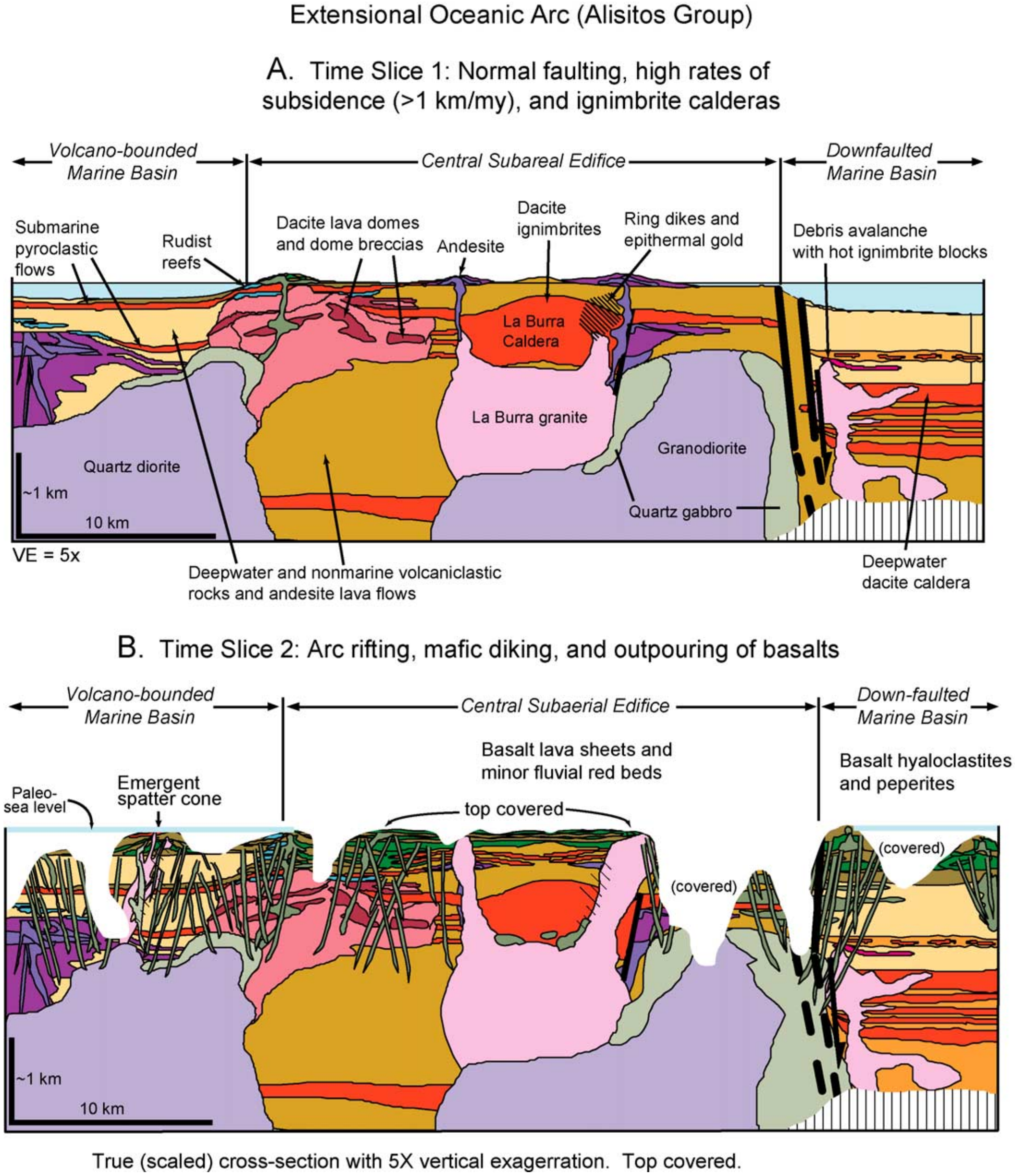


Fig. 10. Reconstruction of the phase 2 fringing extensional arc (Figs. 2 and 11) in two time slices. Time slice 1 is a reconstruction of the andesitic to dacitic arc before the onset of basalt diking and volcanism. Time slice 2 is a true scaled down-dip view of the segment of the Alisitos arc boxed in Fig. 3; the top is covered by Quaternary canyon fill as well as paleocanyon fills of the Cretaceous to Paleocene Rosario Group (Figs. 2 and 17). The entire 4-km-thick section (time slices 1 and 2) accumulated in about 1.5 Ma, at 111–110 Ma.

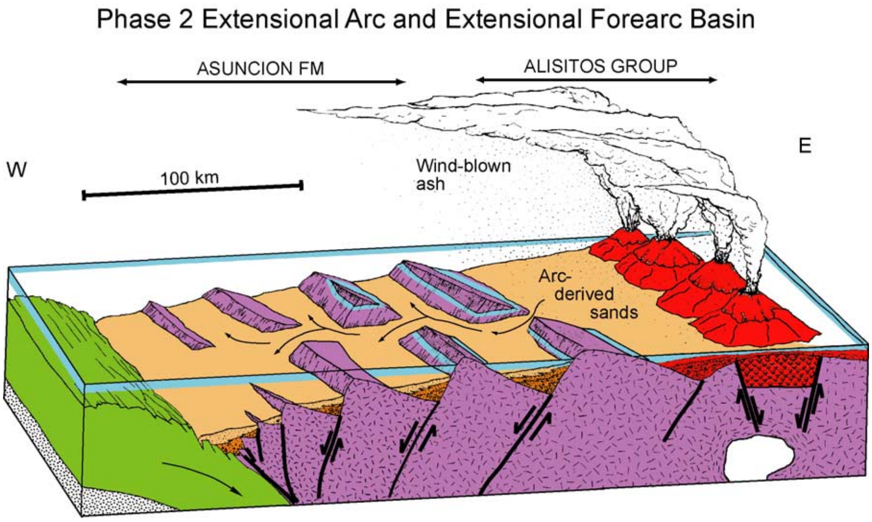


Fig. 11. Reconstruction of the arc to forearc region during Phase 2. Syndepositional normal faults drop basins to bathyal water depths both within the fringing oceanic arc (Fig. 10) and in the forearc region (Fig. 12). There is no evidence for an accretionary wedge at this time; instead, basement rocks (oceanic arc–ophiolite terranes of phase 1, Fig. 12) form horsts and grabens that downstep to the trench (Busby-Spera and Boles, 1986).

American) forearc basin and subduction complex turbidites, perhaps by subduction (Wetmore et al., 2002). I continue to prefer the simpler model that the Alisitos arc fringed North America (Fig. 2, Phase 2).

4. Phase two: extensional fringing arc and extensional forearc

Deep marine strata inboard of the Alisitos arc have interfingering arc-derived and continent-derived

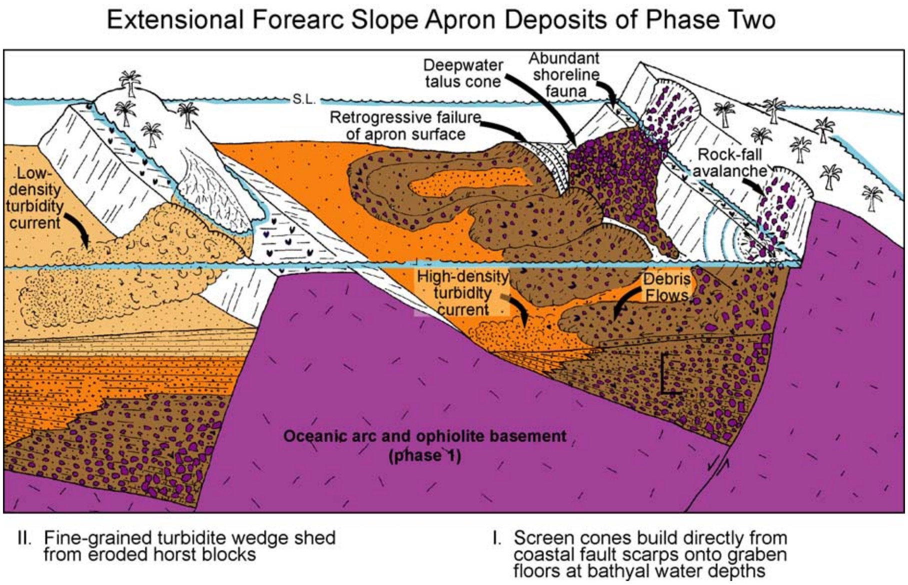


Fig. 12. Reconstruction of the phase 2 extensional forearc basin complex (Asuncion Formation of south Vizcaino Peninsula, Fig. 1).

sedimentary and volcaniclastic rocks, indicating that the Alisitos arc represents a fringing arc separated from the continent by a narrow backarc basin, (Phillips, 1993). The inboard position of the arc axis relative to phase 1, the inferred narrowness of the backarc basin, and the lack of backarc ophiolite remnants, all suggest that extension was more moderate in phase 2 (relative to phase 1), probably due to a lower angle of subduction.

In this section, we present evidence that the Baja California convergent margin was at least mildly extensional in Early Cretaceous time, with well-preserved syndepositional normal faults and high rates of subsidence in both the arc region (Fackler Adams and Busby, 1998) and the forearc region (Busby-Spera and Boles, 1986). There is no evidence for an accretionary wedge at this time (Sedlock, 1988, 1993, 1996); instead normally faulted crystalline basement extended all the way to the trench (Busby-Spera and Boles, 1986; Figs. 2 and 4). Although backarc and intra-arc extension are widely recognized processes, forearc extension is less well understood, even though it has been identified along many convergent margins, both modern and ancient (Moberly et al., 1982; Flint and Turner, 1988; Flinch and Bally, 1992; Wessel et al., 1994; Imperato, 1996). About 45% of modern convergent margins are nonaccretionary (Ingersoll and Busby, 1995), and forearc extension was probably even more common in the geologic past, when there were more young subduction zones in existence. Extension of the arc and forearc created deep basins stuffed with volcaniclastic rocks; these added to the growth of the continental margin, so distinctive features of them are described here.

4.1. Fringing island arc of phase two

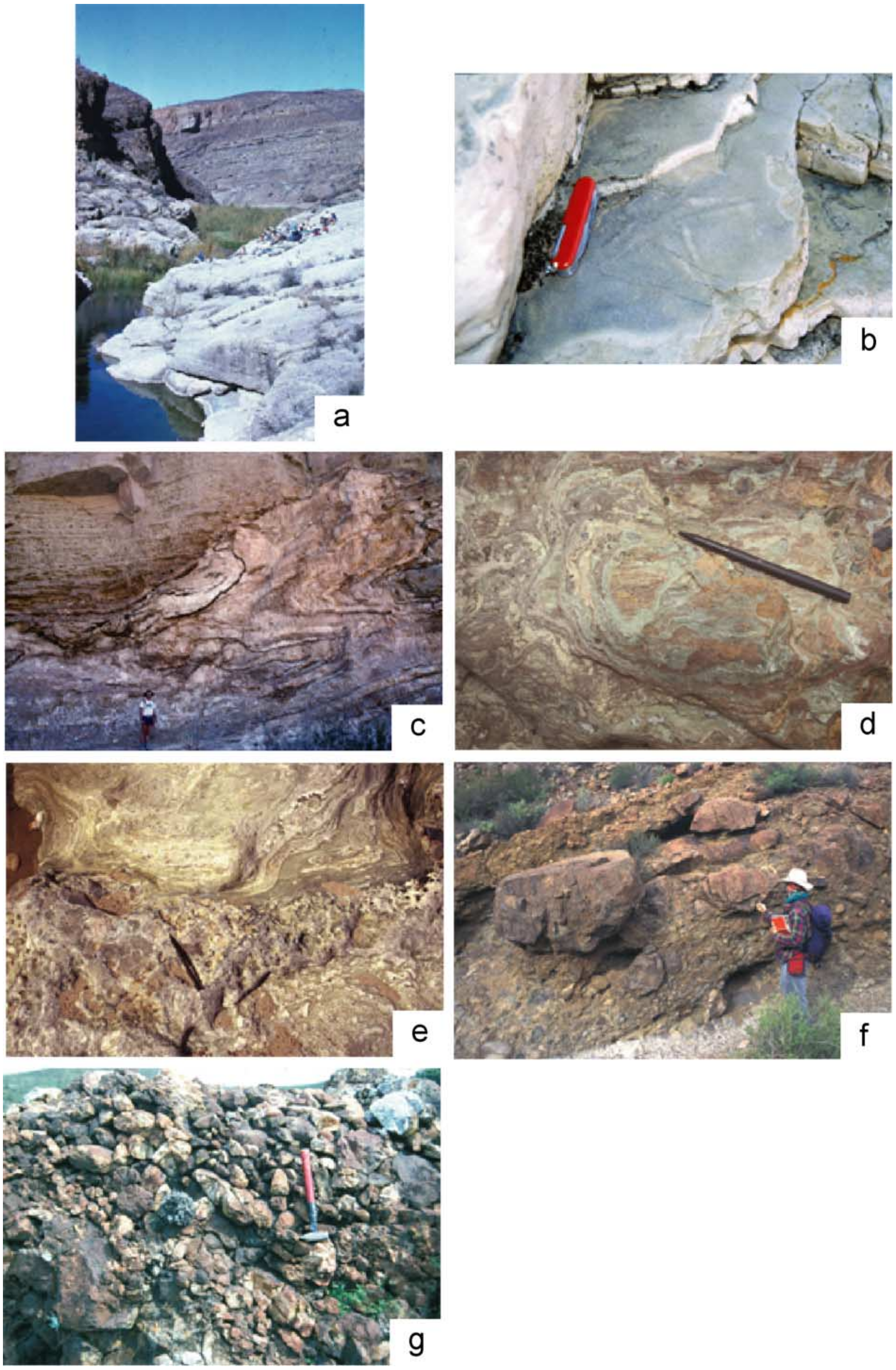
The Alisitos arc is an approximately 300 × 30-km oceanic arc terrane that lies in the western wall of the Peninsular Ranges batholith south of the modern Agua Blanca fault zone (Fig. 3). We have completed detailed mapping and dating of a 50 × 30-km Rosario segment of the Alisitos arc terrane (Fackler Adams and Busby, 1998; Figs. 2 and 10), as well as reconnaissance mapping in the 50 × 30-km San Quentin segment to the north (Busby et al., 2003). The size, unusually good exposure (Fig. 13a), and excellent preservation (Fig. 13b) of the fringing-arc terrane permits comparison of its stratigraphy and structure with those of modern fringing-arc systems.

The El Rosario segment of the Alisitos arc forms a west-dipping monoclinal section approximately 4000 m thick, intruded by contemporaneous hypabyssal and plutonic rocks (Fackler Adams and Busby, 1998; Fig. 10). The El Rosario segment of the Alisitos arc was subaerial around its main eruptive center, and was flanked by marine basins to the present-day north and south (Figs. 10 and 13a,c). The southern marine basin is a "volcano-bounded basin" (as defined by Smith and Landis, 1995), where strata accumulated in the low areas between constructive volcanic centers, in shallow water to deepwater environments (Fig. 13a,b). The northern marine basin, in contrast, is a "fault-bounded basin", which was downthrown into deep water relative to the main subaerial eruptive center along a steeply dipping fault zone (Fig. 10A).

Several characteristics of the Alisitos arc may be used to distinguish volcano-bounded- and fault-bounded-intra-arc basins in other settings. The rugged

Fig. 13. Outcrop photos of the phase 2 Alisitos fringing arc in the segment shown in Figs. 3 and 10. (a) Deep canyons cut into the western flank of the Peninsular Ranges provide outstanding exposures of the Alisitos arc terrane. In the foreground are students examining a subaqueous pyroclastic flow deposit (white, with crude stratification); on the far ridge are marine mudstones and sandstones, overlain by a resistant rudist reef (at ridge crest). (b) Metamorphic grade and deformational structures increase eastward within the Alisitos arc terrane, but most of it is well-preserved enough to display delicate primary sedimentary structures, such as this sea star resting trace. (c) Soft sediment folding produced by fluidizarion of wet substrate where hot pyroclastic flows entered the sea. (d) Lineations in the tuff of Aguajito, a dacitic welded tuff whose eruption formed the La Burra caldera (Fig. 10A). Lineations in welded tuffs form from stretching of pumices in ignimbrites deposited at very high temperatures (ultrawelded or "lava like" ignimbrites). (e) Peperite, defined as interaction of magma and wet sediment. This peperite formed where very hot ignimbrite slabs avalanched into a deep marine basin, along with wet volcaniclastic debris. The margins of the hot ignimbrite blocks interacted explosively with the wet volcaniclastic host in the avalanche deposit, causing wide zones of complex mixing to occur, in both fluidal and brittle states. (f) Subaerial dacite lava dome breccia deposits, formed of tightly packed monlithic dacite blocks in nonsorted, nonstartified deposits tens to hundreds of meters thick. (g) Relatively well-rounded, well-sorted polylithic volcanic cobble to boulder conglomerate on the subaerial edifice. Rounding is rare, due to very short transport distances and residence times in fluvial or shallow-marine environments.

down-faulted flank of the edifice produced mass wasting, plumbed large-volume eruptions to the surface, and caused pyroclastic flows to disintegrate into turbulent suspensions that mixed completely with water. In contrast, gentler slopes on the opposite flank allowed pyroclastic flows to enter the sea with integ-



rity, and supported extensive buildups of bioherms (Fackler Adams and Busby, 1998; Busby et al., in press). Rare beach conglomerates (Fig. 13g) are also restricted to the volcano-bounded basin. Slumping and other mass wasting events were rare and small in scale in the volcano-bounded basin relative to the fault-bounded basin, where topography was steeper and seismicity more common.

We recognize two evolutionary stages in the El Rosario–San Quentin (100 km long) segment of the Alisitos arc terrane: (I) extensional oceanic arc, characterized by intermediate to silicic explosive and effusive volcanism, culminating in caldera-forming silicic ignimbrite eruptions at the onset of arc rifting, and (II) rifted oceanic arc, characterized by mafic effusive and hydroclastic rocks and abundant dike swarms. New U–Pb zircon data from the volcanic and plutonic rocks of both phases indicate that the entire 4000-m-thick section accumulated in about 1.5 Ma, at 111–110 Ma (Busby et al., 2003). Two types of units are widespread enough to permit tentative stratigraphic correlation across much of this 100-km-long segment of the arc: a welded dacite ignimbrite erupted from La Burra caldera, and a deepwater debris–avalanche deposit, both shown in Fig. 10.

Fig. 10A is a reconstructed cross-section of part of the Alisitos arc, in a time frame interpreted to represent an extensional oceanic arc. As discussed above, incipient rifting of an oceanic arc and the resultant change in stress regime produce silicic calderas (Taylor et al., 1990; Gill et al., 1992; Clift, 1995; Arculus et al., 1995). Caldera collapse on the central emergent edifice ponded welded dacite ignimbrite to a thickness of at least 3 km (La Burra caldera, Fig. 10A). The subaerial outflow facies is up to 300 m thick, and densely welded even where only a few meters thick (Fig. 13d). Outflow ignimbrite traversed a dacite dome complex to the south (Fig. 13f), and entered the volcano-bounded marine basin to the south, where it is not welded, but nonetheless shows evidence of heat retention in some flow units (Fig. 13c; Fackler Adams, 1997). In the northern fault-bounded basin, the tuff of Aguajito occurs as blocks, as much as 150 m across, emplaced as part of a debris–avalanche deposit shed into deep water from the faulted basin margin (Fig. 10A). These blocks were hot enough to deform plastically, and to form peperite with the debris–avalanche matrix (Fig. 13e).

Fig. 10B is a reconstructed cross-section of part of the Alisitos arc, in a time frame interpreted to represent arc rifting. This is marked by outpouring of basalt lava flows and contemporaneous emplacement of dike swarms that both cross cut and pass upward into sills and lava flows (Fackler Adams, 1997). The sharp contact between stage 1 and stage 2 strata, and the overwhelming predominance of lavas over fluvial or marine volcaniclastic rocks, indicates that the onset of rift basalt volcanism was abrupt and voluminous. As a result, the central edifice was transformed from a subaerial stratovolcano with a summit caldera, to a basalt lava plateau fringed by small basaltic cones (Fig. 10B).

In summary, evidence for arc extension and rifting includes syndepositional normal faults, very high subsidence rates, debris avalanche deposits, and silicic calderas, culminating in widespread basalt volcanism.

4.2. Extensional forearc of phase two

Evidence for extension in the forearc region during phase two includes syndepositional normal faults, with fault scarps fringed by coarse-grained slope apron deposits (Figs. 2, Phase 2, 11 and 12; Busby-Spera and Boles, 1986). The slope apron deposits are thick, laterally extensive wedges that built outward from coastal normal fault scarps onto graben floors at bathyal water depths (Fig. 12). Time frame I, shown on the right side of Fig. 12, illustrates the peak of extension. Rock-fall avalanches produced talus aprons (Fig. 14a) that pass basinward into debris flow aprons (Fig. 14b), where retogressive failures produced high-density turbidity currents (Fig. 14c). Time frame II, shown on the left side of Fig. 12, illustrates waning extension, when the eroded horst block provided sand to a fine-grained turbidite wedge. These thin-bedded bathyal marine sandstones are dominated by Bouma C division cross lamination (Fig. 14d) and convolute lamination, suggesting deposition from low-density turbidity currents flowing down steeply sloping aprons.

The forearc grabens must have stepped downward toward the trench, because arc-derived detritus was able to make its way into each sub-basin, and paleocurrent directions are consistently toward the west (Fig. 11; Busby-Spera and Boles, 1986). This is also a characteristic of modern extensional forearcs, such as Peru (Moberly et al., 1982) the Marianas (Wessel et al., 1994) and north Chile (Buddin et al., 1993).

Phase two ended when the Alisitos arc was thrust beneath Pleozoic–Mesozoic continental margin terranes of the present day eastern Peninsular Ranges along high-angle reverse faults (see phase three, Fig. 2). As discussed above, the paleotectonic setting remains controversial, but I interpret this event to record back-arc basin closure as the angle of subduction shallowed (Fig. 2), augmented by an increase in convergence rate proposed for that time by Engebretson et al. (1985).

5. Phase three: compressional arc

In the third phase of subduction (Late Cretaceous to Early Paleocene time, ca. 100–50 Ma), a high-standing continental arc was established in the present-day eastern Peninsular Ranges (Fig. 2, Phase 3). All workers are in agreement on this, although proposed uplift mechanisms vary.

Evidence for arc uplift includes (1) the presence of ca. 105–95 Ma reverse faults within the arc (Griffith, 1987; Todd et al., 1988; Goetz et al., 1988; Goetz, 1989; George and Dokka, 1994; Thomson and Girty, 1994), and (2) the coeval sudden influx into forearc basins of coarse-grained sediment eroded from relatively deep structural levels of the arc (Barnes, 1984; Busby-Spera and Boles, 1986; Kimbrough et al., 2001). Lovera et al. (1997) used detrital K-feldspar in Peninsular Ranges forearc basins north of the border to infer very high mean denudation rates for ca. 100 Ma plutons (about 1 km/Ma). George and Dokka (1994) attributed the cooling of 94–93 Ma plutons in the northern Peninsular Ranges batholith to 10–14 km of exhumation along the Santa Rosa mylonite, in only 2 or 3 million years. They attribute this extremely high rate to east-directed extension, at rates of 5–7 km/Ma (also see Erskine and Wenk, 1985; Gastil et al., 1992). George and Dokka (1994) infer that this extension resulted from failure of an unstable crustal welt that formed at 99–94 Ma, by west-vergent thrusting and crustal thickening of 10–18 km (also see Engel and Schultejann, 1984; Todd et al., 1988; Goodwin and Renne, 1991; Grove, 1993). Alternatively, Kimbrough et al. (2001) used detrital zircon from forearc strata of the Vizcaino–Cedros region to infer that this uplift event was triggered by the emplacement of a batholith (La Posta pluton intrusions) at about 96 Ma. Emplacement of tonalite on this scale, however, requires a previously thickened crust (Kimbrough et al., 2001); therefore, I argue that this magmatism occurred in response to the same increased plate convergence rate that caused shortening, crustal thickening and tectonic uplift.

Intra-arc basins from phase three are not preserved, due to arc uplift and erosional and tectonic denudation, but forearc basins provide a record of arc unroofing. Forearc basins also provide evidence for a compressional strain regime throughout the Late Cretaceous to Paleocene, with a dextral strike slip component in the late Late Cretaceous. For these reasons, the forearc basins are described in this section.

5.1. Residual forearc basin and trench–slope basins, Vizcaino–Cedros region

Strongly coupled subduction along the Late Cretaceous compressional arc (Fig. 2C) resulted in accretion of blueschist metamorphic rocks (Sedlock, 1988), and development of a Cenomanian to Campanian residual forearc basin behind the growing accretionary wedge (Vizcaino Peninsula, Figs. 1 and 4). Collapse of the overthickened accretionary wedge resulted in development of detachment faults at depth (Sedlock, 1993, 1996); these pass upward into high-angle normal faults bounding Cenomanian to Coniacian forearc extensional basins at the surface of the wedge (Figs. 15 and 16; Smith and Busby, 1993).

A rapid transition from a mildly extensional arc to a strongly compressional arc is recorded in mid-Cretaceous strata that occur in outboard parts of the forearc basin complex (Valle Group of Vizcaino–Cedros region, Fig. 1). Gravelly sediment gravity flows with boulders up to 1 m in diameter flooded the entire region, depositing a conglomerate sheet 150–200 m thick overlain by a 2–3-km thick section of sandstone and conglomerate (Fig. 16a,b). Clast compositions show a dramatic change, from an undissected oceanic arc provenance to a deeply dissected continental arc provenance (Busby-Spera and Boles, 1986).

Conglomerates of the Valle Group on Cedros Island fill a deep-marine half-graben structure that formed by reactivation of a Jurassic fault between rifted arc and ophiolite basement of phase 1 (Smith

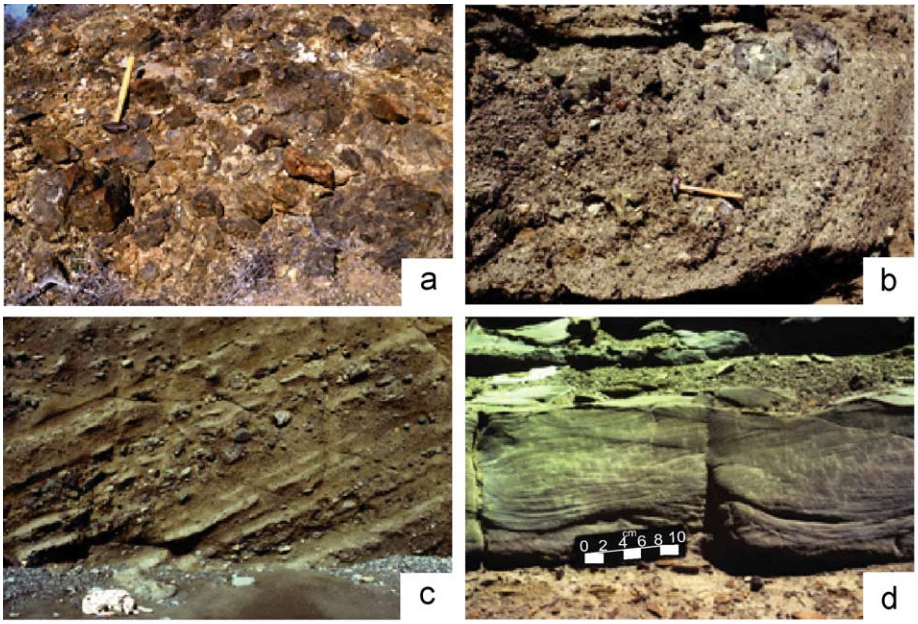


Fig. 14. Outcrop photos of slope apron deposits in the phase two Asuncion Formation extensional forearc basins shown in Figs. 11 and 12. (a) Fault talus breccia, composed of tightly packed monolithic angular clasts derived from in situ basement breccias. These were shed from subaerial to submarine fault scarps, and contain resedimented littoral fauna, as well as bathyal marine foraminifera (Busby-Spera and Boles, 1986). (b) Deepwater sandy debris flow deposit, with thin basal zone of inverse grading, massive unsorted interior, and protruding clasts at top. (c) Crudely stratified, poorly sorted breccia-sandstone beds, probably deposited from high-density turbidity currents. (d) Closeup of typical thin-bedded deepwater sandstones on the slope apron, dominated by Bouma C division cross lamination.

and Busby, 1993; Fig. 8). The axis of the half-graben acted as a submarine canyon that funneled sediment gravity flows from the arc source toward the present-day southwest, while locally derived megablocks fell from the fault scarps (Fig. 16c,d). Meanwhile, the shoulder of the half graben was draped by sandy turbidity currents that ramped up and back down the graben shoulder (Fig. 15). Propagation of a second normal fault into the area (Coloradito fault) resulted in development of a transfer zone ramp, which shed siltstone–mudstone blocks off of the horst block shown in stage I, into the enlarged basin of stage II (intraformational olistostrome, Figs. 15 and 16e). Ongoing faulting and seismicity resulted in oversteepened and unstable slopes that generated numerous submarine landslide scars (Fig. 16f), producing apron deposits of chaotic mudstone–sandstone (Fig. 15) and soft-sediment folds (Fig. 16g). As the basin filled, turbidite channels (Fig. 16h) were no longer confined to the master fault-proximal end of the basin. Note that subsidiary normal faults occur across the width of the basin, but these are not shown in the reconstruction (Fig. 15) for simplicity's sake.

The near coincidence of peak blueschist metamorphism in the lower plate of the convergent margin (Baldwin and Harrison, 1989) with the onset of normal faulting in the upper plate (Smith and Busby, 1993) supports blueschist unroofing models calling for tectonic denudation soon after peak metamorphism, due to gravitational collapse at the top of the overthickened (compressional) accretionary wedge (e.g. Platte, 1986; Jayko et al., 1987). It is important to recognize that trench–slope basins may form by upper crustal extension on top of a shortening accretionary wedge in an arc system that is regionally compressional.

5.2. Arc massif forearc strike slip basins

Arc magmatism gradually migrated eastward (inboard) within the eastern Peninsular Ranges during Late Cretaceous time (Silver, 1986), so that by Campanian time (or perhaps Turonian time), the Early

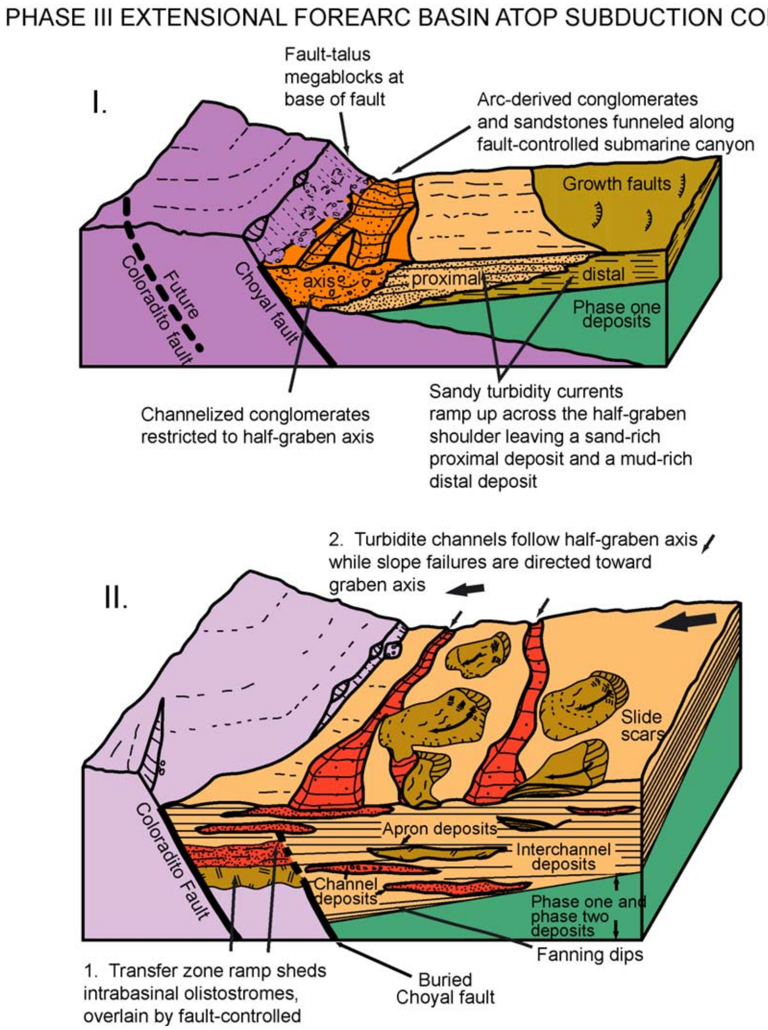
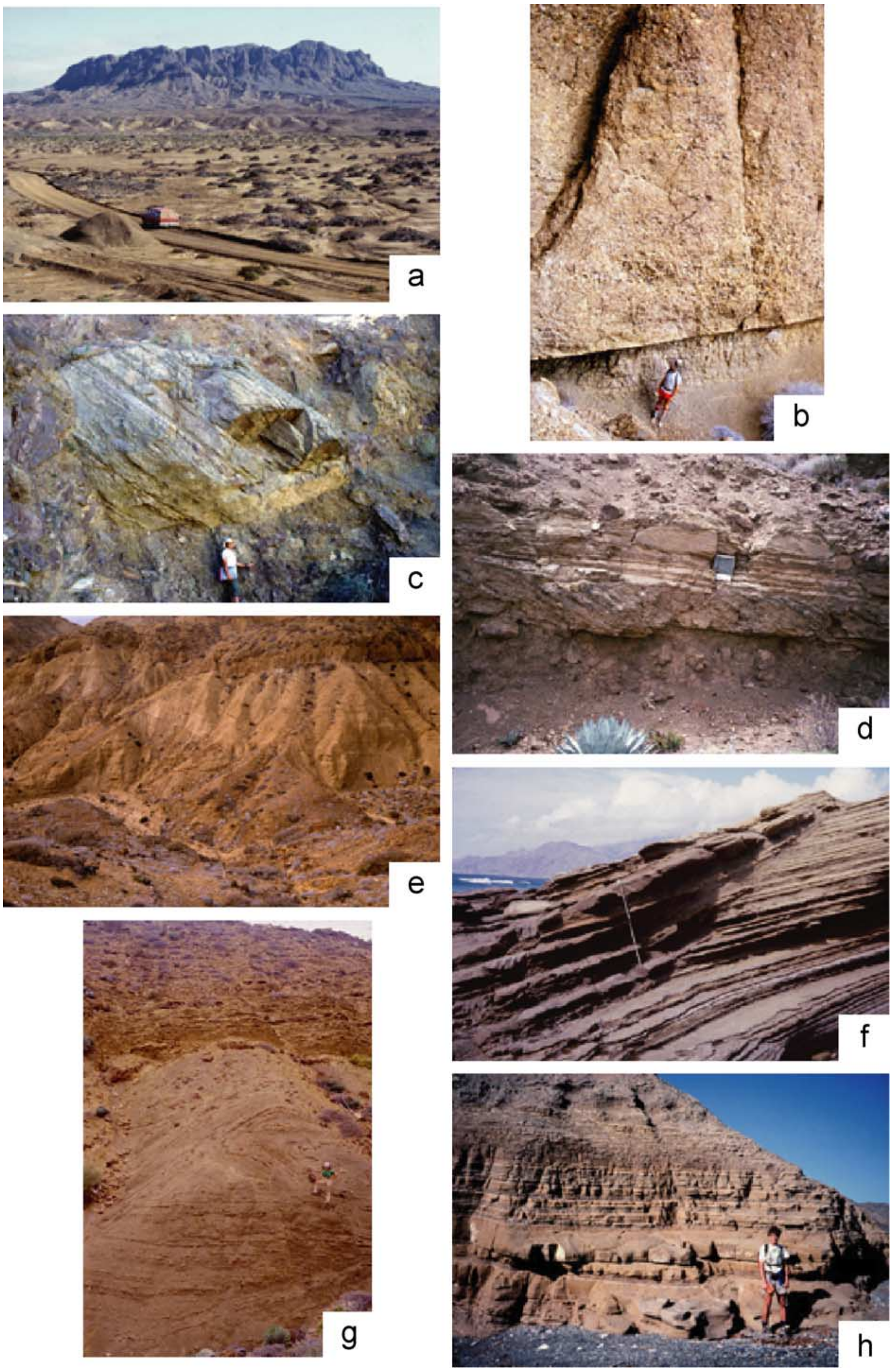


Fig. 15. Reconstruction of a deep-water extensional forearc basin that formed atop an accretionary wedge during Phase 3 (Fig. 2) on Cedros Island (Fig. 5).

Cretaceous arc basement of phase 2 became the substrate for the Peninsular Ranges forearc basin complex (Figs. 1, 2 and 3). In this section, I present stratigraphic and structural evidence that this forearc basin complex formed as dextral strike slip basins in response to oblique convergence (Busby et al., 1998) proposed for this time frame by plate reconstruction models of Engebretson et al. (1985) and Glazner (1991). Forearc strike slip basins in Baja California provide further evidence for the compressional nature of the Late Cretacous arc (phase 3, Fig. 2), since coupling is the primary factor controlling the development of strike slip faults in the upper plates of convergent margins (Jarrard, 1986).

The Rosario embayment (Figs. 1 and 17) is by far the most areally extensive, least deformed or altered, and best-exposed forearc basin segment of the Peninsular Ranges. Our studies demonstrate several lines of evidence for a strike-slip origin of the Peninsular Ranges forearc basin complex (Morris and Busby-Spera, 1988, 1990; Morris et al., 1989; Morris, 1992; Morris and Busby, 1996). Rapid vertical alternation of nonmarine and bathyal marine strata indicates alternating uplift and downdropping of the basin (“porpoising", Fig. 17). Very high rates of tectonic subsidence and sedimentation are documented for the formation we have dated best, using single crystal $^{40}Ar/^{39}Ar$ dates on tuffs (Renne et al., 1991). In our mapping of the Rosario embayment, we have recognized a distinctive structural style used to identify strike slip basins. This



style consists of faults with reverse-slip separation and normal-slip separation that develop simultaneously with grabens and arches, in positions that vary rapidly through time (Crowell, 1982; Wood et al., 1994; Nilsen and Sylvester, 1995; Barnes and Audru, 1999; Barnes et al., 2001). Rapid alternation of contractional and extensional events is a hallmark of modern strike-slip forearc basins (Kimura, 1986; Geist et al., 1988).

In fewer than 5 my (early Campanian), basal non-marine strata (Bocana Roja Formation, Fig. 17B) were folded and cut by high-angle reverse faults, dropped to bathyal depths along basin-bounding faults (Punta Baja Formation), and then uplifted above sea level and weakly folded again during renewed reverse-slip separation along intrabasinal faults (Escarpa Member of the El Gallo Formation). In late Campanian time, a series of westward-downstepping half-grabens formed along the eastern margin of the forearc basin (Fig. 17A), along faults with slickensides that show a right-slip component of displacement. The eastern basin margin fault zone was first recognized by Gastil and Allison (1966). Transverse alluvial fan–fluvial systems (Castillo and Disecado Members of the El Gallo Formation, Fig. 17) were established whose deposits indicate very rapid subsidence rates of 600 m/my (Renne et al., 1991; Fulford and Busby-Spera, 1993), consistent with deposition in a strike slip basin. Shortly before the Campanian–Maastrichtian boundary, the basin was tilted westward so that the eastern margin was uplifted to subaerial environments, passing rapidly westward (basinward) to bathyal depths (Figs. 18 and 19). This tilting resulted in valley incision along the eastern margin, and marine transgression in the western part of the basin (Rosario Formation lower sequence, Fig. 17). Finally, contraction of the forearc basin in Paleocene time resulted in development of a broad syncline along the axis of the basin, causing incision of the basin margin and resedimentation of conglomerates into the basin axis on coarse-grained subaqueous deltas (Sepultura Formation, Fig. 17).

Outstanding exposure in the Rosario forearc basin permits detailed studies of depositional systems in a very tectonically active basin (Figs. 18, 19 and 20). Alluvial fan–fluvial systems of the El Gallo Formation (Fulford and Busby-Spera, 1993) provide a sedimentologic record of intrabasinal reverse faulting (Escarpa Member) and basin-margin normal faulting (Castillo–Desecado Members, Figs. 17 and 21a,b). High subsidence and sedimentation rates led to development of very thick fluvial overbank successions with minimal soil development (Fig. 21c), promoting preservation of abundant dinosaur fossils (Morris, 1974), as well as fossilized wood fragments (Fig. 21d).

The latest Campanian to Mastrichtian Rosario Formation records bathyal marine sedimentation along the length of the Peninsular Ranges forearc basin complex, from El Rosario (Fig. 3) to southern California; George and Dokka (1994) correlated this abrupt increase in paleo-water depths with a Late Cretaceous (ca. 76 Ma), 4–5-km exhumation event inferred from fission track studies of apatites in the northern Peninsular Ranges. These were both attributed to the onset of Laramide flat slab subduction (George and Dokka, 1994), although a mechanism for increased forearc basin subsidence was not given.

The San Carlos section of the Rosario Formation (Fig. 17A) is recognized as a submarine canyon (Fig. 18) because it is an erosively based feature filled with bathyal marine deposits and bound laterally by contemporaneous bathyal marine slope deposits (Morris and Busby-Spera, 1988). Its U-shaped base is typical of the lower reaches of submarine canyons (Shepard

Fig. 16. Outcrop photos of the phase 3 extensional forearc basin atop subduction complexes on the Vizcaino Peninsula and Cedros Island (Figs. 4, 5 and 15). (a) View of Valle Group on the north Vizcaino Peninsula, showing Late Cretaceous conglomerate hundreds of meters thick (upper, bold outcrops), derived from an uplifted, dissected continental arc (phase 3, Fig. 2). (b) Closeup of forearc strata shown in Fig. 16a, showing sedimentary record of abrupt transition from undissected oceanic arc source (phase 2, Fig. 2) to uplifted, dissected continental arc source (phase 3, Fig. 2). (c) Avalanche megablock of Jurassic Gran Canon Formation derived from Choyal fault scarp, encased in Cretaceous arc-derived conglomerates funneled along the axis of a fault-controlled submarine canyon (Fig. 15). (d) Channelized pebbly sandstone in submarine canyon axis (Fig. 15), with medium-scale bedform that can be traced 30 m in the down-paleocurrent direction, indicating deposition from a sustained, steady turbidity current. (e) Intraformational olistostrome formed by tilting on transfer zone ramp between the overlapping tips of two normal faults (Choyal and Coloradito faults, Fig. 15); 40-m thickness of it shown here, capped by canyon axis turbidite conglomerates (resistant beds at top of photo). (f) Slide scar cut onto turbidites and onlapped by turbidites; slide scars lie on the shoulder of the half graben and face toward its axis (Fig. 15). (g) Soft sediment folds in thin-bedded turbidites on the half graben shoulder; this verge toward the half-graben axis. (h) Fining- and thinning-upward sequence of sandstone beds filling a turbidite channel.

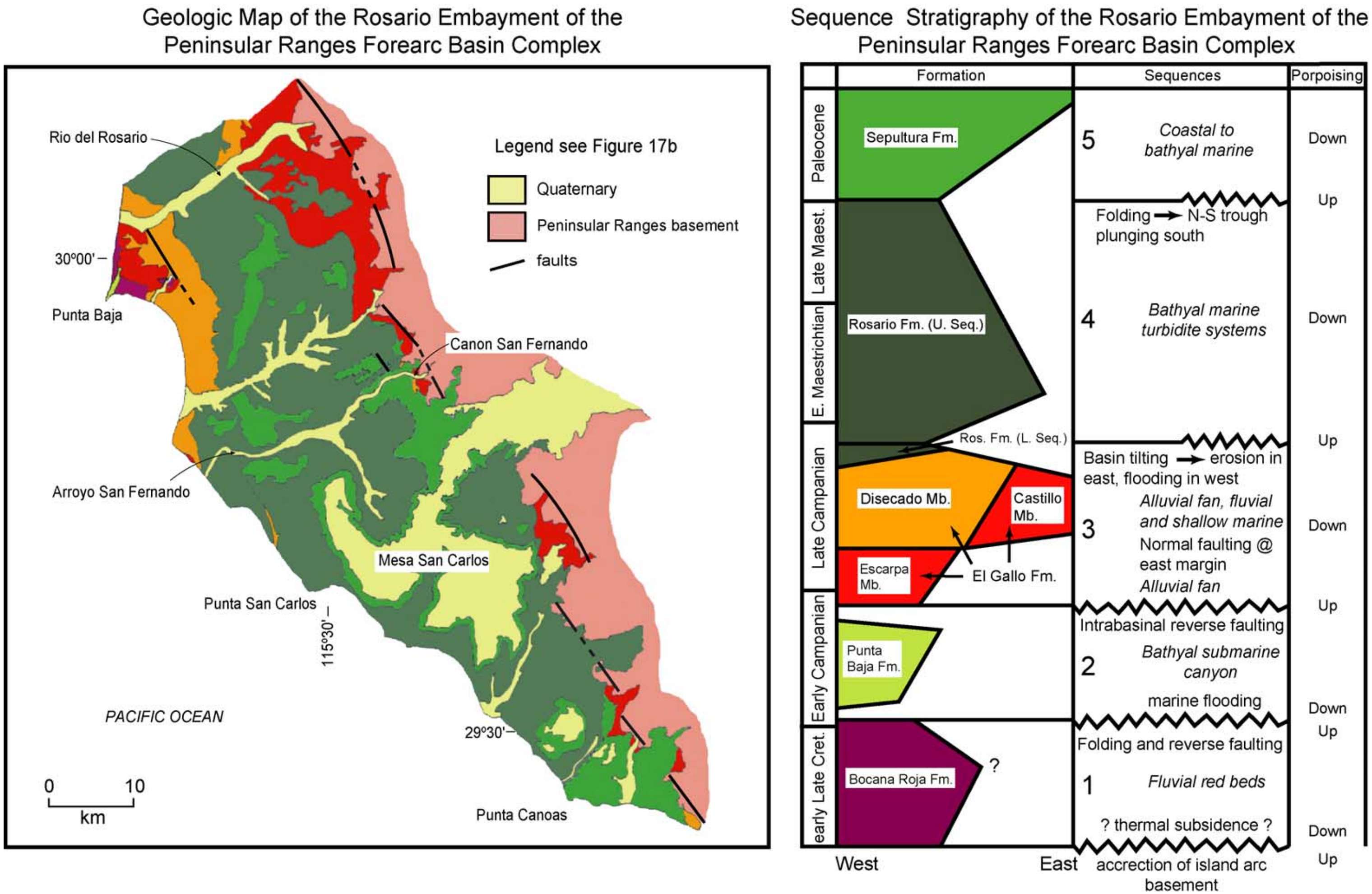


Fig. 17. Geologic map and tectonstratigraphic chart of strike-slip forearc basin built upon arc massif basement during phase 3 (Fig. 2): the Rosario forearc basin (locality shown in Fig. 3). Rapidly alternating contractional and extensional events are shown; these resulted in "porpoising" of the basin between nonmarine and bathyal marine environments. Formations defined by Kilmer (1963); lower and upper sequence of Rosario Formation defined by Morris and Busby (1996).

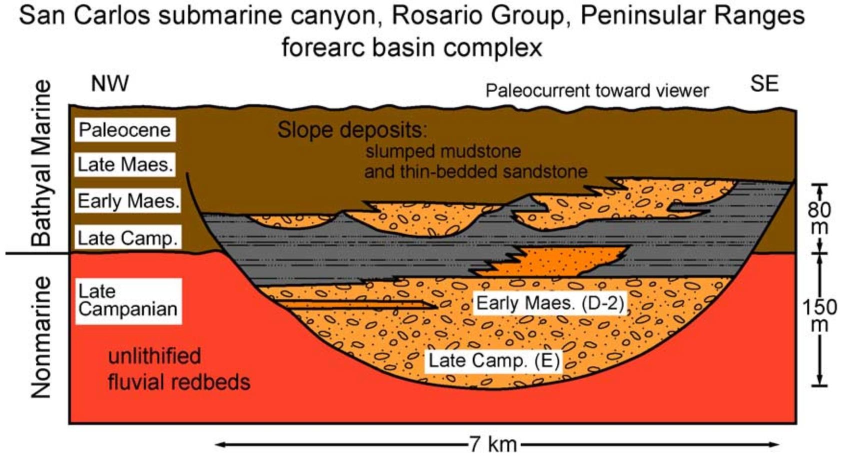


Fig. 18. Reconstructed cross-section of a bathyal submarine canyon of the Rosario Formation in the Punta San Carlos area (Fig. 17). Note vertical exaggeration.

and Dill, 1966), consistent with the presence of lower bathyal marine foraminifera. This indicates that the basin deepened very rapidly from east to west. The lower conglomerate sandstone unit (Fig. 18) consists of amalgamated channels filled with conglomerate or sandstone (Fig. 21e), interstratified with intraformational slide blocks up to 100 m long (Fig. 21f), with broken margins containing abundant injection structures (Fig. 21g). Load structures are common, and in places these detached completely from the gravelly sediment gravity flow into liquefied sands to form giant psuedonodules (Fig. 21h), suggestive of high sedimentation rates. The middle mudstone–sandstone unit (Fig. 18) consists of slumped mudstones, indicating high axial gradients on the canyon floor, as well as a single aggradational turbiditic sandstone channel with abundant traction structures. The upper conglomerate-sandstone unit is similar to the lower one except that it is smaller in volume (Fig. 18). The coarse grain size, large slide sheets, liquifaction structures and

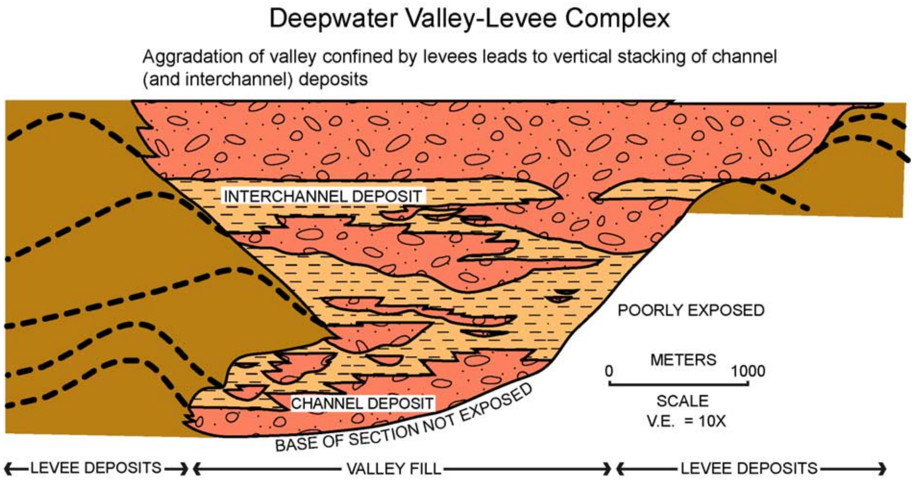


Fig. 19. Reconstruction of a deepwater valley levee complex of the Rosario Formation in the Arroyo San Fernando area (Fig. 17). Note vertical exaggeration.

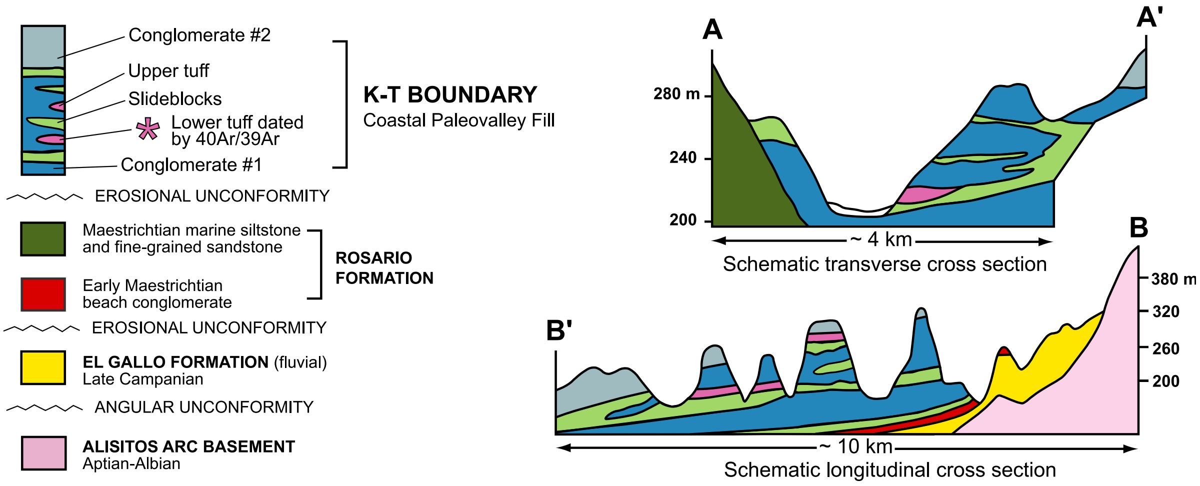


Fig. 20. Transverse and longitudinal cross-sections through Cretaceous–Tertiary boundary coastal paleovalley in Canon San Fernando (Fig. 17), modified from Busby et al. (2002). The 5-km-wide and >15-km-long valley was carved by catastrophic landsliding of Rosario Formation (dark green), and filled with Rosario-derived giant slide sheets (pale green) interstratified with massive shallow-marine conglomerates (dark blue). Pumice lapilli tuffs (pink) provide age controls on the section (Fig. 21l).

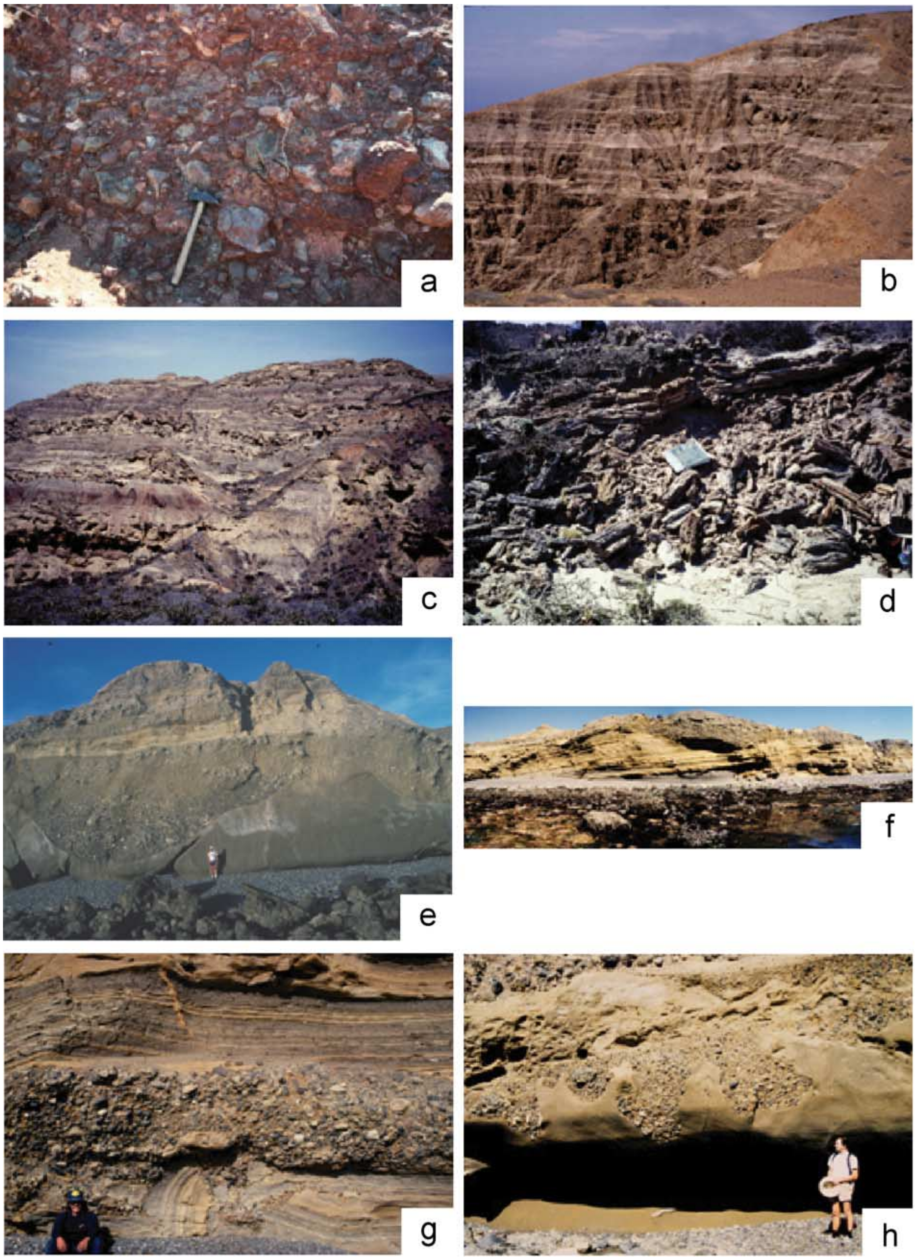
inferred deep basin, steep axial canyon gradient and high sedimentation rates are all consistent with deposition in a very tectonically active basin.

The Arroyo San Fernando section of the Rosario Formation (Fig. 17) is interpreted to represent a deepwater valley–levee complex (Fig. 19) because it consists of stacked, bathyal marine channel–interchannel turbidites (Fig. 21i) that aggraded vertically due to confinement by bathyal marine levee deposits (Fig. 21j; Morris and Busby-Spera, 1990). Comparisons with seismic data on young subsurface systems suggest that the Arroyo San Fernando levees may be more intensely slumped than many others, perhaps due to frequent seismic activity in the forearc strike slip basin (Dykstra and Kneller, 2002).

The Punta Canoas section of the Sepulture Formation (Fig. 17 and 21k) was dominated by marine conglomerates that may be the sedimentary record of a late Early Paleocene (ca. 62 Ma) uplift event inferred from fission track thermochronology of the Peninsular Ranges in California (Dokka, 1984). An earlier influx of conglomerates at the Cretaceous–Tertiary boundary, however, is hypothesized to record resedimentation events triggered by seismicity caused by the Chicxulub bolide impact (Busby et al.,

Fig. 21. Outcrop photos of the Rosario forearc strike slip forearc basin of phase 3 (Figs. 2, 3 and 17). (a) Fault talus breccia of the Castillo member of the El Gallo Formation (Fig. 17), shed from basin-margin normal fault cut through Alisitos arc basement during a period of basin downdropping. (b) Alluvial fan conglomerate and sandstone of the Escarpa member of the El Gallo Formation (Fig. 17), shed from intrabasinal reverse fault produced during a period of basin uplift. Height of cliff face is about 150 m. (c) Overbank and lesser channelized fluvial sandstones and siltstones of the Desicado member of the El Gallo Formation, deposited during a period of basin extension and subsidence (Fig. 17). (d) "Logjam" in a fluvial channel of the El Gallo Formation (map case for scale). (e) Conglomerate-filled scours in the lower conglomerate-sandstone unit of the San Carlos submarine canyon (Fig. 18). (f) 100 m long intraformational slide sheet within in the lower conglomerate-sandstone unit of the San Carlos submarine canyon. (g) Margin of the intraformational slide sheet shown in F, with arrested development of injection structures (clastic dikes of sandstone and conglomerate), caught in the act of breaking off the margins of the slide sheet. (h) Large load structures in the lower conglomerate-sandstone unit of the San Carlos submarine canyon (Fig. 18). (i) Valley fill deposits of the Arroyo San Fernando deepwater valley–levee complex (Fig. 19), showing interstratified channelized conglomerate-sandstone units and laterally extensive interchannel sandstone-mudstone units. (j) Slumped levee deposits of the Arroyo San Fernando valley–levee complex. (k) Late Early Paleocene coarse-grained delta system at Punta Canoas (Fig. 17). (l) Pumice lapilli tuff interstratified with the slide sheets and conglomerate-sandstone units of the Canon San Fernando coastal paleovalley; dates on this date show that the mass wasting deposits formed at the Cretaceous–Tertiary boundary. (m) Mass wasting deposits in a coastal paleovalley formed and filled at the Cretaceous–Tertiary boundary (Fig. 20).

2002). These Cretaceous–Tertiary boundary conglomerates occur in a coastal paleovalley (Fig. 20) preserved in present-day Canon San Fernando (Fig. 17). This coastal paleovalley (Fig. 20) formed by massive gravitational collapses, and rapidly filled with coastal (shallow marine and lesser fluvial) gravels and sands, as well as slide sheets of marine mudstone that range from meters to kilometers in length (Fig. 21m). $^{40}Ar/^{39}Ar$ dates on pumice lapilli tuffs from the base and the top of the section (Fig. 21l) are indistinguishable in age from Haitian tektites dated by the same lab, and show that the section accumulated very rapidly (Busby et al., 2002, and unpublished data).



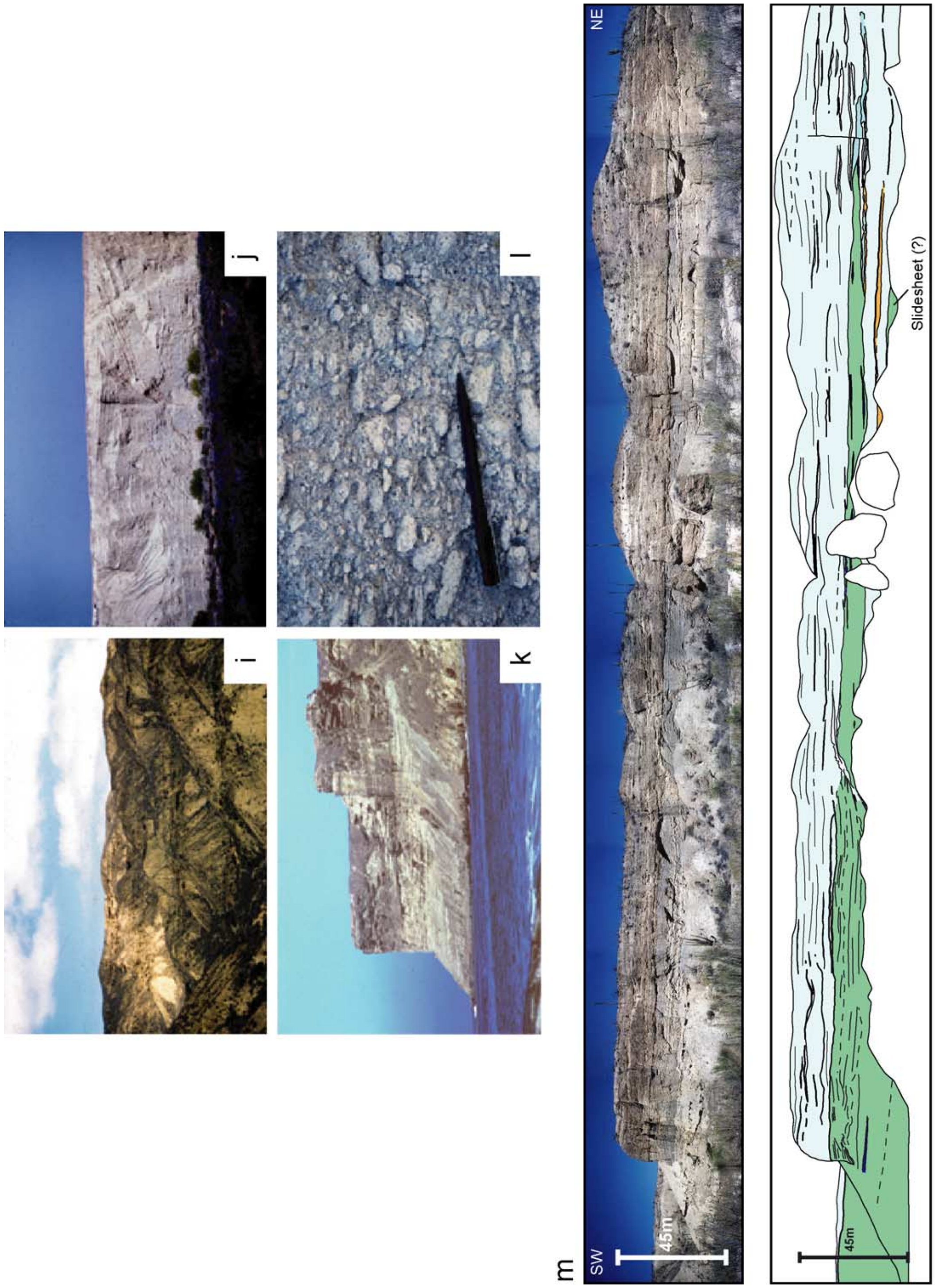


Fig. 21 (*continued*).

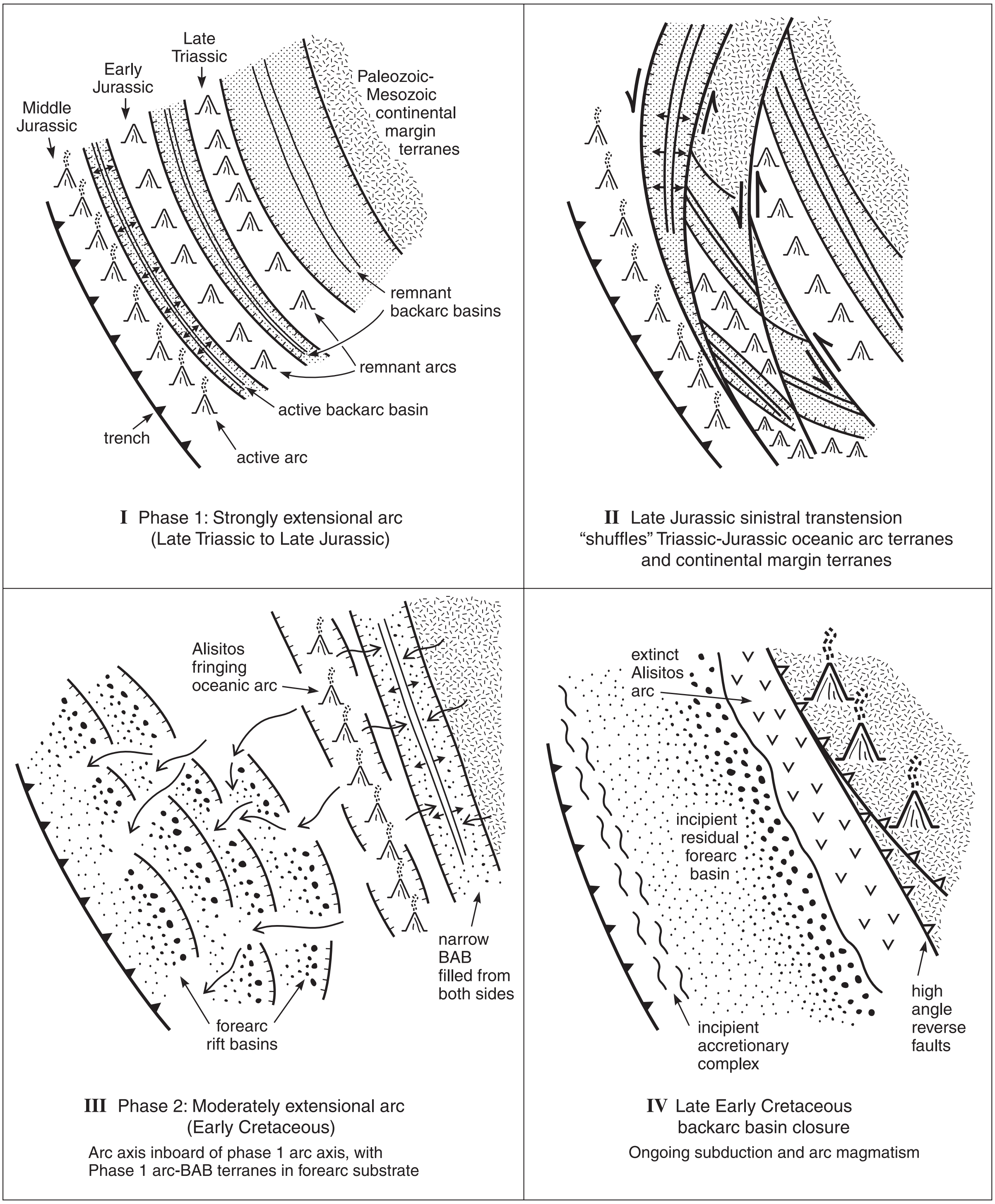


Fig. 22. Schematic tectonic reconstruction of the Baja California margin through Mesozoic time.

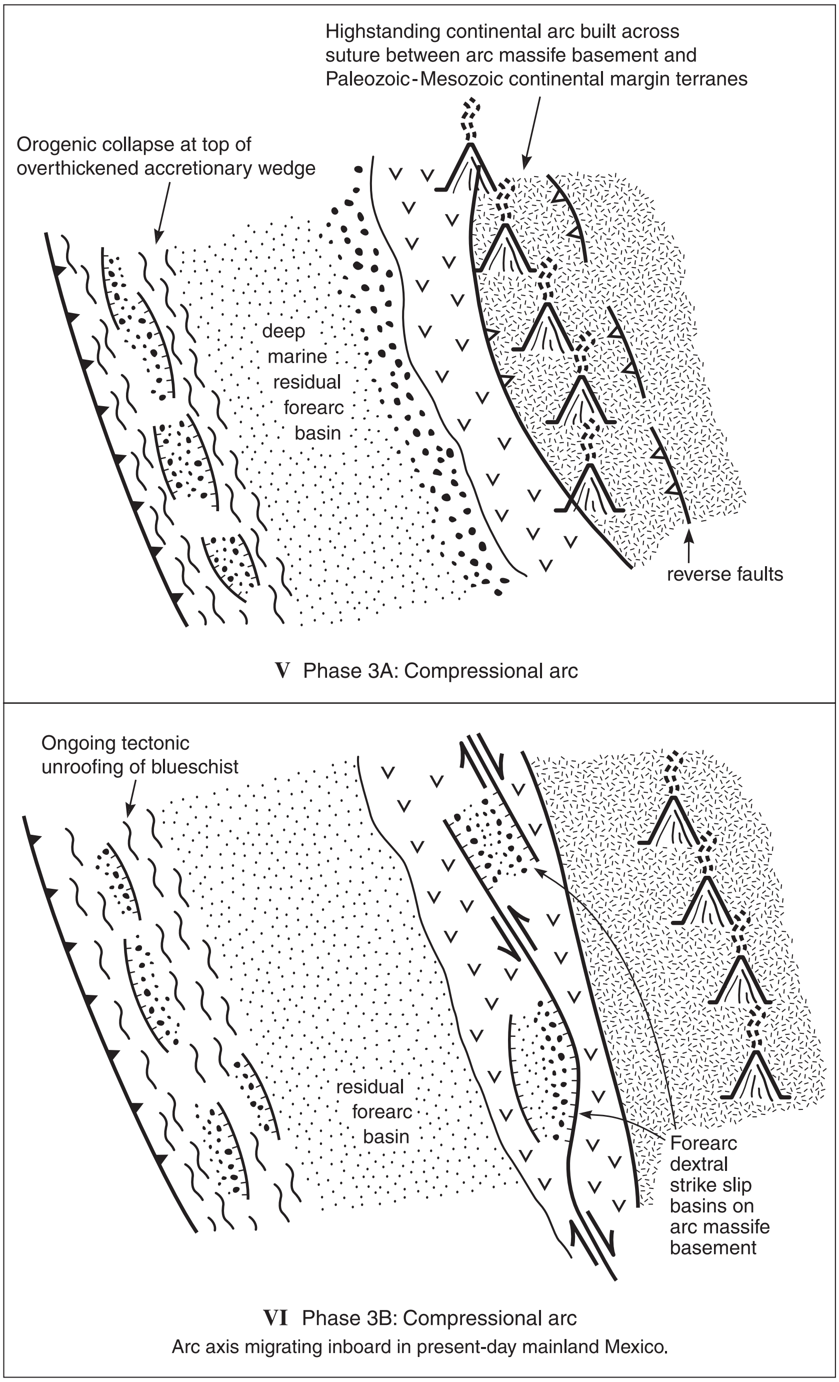


Fig. 22 (*continued*).

6. Plate margin-scale discussion and conclusions

Mesozoic to Paleocene rocks of the Baja California Peninsula provide an opportunity to study a very well-exposed, areally extensive, long-lived convergent-margin complex. Like modern convergent margins, it shows an overall evolution from extensional to compressional strain regimes, divided into three phases in this paper (Fig. 22).

Extension characterized the early history of not only the Mesozoic arc of Baja California, but also the Mesozoic arc of the southwestern Cordilleran United States and South America. This superregional tectonic regime may have been controlled largely by slab age (Busby-Spera et al., 1990a) because the paleo-Pacific Ocean basin at the time of breakup of Pangea was probably composed of large, relatively old, cold plates. Early Mesozoic extension of continental crust in the southwestern United States created intra-arc and forearc graben-depressions filled with kilometers of volcanic and sedimentary rock (Busby-Spera, 1988a,b; Saleeby and Busby-Spera, 1992); these contributed significantly to the growth of the continental margin. Similarly, Mesozoic extension of transitional to oceanic crust in outboard parts of the convergent margin created arc-related basins filled with volcanic and sedimentary rock (Saleeby and Busby-Spera, 1992; Busby et al., 1998); these nonsubductible items were accreted to present-day California and Baja California during the later, compressional stage of convergence, causing growth of the continent.

Phase one in Baja California, strongly extensional arc–ophiolite systems of Late Triassic to Late Jurassic age (Fig. 22, time frame I), has analogues in the fringing arc–ophiolite terranes of the Klamath–Sierran and Coast Ranges of California (Saleeby and Busby-Spera, 1992), although some workers argue that these terranes are exotic to both of the Californias (Dickinson and Lawton, 2001). These terranes consist of rifted oceanic arcs and suprasubduction zone ophiolites directly overlain by arc volcanic–volcaniclastic rocks.

The Late Triassic to Late Jurassic arc–ophiolite systems of Baja California appear to differ from those of northern California by lacking evidence for Middle and Late Jurassic shortening events (e.g. thrust faults, folds and cleavage). These shortening events in northern California have been attributed to accretion of exotic or fringing arcs along subduction zones of controversial numbers and polarities (Schweickert and Cowan, 1975; Dickinson et al., 1996; Dilek and Moores, 1993, 1995; Godfrey and Dilek, 2000). Alternatively, the Middle and Late Jurassic shortening events in northern California have been interpreted to record transpressional deformation along a single subduction zone that dipped eastward under the continent (Saleeby and Busby-Spera, 1992). In this alternative model, a broad right bend in the sinistral oblique subduction margin caused shortening along the northern California segment of the convergent margin, while at the same time, to the south in southern California and southern Arizona, a broad left bend in the sinistral oblique subduction margin caused transtension (Busby-Spera et al., 1990b; Adams et al., 1997; Busby et al., in press). I propose here that this southern transtensional belt extended farther south into the Mexican margin, resulted in amalgamation of oceanic arc terranes and continental margin terranes without causing significant shortening (Fig. 22, time frame II).

Phase two in Baja California, moderately extensional fringing arc–forearc systems of Early Cretaceous age (Fig. 22, time frame III), has an analogue in the fringing arc of the northern Peninsular ranges in California (Santiago Peak Volcanics), although some workers argue that the arc in Baja California (Alisitos Group) is an exotic rather than fringing arc (Wetmore et al., 2002). This fringing arc does not have an analogue in northern California, where shortening along the Middle to Late Jurassic convergent margin resulted in collapse of fringing arcs against the continental margin to form a thickened crustal basement for the Early Cretaceous arc. Instead, the Mexican convergent margin was characterized by ongoing extension, similar to the Cretaceous convergent margin of South America. This extension ended in late Early Cretaceous time, when decreasing slab dip and increasing convergence resulted in backarc basin closure and development of a compressional continental arc (Fig. 22, time frame IV).

Phase three in Baja California, compressional arc system of Late Cretaceous age, (Fig. 22, time frame V), consists of a high-standing continental arc, with accretion of blueschist metamorphic rocks and development of extensional forearc basins formed by gravitational collapse of the accretionary wedge.

These all have direct analogues in the Klamath–Sierra arc and the Franciscan complex of California. Later in phase three, oblique strongly coupled subduction resulted in development of Late Cretaceous to Paleocene forearc strike slip basins in the present day Peninsular Ranges (Fig. 22, time frame VI). Late Cretaceous forearc basins of strike slip origin have not been recognized in California, but intra-arc strike slip faults of that age are present in the Sierra Nevada batholith (e.g. Busby-Spera and Saleeby, 1990; Glazner, 1991).

All three phases of convergence contributed substantially to the growth of the North American continental margin.

The schematic tectonic model presented in Fig. 22 takes the view that all of the Mesozoic elements that added to the growth of the Mexican margin were formed and accreted within the upper plate of the convergent-margin onshore and offshore of North America (except for oceanic materials offscraped into the accretionary wedge of phase three). This is in contrast with the other end member model, which proposes that many of the elements are exotic to North America, and were accreted by closure of major (not marginal) ocean basins (Dickinson and Lawton, 2001; Umhoefer, in press). My more "fixist" model is consistent with the geologic data in hand, and must be considered a viable option. It may explain similarities in Mesozoic tectonic styles between Baja California and adjacent parts of North America and South America.

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References

Abbott, P.L., Hanson, A.D., Thomson, C.N., Bradshaw, K.D., Pollard, W.J., Seeliger, T.E., 1993. Geology of the Paleocene Sepultura Formation, Mesa de la Sepultura, Baja California, Mexico. Ciencias Marinas 19 (1), 75–93.

Adams, B.F., Busby, C.J., 1994. Stratigraphic evolution of the Lower Cretaceous Alisitos arc, Baja California (Mexico). Abstracts with Programs - Geological Society of America 26 (7), 147.

Adams, B.N., Busby, C.J., Mattinson, J.M., 1997. Jurassic magmatism and sedimentation in the Palen Mountains, southeastern California: implications for regional tectonic controls on the Mesozoic magmatic arc. Geological Society of America Bulletin 109, 1464–1484.

Allison, E.C., 1955. Middle Cretaceous Gastropoda from Punta China, Baja California, Mexico. Journal of Paleontology 29, 400–432.

Arculus, R.J., Gill, J.B., Cambray, H., Chen, W., Stern, R.J., 1995. Geochemical evolution of arc systems in the western Pacific: the ash and turbidite record recovered by drilling. In: Taylor, B., Natland, J. (Eds.), Active Margins and Marginal Basins of the Western Pacific. American Geophysical Union Monograph, vol. 88, pp. 45–65.

Baldwin, S.L., Harrison, T.M., 1989. Geochronology of blueschists from west-central Baja California and the timing of uplift in subduction zones. Journal of Geology 97, 149–163.

Barnes, D.A., 1984. Volcanic arc derived, Mesozoic sedimentary rocks, Vizcaino Peninsula, Baja California Sur, Mexico. In: Frizzell, V.A. (Ed.), Geology of the Baja California Peninsula. Book, 39. Pacific Section Society Economic Paleontologists and Mineralogists, Los Angeles, pp. 119–133.

Barnes, P.M., Audru, J.-C., 1999. Recognition of active strike-slip faulting from high-resolution marine seismic reflection profiles: Eastern Marlborough fault system, New Zealand. Geological Society of America Bulletin 111, 538–559.

Barnes, P.M., Sutherland, R., Davy, B., Delteil, J., 2001. Rapid

creation and destruction of sedimentary basins on mature strike-slip faults: an example from the offshore Alpine Fault, New Zealand. Journal of Structural Geology 23, 1727–1739.
Beggs, J.M., 1984. Volcaniclastic rocks of the Alisitos Group, Baja California, Mexico. In: Frizzell, V.A. (Ed.), Geology of the Baja California Peninsula. Book, 39. Pacific Section Society Economic Paleontologists and Mineralogists, Los Angeles, pp. 43–52.
Boehlke, J.E., Abbott, P.L., 1986. Punta Baja Formation, a Campanian submarine canyon fill, Baja California, Mexico. In: Abbott, P.L. (Ed.), Cretaceous Stratigraphy Western North America, vol. 46. Pacific Section S.E.P.M, Los Angeles, pp. 91–102.
Boles, J.R., Landis, C.A., 1984. Jurassic sedimentary melange and associated facies, Baja California, Mexico. Geological Society of America Bulletin 95, 513–521.
Bottjer, D.J., Link, M.H., 1984. A synthesis of Late Cretaceous southern California and northern Baja California paleogeography. In: Crouch, J.K., Bachman, S.B. (Eds.), Tectonics and sedimentation along the California margin. Book 38. Pacific Section, Society of Economic Paleontologists and Mineralogists, Los Angeles, pp. 79–90.
Buddin, T.S., Stimpson, I.G., Williams, G.D., 1993. North Chilean forearc tectonics and Cenozoic plate kinematics. Tectonophysics 220 (1–4), 193–203.
Busby, C.J., Smith, D.P., Morris, W.R., Adams, B., 1998. Evolutionary model for convergent margins facing large ocean basins: Mesozoic Baja California (Mexico). Geology 26 (3), 227–230.
Busby, C.J., Yip, G., Blikra, L., Renne, P., 2002. Pacific margin example of catastrophic sedimentation triggered by K/T bolide impact. Geology 30, 687–690.
Busby, C.J., Mattinson, J., Fackler-Adams, B., DeOreo, S., 2003. Stratigraphic evolution of the Early Cretaceous Alisitos Arc from San Quintin to Mission San Fernando (Baja California, Mexico). Geological Society of America, Abstracts with Programs 35 (4), 76–77.
Busby, C.J., Bassett, K., Riggs, N., Steiner, M., in press. Climatic and tectonic controls on Jurassic intra-arc basins related to northward drift of North America. Geological Society of America, Special Paper, 40 manuscript pages.
Busby-Spera, C.J., 1987. Lithofacies of deep marine basalts emplaced on a Jurassic backarc apron, Baja California (Mexico). Journal of Geology 95, 671–686.
Busby-Spera, C.J., 1988a. Evolution of a Middle Jurassic back-arc basin, Cedros Island, Baja California: evidence from a marine volcaniclastic apron. Geological Society of America Bulletin 100, 218–233.
Busby-Spera, C.J., 1988b. Development of fan-deltoid slope aprons in a convergent-margin tectonic setting: Mesozoic, Baja California, Mexico. In: Nemec, W., Steel, R.J. (Eds.), Fan Deltas: Sedimentology and Tectonic Settings. Blackie and Son, pp. 419–429.
Busby-Spera, C.J., Boles, J.R., 1986. Sedimentation and subsidence styles in a Cretaceous forearc basin, southern Vizcaino Peninsula, Baja California, Mexico. In: Abbott, P.L. (Ed.), Cretaceous Stratigraphy, Western North America. Pacific Section, Society of Economic Paleontologists and Mineralogists, Los Angeles, pp. 79–90.
Busby-Spera, C.J., Saleeby, J.B., 1990. Intra-arc strike slip fault exposed at batholithic levels in the southern Sierra Nevada, California. Geology 18, 255–259.
Busby-Spera, C.J., White, J.D.L., 1987. Variation in peperite textures associated with differing host sediment properties. Bulletin Volcanologique 49 (6), 765–776.
Busby-Spera, C.J., Morris, W.R., Abbott, P., 1988. Geologic field guide to selected Cretaceous forearc and intra-arc rocks of Baja California Del Norte, Mexico. GSA Field Trip Guide, Cordilleran Section Meeting. U.N.L.V. Dept. of Geosciences Special Pub., vol. 2, pp. 311–330.
Busby-Spera, C.J., Schermer, E.R., Riggs, N.R., Mattinson, J.M., 1990a. Controls on basin subsidence within continental magmatic arcs. International Sedimentological Congress, 13th, Nottingham, England. Abstracts: International Association of Sedimentologists, 70–71.
Busby-Spera, C.J., Mattinson, J.M., Riggs, N.R., Schermer, E.R., 1990b. The Triassic–Jurassic magmatic arc in the Mojave–Sonoran deserts and the Sierran–Klamath region; similarities and differences in paleogeographic evolution. In: Harwood, D., Miller, M. (Eds.), Paleozoic and Early Mesozoic Paleogeographic Relations; Sierra Nevada, Klamath Mountains, and Related Terranes. Special Paper - Geological Society of America, vol. 255, pp. 325–338.
Clift, P.D., 1995. Volcaniclastic sedimentation and volcanism during the rifting of western Pacific backarc basins. In: Taylor, B., Natland, J. (Eds.), Active margins and marginal basins of the western Pacific. American Geophysical Union Monograph, vol. 88, pp. 67–96.
Critelli, S., Marsaglia, K., Busby, C., 2002. Tectonic and sedimentary controls on the composition of backarc volcaniclastic rocks. Geological Society of America Bulletin 114 (5), 515–527.
Crowell, J.C., 1982. The tectonics of Ridge basin, southern California. In: Crowell, J.C., Link, M.H. (Eds.), Geologic History of Ridge Basin Southern California: SEPM Series Book #22, Pacific Section. Society of Economic Paleontologists and Mineralogists, Los Angeles, pp. 25–41.
Cunningham, A.B., Abbott, P.L., 1986. Sedimentology and provenance of the Upper Cretaceous Rosario Formation south of Ensenada, Baja California, Mexico. In: Abbott, P.L. (Ed.), Cretaceous Stratigraphy Western North America, vol. 46. Pacific Section, S.E.P.M, Los Angeles, pp. 103–118.
Dewey, J.F., 1980. Episodicity, sequence and style at convergent plate boundaries. Special Paper - Geological Association of Canada, vol. 20, pp. 553–573.
Dickinson, W.R., Lawton, T.F., 2001. Carboniferous to Cretaceous assembly and fragmentation of Mexico. Geological Society of America Bulletin 113 (9), 1142–1160.
Dickinson, W.R., Hopson, C.A., Saleeby, J.B., 1996. Alternate origins of the Coast Range Ophiolite (California): introduction and implications. GSA Today 6, 1–10.
Dilek, Y., Moores, E.M., 1993. Across-strike anatomy of the Cordilleran orogen at 40°N latitude: implications for the Mesozoic paleogeography of the western United States. In: Dunn, G., McDougall, K. (Eds.), Mesozoic Paleogeography of the Western United States—II. Book 71. Pacific Section, S.E.P.M, Los Angeles, pp. 333–346.

Dilek, Y., Moores, E.M., 1995. Geology of the Humboldt igneous complex, Nevada, and tectonic implications for the Jurassic magmatism in the Cordilleran orogen. In: Miller, D.M., Busby, C. (Eds.), Jurassic Magmatism and Tectonics of the North American Cordillera: Boulder, Colorado. Special Paper - Geological Society of America, vol. 299, pp. 229–248.

Dokka, R.K., 1984. Fission-track geochronologic evidence for Late Cretaceous mylonitization and early paleocene uplift of the northeastern Peninsular Ranges, California. Geophysical Research Letters 11, 46–49.

Dykstra, M., Kneller, B., 2002. Arroyo San Fernando: a Cretaceous submarine valley–levee complex built within and out of a submarine canyon. Abstracts with Programs - Geological Society of America 34 (6), 508.

Engebretson, D.A., Cox, A., Gordon, R.G., 1985. Relative motions between oceanic and continental plates in the Pacific Basin. Special Paper - Geological Society of America, vol. 206. 59 pp.

Engel, A.E., Schultejann, P.A., 1984. Late Mesozoic and Cenozoic tectonic history of south central California. Tectonics 3, 659–675.

Erskine, B.G., Wenk, H.R., 1985. Evidence for Late Cretaceous crustal thinning in the Santa Rosa mylonite zone, southern California. Geology 18, 1173–1177.

Fackler Adams, B.N., 1997. Volcanic and sedimentary facies, processes, and tectonics of intra-arc basins: the Jurassic continental arc of California and the Cretaceous oceanic arc of Baja California. Unpublished PhD thesis. University of California at Santa Barbara. 248 pp.

Fackler Adams, B.N., Busby, C.J., 1998. Structural and stratigraphic evolution of extensional oceanic arcs. Geology 26 (8), 735–738.

Filmer, P.E., Kirschvinck, J.L., 1989. A paleomagnetic constraint on the Late Cretaceous paleoposition of northwestern Baja California, Mexico. Journal of Geophysical Research 94 (B6), 7332–7342.

Fiske, R.S., Naka, J., Iizasa, K., Yuasa, M., 1995. Caldera-forming submarine pyroclastic eruption at Myojin Knoll, Izu-Bonin Arc. JAMSTEC Journal of Deep Sea Research 11, 315–322.

Fiske, R.S., Naka, J., Iizasa, K., Yuasa, M., 2001. Submarine silicic caldera at the front of the Izu-Bonin arc, Japan: voluminous seafloor eruptions of pumice. Geological Society of America Bulletin 113, 813–824.

Flinch, J.F., Bally, A.W., 1992. Extensional basins in a forearc setting, Betic and Rif Cordilleras. AAPG Annual Convention Abstr., 41–42.

Flint, S., Turner, P., 1988. Alluvial fan and fan-delta sedimentation in a forearc extensional setting; the Cretaceous Coloso basin of northern Chile. In: Nemec, W., Steel, R.J. (Eds.), Fan Deltas: Sedimentology and Tectonic Settings. Blackie and Son, Glasgow, pp. 387–399.

Fulford, M.M., Busby-Spera, C.J., 1993. Tectonic controls on nonmarine sedimentation in a Cretaceous fore-arc basin, Baja California, Mexico. In: Frostick, L.E., Steel, R.J. (Eds.), Tectonic Controls and Signatures in Sedimentary Successions. International Association of Sedimentologists Special Publication, vol. 20, pp. 301–333.

Gastil, R.G., 1985. Terranes of peninsular California and adjacent Sonora. In: Howell, D.G. (Ed.), Tectonostratigraphic Terranes of the Circum-Pacific Region: Houston, TX. Circum-Pacific Council for Energy and Mineral Resources, pp. 273–284.

Gastil, G., Allison, E.C., 1966. An upper Cretaceous fault-line coast. AAPG Bulletin (abstr.) 50, 647–648.

Gastil, R.G., Miller, R.H., 1981. Lower Paleozoic strata on the Pacific plate of North America. Nature 292 (5826), 828–830.

Gastil, R.G., Phillips, R.C., Allison, E.C., 1975. Reconnaissance geology of the state of Baja California. Geological Society of America Memoir 140 (170 pp.).

Gastil, R.G., Morgan, G.J., Krummenacher, D., 1978. Mesozoic history of Peninsular California and related areas east of the Gulf of California. In: Howell, D.G., McDougall, K.A. (Eds.), Mesozoic Paleogeography of the Western United States: Pacific Coast Paleogeography Symposium, vol. 2. Pacific Section Society of Economic Paleontologists and Mineralogists, Los Angeles, pp. 107–115.

Gastil, G., Wracher, M., Strand, G., Kear, L.L., Eley, D., Chapman, D., 1992. The tectonic history of the southwestern United States and Sonora, Mexico during the past 100 million years. Tectonics 7, 327–341.

Geist, E.L., Childs, J.R., Scholl, D.W., 1988. The origin of summit basins of the Aleutian Ridge: implications for block rotation of an arc massif. Tectonics 7, 327–341.

Gill, J.B., Seales, C., Thompson, P., Hochstaedter, R., A.G., Dunlap, C., 1992. Petrology and geochemistry of Pliocene–Pleistocene volcanic rocks from the Izu Arc, Leg 126. Proceedings of the Ocean Drilling Program, Scientific Results, College Station 126, 383–404.

Glazner, A., 1991. Plutonism, oblique subduction and continental growth: an example from the Mesozoic of California. Geology 19, 784–786.

George, P.G., Dokka, R.K., 1994. Major Late Cretaceous cooling events in the eastern Peninsular Ranges, California, and their implications for Cordilleran tectonics. Geological Society of America Bulletin 106, 903–914.

Godfrey, N.J., Dilek, Y., 2000. Mesozoic assimilation of oceanic crust and island arc into the North American continental margin in California and Nevada: insights from geophysical data. In: Dilik, Y., Moores, E.M., Elthon, D., Nicolas, A. (Eds.), Ophiolites and Oceanic Crust: New insights from Field Studies and the Ocean Drilling Program: Boulder, Colorado. Geological Society of America Special Paper, vol. 349, pp. 365–382.

Goetz, C.W., 1989. Geology of the Rancho El Rosarito area: evidence for latest Albian, east over west, ductile thrusting in the Peninsular Ranges (MS thesis), California State University at San Diego. 134 pp.

Goetz, C.W., Girty, G.H., Gastil, R.G., 1988. East over west ductile thrusting along a terrane boundary in the Peninsular Ranges: Rancho El Rosarito, Baja California, Mexico. Abstracts with Programs - Geological Society of America 20 (3), 164.

Goodwin, L.B., Renne, P.R., 1991. Effects of progressive mylonitization on Ar retention in biotites from the Santa Rosa mylonite zone, California, and thermochronologic implications. Contributions to Mineralogy and Petrology 108, 283–297.

Graham, S.A., Dickinson, W.R., Ingersoll, R.V., 1975. Himalayan–

Bengal model for flysch dispersal in Appalacjian–Ouachita system. Geological Society of America Bulletin 86, 273–286.

Grove, M., 1993, Thermal histories of southern California basement terranes (PhD thesis), University of California.

Griffith, R.C., 1987. Geology of the southern Sierra Calamuje area: structural and stratigraphic evidence for latest Albian compression along a terrane boundary, Baja California, Mexico (MS thesis), San Diego, California. 119 pp., San Diego State University.

Gromet, L.P., Silver, L.T., 1979. Profile of rare earth element characteristics across the Peninsular Ranges Batholith near the international border, southern California, U.S.A. and Baja California, Mexico. In: Abbott, P.L., Todd, V.R. (Eds.), Mesozoic Crystalline Rocks: Peninsular Ranges Batholith and Pegmatites; Point Sal Ophiolite. San Diego State University, pp. 133–142.

Hickey, J.J., 1984. Stratigraphy and composition of a Jura-Cretaceous volcanic arc apron, Punta Eugenia, Baja California Sur, Mexico. In: Frizzell, V.A. (Ed.), Geology of the Baja California Peninsula. Book 39. Pacific Section Society Economic Paleontologists and Mineralogists, Los Angeles, pp. 149–160.

Iizasa, K., Fiske, R.S., Ishizuka, O., Yuasa, M., Hshimoto, J., Naka, Y., Horii, Y., Fujiwara, Y., Imai, A., Koyama, S., 1999. A Kuroko-type polymetallic sulfide deposit in a submarine silicic caldera. Science 283, 975–977.

Imperato, D.P., 1996. Timing and style of subduction-related deformation, central Great Valley, California. American Association of Petroleum Geologists Annual Meeting Abstr., 69.

Ingersoll, R.V., 1995. Remnant ocean basins. In: Busby, C.J., Ingersoll, R.V. (Eds.), Tectonics of Sedimentary Basins. Blackwell, Cambridge, MA, pp. 363–392.

Ingersoll, R.V., Busby, C.J., 1995. Tectonics of sedimentary basins. In: Busby, C.J., Ingersoll, R.V. (Eds.), Tectonics of Sedimentary Basins. Blackwell, Cambridge, MA, pp. 1–52.

Jarrard, R.D., 1986. Relations among subduction zone parameters. Reviews of Geophysics 24, 184–217.

Jayko, A.S., Blake, M.C., Harms, T., 1987. Attenuation of the Coast Range Ophiolite by extensional faulting and nature of the Coast Range "thrust", California. Tectonics 6, 475–488.

Johnson, S., Tate, M.C., Fanning, C.M., 1999. New geologic mapping and SHRIMP U–Pb zircon data in the Peninsular Ranges batholith, Baja California: evidence for a suture? Geology 27 (8), 743–746.

Karig, D.E., 1983. Temporal relationships between backarc basin formation and arc volcanism with special reference to the Philippine Sea. In: Hayes, D.E. (Ed.), The Tectonic and Geologic Evolution of Southeast Asian Seas and Islands, Part 2. American Geophysical Union Monograph, vol. 27, pp. 318–325.

Kilmer, F.H., 1963. Cretaceous and Cenozoic stratigraphy and paleontology, El Rosario area. PhD thesis (Unpub.). University of California, Berkeley. 149 pp.

Kilmer, F.H., 1977. Reconnaissance geology of Cedros Island, Baja California, Mexico. Southern California Academy of Sciences Bulletin 76, 91–98.

Kilmer, F.H., 1984. Geology of Cedros Island, Baja California, Mexico. Humboldt State University, Arcata. 69 pp.

Kimbrough, D.L., 1984. Paleogeographic significance of the Middle Jurassic Gran Canon Formation, Cedros Island, Baja California. In: Frizzell, V.A. (Ed.), Geology of the Baja California Peninsula. Book 39. Pacific Section Society Economic Paleontologists and Mineralogists, Los Angeles, pp. 107–118.

Kimbrough, D.L., 1985. Tectonstratigraphic terranes of the Vizcaino Peninsula and Cedros and San Benito Islands, Baja California, Mexico. Circum-Pacific Council for Energy and Mineral Resources Earth Sciences Series 1, 285–298.

Kimbrough, D.L., Herzig, C.T., 1994. Late Jurassic/Early Cretaceous deformation in the perbatholithic basement of the Cretaceous Peninsular Ranges batholith magmatic arc—an expression of the J-2 cusp? Abstracts with Programs - Geological Society of America 26 (2), 63.

Kimbrough, D.L., Hickey, J.J., Tosdal, R.M., 1987. U–Pb ages of granitoid clasts in upper Mesozoic arc-derived strata of the Vizcaino Peninsula, Baja California, Mexico. Geology 15, 26–29.

Kimbrough, D.L., Smith, D.P., Mahoney, J.B., Moore, T.E., Grove, M., Gastil, R.G., Otega-Riivera, A., Fanning, C.M., 2001. Forearc-basin sedimentary response to Late Cretaceous batholith emplacement in the Peninsular Ranges of southern and Baja California. Geology 14, 491–494.

Kimura, G., 1986. Oblique subduction and collision: forearc tectonics of the Kurile arc. Geology 14, 404–407.

Krummenacher, D., Gastil, R.G., Bushee, J., Doupont, J., 1975. K–Ar apparent ages, Peninsular Ranges batholith, southern California and Baja California. Geological Society of America Bulletin 86, 760–768.

Lovera, O.M., Grove, M., Harrison, T.M., Mahon, K.I., 1997. Systematic analysis of K-feldspar $^{40}Ar/^{39}Ar$ step heating results: I. Significance of activation energy determinations. Geochemica et Cosmochemica Acta 61 (5), 3171–3192.

Marsaglia, K.M., 1995. Interarc and backarc basins. In: Busby, C.J., Ingersoll, R.V. (Eds.), Tectonics of Sedimentary Basins. Blackwell, Cambridge, MA, pp. 299–330.

Moberly, R., Shepherd, G.L., Coulbourn, W.T., 1982. Forearc and other basins, continental margin of northern and southern Peru and adjacent Ecuador and Chile. In: Leggett, J.K. (Ed.), Trench–Forearc Geology: Sedimentation and Tectonics on Modern and Ancient Active Plate Margins. Geological Society of London, pp. 171–189.

Moore, T.E., 1984. Sedimentary facies and composition of Jurassic volcaniclastic turbidites at Cerro El Calvario, Vizcaino Peninsula, Baja California Sur, Mexico. In: Frizzell, V.A. (Ed.), Geology of the Baja California Peninsula. Book 39. Pacific Section Society Economic Paleontologists and Mineralogists, Los Angeles, pp. 131–148.

Moore, T.E., 1985. Stratigraphy and tectonic significance of the Mesozoic tectonstratigraphic terranes of the Vizcaino Peninsula, Baja California Sur, Mexico. Circum-Pacific Council for Energy and Mineral Resources Earth Sciences Series, vol. 1, pp. 315–329.

Morris, W.J., 1974. Upper Cretaceous "El Gallo" formation and its vertebrate fauna. In: Gastil, G., Lillegraven, J. (Eds.), A Guidebook to the Geology of Peninsular California. AAPG-SEPM-SEG Pacific Section, Los Angeles, CA, pp. 60–66.

Morris, W.R., 1992. The depositional framework, paleogeography and tectonic development of the Late Cretaceous through Paleocene Peninsular Range forearc basin in the Rosario Embayment, Baja California, Mexico (PhD thesis), Santa Barbara, Doctoral Dissertation, University of California. 295 pp.

Morris, W.R., Busby, C.J., 1996. The effects of tectonism on the high-resolution sequence stratigraphic framework of non-marine to deep-marine deposits in the Peninsular Ranges forearc basin complex. In: Abbott, P.L., Cooper, J.D. (Eds.), Field Conference Guide 1996, Pacific Section American Association of Petroleum Geologists Guidebook. 73. Book 80. Pacific Section SEPM, pp. 381–408.

Morris, W.R., Busby-Spera, C.J., 1988. Sedimentologic evolution of a submarine canyon in a forearc basin, Late Cretaceous Rosario Formation, San Carlos, Mexico. American Association Petroleum Geologists, Bulletin 72, 717–737.

Morris, W.R., Busby-Spera, C.J., 1990. A submarine fan valley–levee complex in the Upper Cretaceous Rosario Formation, Baja California: implications for turbidite facies models. Geological Society America Bulletin 102, 900–914.

Morris, W.R., Smith, D.P., Busby-Spera, C.J., 1989. Conglomerate facies and processes in deep marine Cretaceous forearc basins of Baja California, Mexico. In: Colburn, I., Abbott, P. (Eds.), Conglomerates and Basin Analysis. Pacific Section Society of Economic Paleontologists and Mineralogists, Los Angeles, pp. 123–142.

Nilsen, T.H., Abbott, P.L., 1981. Paleogeography and sedimentology of Upper Cretaceous turbidites, San Diego, California. American Association of Petroleum Geologists Bulletin 65, 1256–1284.

Nilsen, T.H., Sylvester, A.G., 1995. Strike-slip basins. In: Busby, C.J., Ingersoll, R.V. (Eds.), Tectonics of Sedimentary Basins. Blackwell, Cambridge, MA, pp. 425–457.

Nishimura, A., et al., 1991. Pliocene–Quaternary submarine pumice deposits in the Sumisu Rift, Izu-Bonin Arc. In: Fisher, R.V., Smith, G.A. (Eds.), Sedimentation in Volcanic Settings. Society of Economic Paleontologists and Mineralogists Special Publication, vol. 45, pp. 201–208.

Nishimura, A., Rodolfo, K., Koizumi, A., Gill, J., Fujioka, K., 1992. Episodic deposition of Pliocene–Pleistocene pumice deposits of Izu-Bonin arc, Leg 126. Proceedings of the Ocean Drilling Program, Scientific Results, College Station 126, 3–21.

Patterson, D.L., 1984. Los Chapunes and Valle sandstones: cretaceous petrofacies of the Vizcaino basin, Baja California, Mexico. In: Frizzell, V.A. (Ed.), Geology of the Baja California Peninsula. Book 39. Pacific Section Society Economic Paleontologists and Mineralogists, Los Angeles, pp. 161–182.

Phillips, J.R., 1993. Stratigraphy and structural setting of the Mid-Cretaceous Olvidada Formation, Baja California Norte, Mexico. In: Gastil, R.G., Miller, R.H. (Eds.), The Prebatholithic Stratiography of Peninsular California. Geological Society of America, Boulder, CO, pp. 97–106.

Platte, J.P., 1986. Dynamics of orogenic wedges and the uplift of high-pressure metamorphic rocks. Geological Society of America Bulletin 97, 1037–1053.

Rangin, C., 1978. Speculative model of Mesozoic geodynamics, central Baja California to northeastern Sonora (Mexico). In: Howell, D.G., McDougall, K. (Eds.), Mesozoic Paleogeography of the Western United States. Pacific Section SEPM, Los Angeles, CA, pp. 85–106.

Renne, P.R., Fulford, M.M., Busby-Spera, C.J., 1991. High resolution $^{40}Ar/^{39}Ar$ chronostratigraphy of the Late Cretaceous El Gallo Formation, Baja California de Norte, Mexico. Geophysical Research Letters 18 (3), 459–462.

Saleeby, J.B., Busby-Spera, C.J., 1992. Early Mesozoic tectonic evolution of the western U.S. Cordillera. In: Burchfiel, B.C., et al. (Eds.), The Cordilleran Orogen: Co-terminus U.S.: Geology of North America, vol. G-3. Geological Society of America, Boulder, CO, pp. 107–168.

Schweickert, R.A., Cowan, D.S., 1975. Early Mesozoic tectonic evolution of the western Sierra Nevada, California. Geological Society of America Bulletin 86, 1329–1336.

Sedlock, R.L., 1988. Tectonic setting of blueschist and island-arc terranes of west-central Baja California, Mexico. Geology 16, 205–233.

Sedlock, R.L., 1993. Mesozoic geology and tectonics of blueschist and associated oceanic terranes in the Cedros–Vizcaino–San Benito and Magdalena–Santa Margarite regions, Baja California. In: Dunne, G.C., McDougall, K.A. (Eds.), Mesozoic Paleogeography of the Western United States – II. Book 71. Pacific Section, Society of Economic Paleontologists and Mineralogists, pp. 113–125.

Sedlock, R.L., 1996. Syn-subduction forearc extension and blueschist exhumation in Baja California, Mexico. In: Bebout, G.E., Scholl, D.W., Kirby, S.H., Platte, J.P. (Eds.), Dynamics of Subduction. American Geophysical Union, Monograph, vol. 96, pp. 155–162.

Sedlock, R.L., in press. Four phases of Mesozoic deformation in the Sierra de San Andres ophiolite, Vizcaino Peninsula, west-central Baja California. In: Johnson, S., Paterson, S., Kimbrough, D., Martin-Barajas, A., Girty, G. (Eds.), The Tectonic Evolution of Northwestern Mexico and the Southwestern U.S.A., Geological Society of America Special Paper, 374.

Sedlock, R.L., Ortega-Gutierrez, F., Speed, R.C., 1993. Tectonostratigraphic terranes and tectonic evolution of Mexico. Geological Society America Special Paper 278 (153 pp.).

Shepard, F.P., Dill, R.F., 1966. Submarine Canyons and Other Sea Valleys. Rand McNallet, Chicago. 381 pp.

Silver, L.T., 1986. Observations on the Peninsular Ranges batholith, southern California and Mexico, in space and time. Abstracts with Programs - Geological Society of America 18, 184.

Silver, L.T., Chappal, B.W., 1988. The Peninsular Ranges batholith: an insight onto the evolution of cordilleran batholiths of southwestern North America. Transactions of the Royal Society of Edinburgh Earth Science 79, 105–121.

Silver, L.T., Stehli, F.G., Allen, C.R., 1963. Lower Cretaceous prebatholithic rocks of northern Baja California, Mexico. American Association of Petroleum Geologists Bulletin 47, 2054–2059.

Silver, L.T., Taylor, H.P., Chappell, B., 1979. Some petrological and geochronological observations of the Peninsular Ranges batholith near the international border of the U.S.A. and Mexico. Geological Society of America, Annual Meeting, Field Trip Guide, Boulder, CO, pp. 83–110.

Smith, D.P., Busby, C.J., 1993. Mid-Cretaceous crustal extension

recorded in deep-marine half-graben fill, Cedros Island, Mexico. Geological Society America Bulletin 105, 547–562.
Smith, G.A., Landis, C.A., 1995. Intra-arc basins. In: Busby, C.J., Ingersoll, R.V. (Eds.), Tectonics of Sedimentary Basins. Blackwell, Cambridge, MA, pp. 263–298.
Smith, D.P., Sedlock, R.L., Kimbrough, D.L., Busby-Spera, C.J., 1991. Mesozoic geology of Cedros Island. GSA Annual Meeting Field Trip Guide, San Diego. 20 pp.
Sutherland, M., Wetmore, P.H., Herzig, C., Paterson, S.R., 2002. The early Cretaceous Santiago Peak arc: a continental margin arc built on the North American Triassic–Jurassic accretionary prism of southern and Baja California. Abstracts with Programs - Geological Society of America 34 (6), 43.
Taylor, H.P., Silver, L.T., 1978. Oxygen isotope relationships in plutonic igneous rocks of the Peninsular Ranges Batholith, southern and Baja California, Short Papers of the Fourth International Conference on Geochronology, Cosmochronology, and Isotope Geology, U.S. Geological Survey Open-File Report OF-78-701, 423-426.
Taylor, B., et al., 1990. ALVIN-Sea Beam studies of Simisu rift, Izu-Bonin arc. Earth and Planetary Science Letters 100, 127–147.
Thomson, C.N., Girty, G.H., 1994. Early Cretaceous intra-arc ductile strain in Triassic–Jurassic and Cretaceous continental margin arc rocks, Peninsular Ranges, California. Tectonics 13, 1108–1119.
Todd, V.R., Erskine, B.G., Morton, L., 1988. Metamorphic and tectonic evolution of the northern Peninsular Ranges batholith, southern California. In: Ernst, W.G. (Ed.), Metamorphism and Crustal Evolution of the western United States (Rubey Volume VII). Prentice Hall, Englewood Cliffs, pp. 895–937.
Umhoefer, P., in press, A speculative model on the North America Cordillera in the Early Cretaceous: Tectonic escape related to arc collision of the Guerrero terrane and a change in North America plate motion. In: Johnson, S., Paterson, S., Kimbrough, D., Martin-Barajas, A., Girty, G. (Eds.), The Tectonic Evolution of Northwestern Mexico and the Southwestern U.S.A., Geological Society of America Special Paper 374.
Walawander, M.J., Gastil, R.G., Clinkenbeard, J.P., mcCormick, W.V., Eastman, B.G., Wenicke, R.S., Wardlaw, M.S., Gunn, S.H., Smith, B.M., 1990. Origin and evolution of the zoned La Posta-type plutons, eastern Peninsular Ranges batholith, southern and Baja California. In: Anderson, J.L. (Ed.), The Nature and Origin of Cordilleran Magmatism. Geological Society of America, Boulder, CO, pp. 1–18.
Wetmore, P.H., Schmidt, K.L., Paterson, S.R., Herzig, C., 2002. Tectonic implications for the along-strike variation of the Peninsular Ranges batholith, southern and Baja California. Geology 30, 247–250.
Wessel, J.K., Fryer, P., Wessel, P., Taylor, B., 1994. Extension in the northern Mariana inner forearc. Journal Geophysical Research, B 99 (8), 15181–15203.
Whalen, P.A., Pessagno Jr., E.A., 1984. Lower Jurassic radiolaria, San Hipolito Formation, Vizcaino Peninsula, Baja California Sur. In: Frizzell, V.A. (Ed.), Geology of the Baja California Peninsula. Book 39. Pacific Section Society Economic Paleontologists and Mineralogists, Los Angeles, pp. 53–65.
White, J.D.L., Busby-Spera, C.J., 1987. Deep marine arc apron deposits and syndepositional magmatism in the Alisitos Group at Punta Cono, Baja California, Mexico. Sedimentology (October issue) 34 (5), 911–927.
Wood, R.A., Pettinga, J.R., Bannister, S., Lamarche, G., McMorran, T.J., 1994. Structure of the Hanmer strike-slip basin, Hope Fault, New Zealand. Geological Society of America Bulletin 106, 1459–1473.
Yeo, R.K., 1984. Sedimentology of Upper Cretaceous strata, northern Baja California, Mexico. In: Abbott, P.L. (Ed.), Upper Cretaceous Depositional Systems in Southern California–Northern Baja California. Book 36. Pacific Section, Society of Economic Paleontologists and Mineralogists, Los Angeles, pp. 109–120.