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Tectonic implications of post–30 Ma Pacific and North American relative plate motions

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ABSTRACT

The Pacific plate moved northwest relative to North America since 42 Ma. The rapid half rate of Pacific-Farallon spreading allowed the ridge to approach the continent at about 29 Ma. Extinct spreading ridges that occur offshore along 65% of the margin (Lonsdale, 1991) document that fragments of the subducted Farallon slab became captured by the Pacific plate and assumed its motion prior to the actual subduction of the spreading ridge. This plate-capture process can be used to explain much of the post–29 Ma Cordilleran North America extension, strike slip, and the inland jump of oceanic spreading in the Gulf of California. The Pacific and North American contact zone lengthened with each successive plate capture event, underpinning the parts of western North America directly inland with a strong plate undergoing Pacific relative motion. We suggest that much of the post–29 Ma continental tectonism is the result of the strong traction imposed on the deep part of the continental crust by the gently inclined slab of subducted oceanic lithosphere as it moved to the northwest relative to the overlying continent. The plate-capture hypothesis is distinctly different from theories involving shallow slab gaps. Kinematic problems associated with shallow slab-gap models cause us to question them. This conclusion is consistent with seismic refraction interpretations that suggest there is an inclined layer with high velocities like that of basalt or gabbro at the base of the continental crust beneath much of the Californian margin and the documented reduction of slab-pull forces and density associated with young subducting slabs. Thermal and rheologic modeling suggests that coastal California was a strong zone at all depths allowing it to be firmly linked to Pacific motion. Our model shows that deformed regions such as the basin and range and borderland provinces developed in predicted weak parts of the crustal section, but they have been incompletely linked to the deep plate across the ductile middle and lower crustal layer.

INTRODUCTION

Purpose and Scope

The global-circuit and hot-spot-reference plate reconstruction models (Stock and Molnar, 1988; Engebretson et al., 1985) show that the Pacific plate has moved to the northwest relative to a reference fixed on the interior of North America since the formation of anomaly (A) 18, 41.5–42.5 m.y. ago (Fig. 1; after Stock and Molnar, 1988). The rate of longitudinal displacement (the westerly component) has been about 30 km/m.y., depending on the latitude examined. The half spreading rate on the Pacific/Farallon ridge was slightly greater, allowing the ridge axis to migrate slowly eastward toward

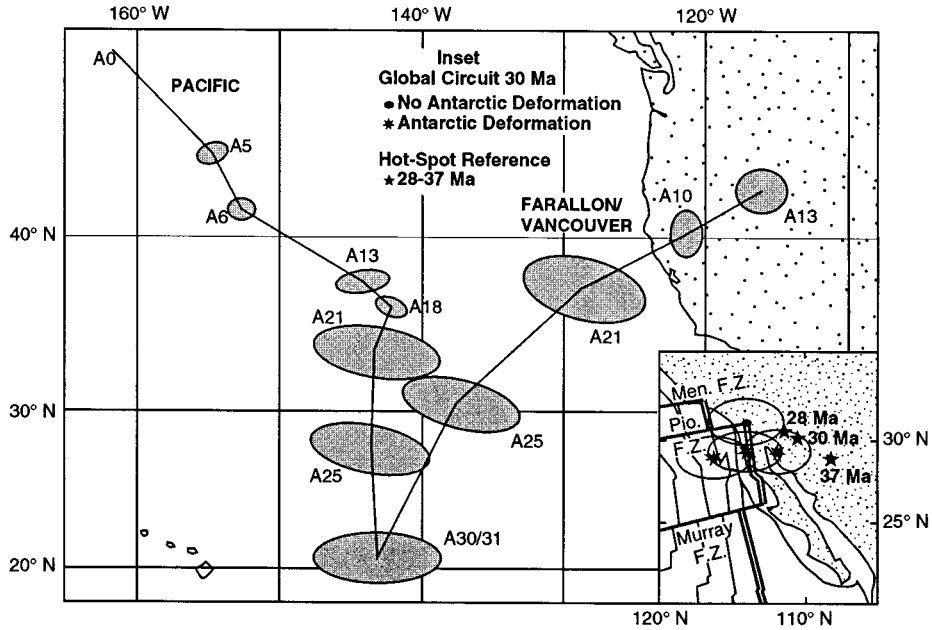

Figure 1. Cumulative displacement of points on the Pacific and Farallon/Vancouver plates relative to fixed North America for the times of anomalies 0, 5, 6, 13, 18, 25, and 30/31 from Stock and Molnar (1988, Fig. 4). A0 = present; A5 ≈ 9–10.5 Ma; A6 ≈ 19.5–20.5 Ma; A13 ≈ 35.5 Ma; A18 ≈ 41.5–42.5 Ma; A25 ≈ 59 Ma; and A30/31 ≈ 69 Ma (Berggren et al., 1985; Kent and Gradstein, 1985). Ellipses indicate uncertainty in point positions. Geographic coordinates set to fixed North American frame. Inset shows position of Pacific/Farallon spreading ridge relative to North America at 30 Ma when initial contact occurred (also from Stock and Molnar, 1988, Fig. 8). Compare position of intersection of spreading ridge with Mendocino fracture zone derived from global-circuit model (Stock and Molnar, 1988) and hot-spot-reference model (Engebretson et al., 1988).

GSA Bulletin; August 1995; v. 107; no. 8; p. 937–959; 15 figures; 1 table; 1 insert.

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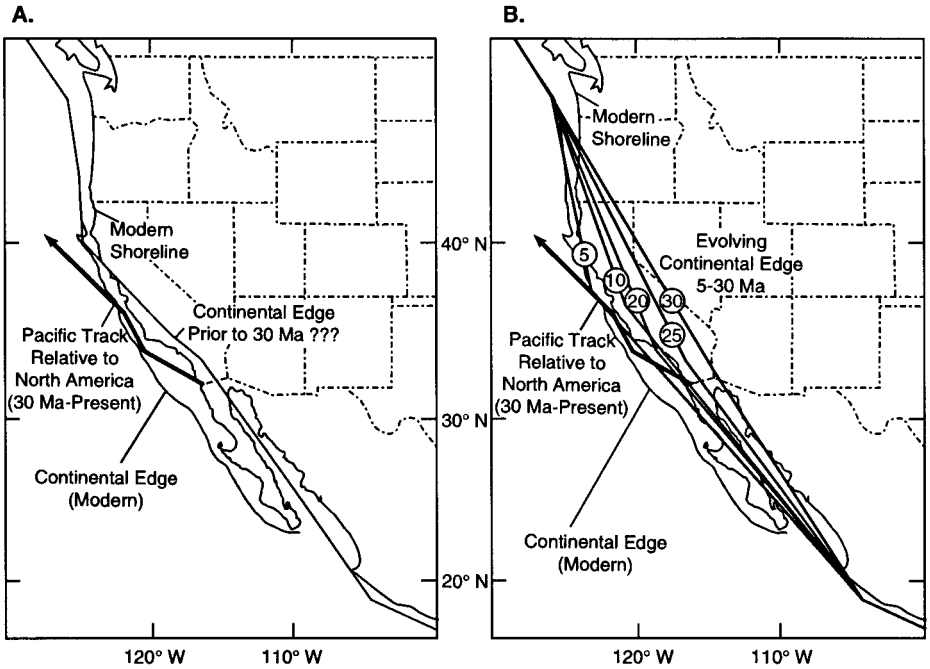

Figure 2. Two possible scenarios for the post–30 Ma evolution of the North American continental edge. Ends of arrows denoting Pacific track relative to North America are positioned at present intersection of A10 and Mendocino fracture zone and at 30 Ma location provided by Stock and Molnar (1988).

A. Continental edge starts at 30 Ma, in favorable configuration to take advantage of future relative plate motions between Pacific and North American plates once contact between those plates has been achieved. Kinematic consequences are as follows: (1) continental edge offshore of Oregon and Washington remains fixed relative to interior of North America; (2) Pacific/North American pole of rotation is stable for 30 Ma; (3) San Andreas position and orientation are relatively fixed; (4) allows for simple shift from subduction and convergence to margin-parallel strike-slip following Pacific/North American contact; (5) makes little or no allowance for margin deformation associated with Cordilleran extension; (6) requires unlikely large bends in pre–30 Ma shoreline "in anticipation" of arrival of spreading ridge at 114 W and 31 N (Stock and Molnar, 1988) 30 m.y. ago.

B. "Bowstring model" used by Severinghaus and Atwater (1990). Assumes that continental edge evolved to present configuration from an approximately straight initial geometry located east of the present margin in response to Pacific plate drift. Kinematic consequences are as follows: (1) continental edge is not fixed relative to continental interior; (2) Pacific/North American pole is unstable; (3) San Andreas deforms with time; (4) transition from subduction is not simple convergence to strike slip; allows for as much post-subduction extension within continent as it does strike-slip faulting; (5) no particular "pre-configuration" of continental edge is necessary—30 Ma position limited only by high-end Cordilleran extension models.

North America. The initial contact of the eastern accreting edge of the Pacific plate with the western edge of the North American plate occurred at 29 Ma just south of the Pioneer fracture zone (Atwater and Severinghaus, 1989). Since 29 Ma the Pacific plate has moved west by about 10.2° of longitude, which corresponds to slightly over 900 km at the latitude of San Francisco. This compares to 8.3° of latitudinal displacement (about 920 km) during the same time period. The segment of the North American edge at the initial contact point has kept pace with the rapid west motion of the Pacific plate for the last 29 m.y.

The Pacific plate moved almost due north relative to North America until the formation of A18, when it assumed a rapid northwest motion track (Fig. 1). It is clear that there has not been any Neogene Pacific plate convergence relative to the stable interior of North America, but Farallon/Vancouver plate relative motion was consistently toward the northeast, and rapidly convergent, throughout the same time period. There is no indication that the Farallon/Vancouver plates were affected by the large change in Pacific relative motion at about 42 Ma (Fig. 2). Because both the Pacific and the Farallon/Vancouver plates have had a significant northward component of motion relative to North America, strike-slip deformation might be expected along the margin at any time during the Tertiary. The big change, once Pacific contact had occurred, was from oblique plate convergence in the Farallon/Vancouver regime to oblique divergence in the Pacific regime.

The importance of the divergent component of Pacific plate motion relative to North America has gone unrecognized or it has been ignored primarily because the modern continental edge is oriented parallel to the plate-motion track (Fig 2A). The tacit assumption that this has always been true has promoted the widespread application of the rigid-pole, transform-fault paradigm to the plate boundary. In that view, the San Andreas transform boundary lies between an active spreading center that lies offshore of the southern part of the Gulf of California and a triple junction near Cape Mendocino (Wilson, 1965; Atwater, 1970; Powell, 1993). Because transform faults lie on small circles about the pole of relative motion, large amounts of extension and/or compression are not associated with them (Wilson, 1965; McKenzie and Parker, 1967; McKenzie and Morgan, 1969). The paradigm works well in this regard for transform faults between rigid plates of oceanic lithosphere because the position of the rotation pole that describes the transform motion remains stable for long time periods. The stable-pole hypothesis fails to describe the San Andreas boundary because it requires a comparable stability in the orientations and relative positions of the fault and the western edge of the North American continent. The fault and continental edge would have had to have been aligned roughly parallel to the relative plate motion direction since the initial plate contact at 29 Ma (Fig. 2A). An overwhelming volume of evidence indicates that this has not been so. Large parts of Cordilleran North America have undergone extension during the last 30 m.y. (e.g., Wernicke, 1992), so the continental edge has no doubt changed orientation and position through time.

Ingersoll (1982) and Severinghaus and Atwater (1990) recognized that the rigid-

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plate transform paradigm fails to accurately describe the Pacific/North American plate boundary. They assumed that the Cascadia continental edge, offshore of Oregon and Washington, was probably not fixed relative to the continental interior and that the continental margin of North America deformed as if pinned to the Pacific plate at the Mendocino fracture zone with pivot points off Cascadia and the southern Gulf of California (Fig. 2B).

In addition to margin deformation, the Pacific/North American plate boundary evolved in the last 29 m.y. from a short simple contact at the subduction interface to the present complex zone, over 2300 km long, that partly resides within the continent and partly includes a small new ocean in the Gulf of California (Atwater, 1970; Atwater and Molnar, 1973; Lonsdale, 1991). Most workers, including Ingersoll (1982) and Severinghaus and Atwater (1990), explained the plate boundary evolution in the context of the migrating-triple-junction paradigm (Atwater, 1970; Atwater and Molnar, 1973). However, there is nothing inherent in that concept to explain the deformation of the margin or the inland jump of the plate boundary.

Lonsdale (1991) has convincingly shown that extinct ridge segments and the corresponding fragments of the Farallon plate are still preserved offshore between 38°N and 35°N (Monterey microplate system), and (Soledad/Guadalupe/Magdalena ridge system) Lonsdale's (1991) observation limits the applicability of the uniformly migrating-triple-junction hypothesis from 30.5°N to 23°N because the active ridge never actually contacted the continent along at least 65% of the length of the plate margin. As such, it can be demonstrated that much of the boundary lengthened through a series of plate-capture events. Nicholson et al. (1994) described some possible consequences of one of these events involving the Monterey microplate. Each time a ridge segment died offshore, the corresponding partially subducted fragment of the Farallon plate must have assumed the rapid west motion of the Pacific plate relative to the interior of North America.

The ultimate fate of the Farallon plate system along segments of the margin that lack fossil spreading ridges offshore is not obvious in the magnetic record. In one such area, between the Mendocino and Farallon fracture zones north of San Francisco, north-south magnetic anomalies that get younger to the east are present at the continental edge and their uniform spacing suggests that there was little change in the half spreading rate with proximity to the continent (Atwater and Severinghaus, 1989). This is the region where the ridge first encountered the continent (Stock and Molnar, 1988). The other area is between the Monterey (at 35°N) and Soledad (at 30.5°N) fossil ridges offshore of southern California and northern Baja California. Here the anomalies form a south-widening fan, and they truncate at a low angle against the continental edge (Lonsdale, 1991). We argue herein that plate capture is still possible in these two areas.

In this paper we examine the plate capture phenomenon with respect to post–30 Ma tectonism in the southwestern United States and northwestern Mexico. We explore a simple model of continental deformation like those of Ingersoll (1982) and Severinghaus and Atwater (1990), but we argue that much of the post–30 Ma extensional tectonism in Cordilleran North America resulted primarily from plate capture and the divergent component of Pacific relative motion. We think the effect of the initial contact of the spreading ridge with the continent was the opposite of the crustal shortening that occurred as a result of India indenting Eurasia. Cordilleran North America expanded westward to fill the widening gap between the rigid Pacific plate and the more coherent interior of North America. An important part of our model revolves around an analysis of the sea-floor spreading history of relevant parts of the eastern Pacific Ocean with respect to the stable interior of North America. This analysis is presented in the context of the global circuit models of Stock and Molnar (1988). Simple kinematic arguments are used to demonstrate that plate capture, or simple changes in ridge dynamics are probably inescapable consequences of a close encounter of ridge and continent under the known conditions of relative plate motions. We conclude that shallow (lithospheric level) slab-free zones (e.g., Dickinson and Snyder, 1979; Zandt and Furlong, 1982; Severinghaus and Atwater, 1989; Furlong et al., 1989) are unlikely to develop for two reasons: (1) because of the unfavorable kinematic consequences created by the rapid west displacement of the Pacific plate and (2) because of the lower strength, reduced slab-pull force, and slightly greater buoyancy associated with young subducting slabs. We further analyze the plate capture phenomenon with respect to strength envelopes in simple lithospheric models. Last, the main elements of post–30 Ma Cordilleran tectonism are explained within the context of the developed model.

Assumptions

We assume that the western edge of North America was not static relative to the interior of North America during the last 30 m.y. and that the present San Andreas geometry can, at best, only be used to define an instantaneous pole of rotation for Pacific/North American relative motion. We carry this one step farther and use the relative motion of the Pacific plate as defined by Stock and Molnar (1988) and Engebretson et al. (1985) to define the longitudinal position of the continental edge through time. Thus, the margin resided at a more easterly position at 30 Ma than it does today, and it migrated west, following the Pacific plate. Maximum limits for the west drift of the Cascadia continental edge are provided by high-end estimates of post–30 Ma extension in the Great Basin (Hamilton and Myers, 1966; Hamilton, 1987; and Wernicke et al., 1988).

We assume that no part of the Pacific plate, or any fragment of the Farallon plate once attached to the Pacific plate, has been subducted beneath North America. The rapid divergence of the Pacific plate relative to the North American interior precludes such subduction. Our analysis requires that, once in contact, the west edge of Cordilleran North America keeps up with the Pacific plate edge, but does not overtake it. Thus, the magnetic anomaly pattern preserved offshore today represents the entire accretion record that formed the east margin of the Pacific plate. The validity of this assumption is dependent on the shape of the continental margin to some extent. Segments of the margin with westerly orientations might become sites of local underthrusting, especially during times following small changes in the relative plate motions.

We assume that continental rifting and sea-floor spreading are passive processes with respect to fluid flow in the mantle beneath the mechanical boundary layer. Thus, (1) plate geometry and kinematics reflect the mechanical properties of the solid lithosphere, rather than that of the fluid mantle (Davies, 1988; McKenzie and Bickle, 1988); (2) Cordilleran rifting is the result of global lithosphere dynamics rather than the product of some local convectional abnormality in the mantle; and (3) melt generation at spreading centers results primarily from upwelling of mantle under adiabatic conditions

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at a rate that is controlled entirely by plate divergence (McKenzie and Bickle, 1988). The passive-rift assumption is commonplace in studies of the oceanic realm, but it is less commonly made in analyses of continental rifts, particularly in the case of the Cordilleran Basin and Range structural province (i.e., Dickinson and Snyder, 1979; Elston, 1984; Gans et al., 1989; Coney, 1987).

Our reason for assuming a passive origin for post–30 Ma Cordilleran extension is simple. We consider the longitudinal displacement of the Pacific plate relative to the interior of North America to be well documented (Stock and Molnar, 1988; Engebretson et al., 1985). It is important to note that this displacement was occurring prior to Pacific/North American contact and that it remained unchanged by that contact (Sock and Molnar, 1988). We show that once contact occurred, the part of the western edge of North America that was in contact was also displaced by the same amount as the Pacific plate. Thus, at least a coarse link between the amounts of post–30 Ma Cordilleran extension and Pacific plate relative motion is easy to establish.

METHODS USED FOR RECONSTRUCTIONS

Our model relies on palinspastic reconstructions of Pacific and North American features. Stock and Molnar (1988) and Engebretson et al. (1985) provided the framework for such reconstructions. There are important conceptual differences between the global-circuit and hot-spot models (Atwater, 1989), but actual differences in the predicted relative locations are relatively small when only the last 30 m.y. are considered (Fig. 1). We use the position of initial Pacific/North American contact at the intersection of A10 (roughly 30 Ma) and the Pioneer fracture zone as provided by the global-circuit model with no Antarctic deformation (≈114°W and 31°N, relative to fixed North America; Stock and Molnar, 1988) throughout this paper. Relative positioning for each reconstruction is based on this 30 Ma position and the Pacific motion track given by Stock and Molnar (1988). Figure 3A shows the relative motion track that we use in our reconstructions in 1 m.y. increments, based on linear interpolation between positions provided by Stock and Molnar (1988). Translations were done planimetrically on a Mercator projection, so some problems arise with scaling and the orientation of features. Scale distortion is minimal in the 30°N–40°N range, but it increases to the south. A proper spherical translation produces an ≈8° counterclockwise rotation of linear features at 30 Ma with respect to the fixed North American graticule (Stock and Molnar, 1988). We rectify the angular deficiencies of the planimetric shift by adding a linearly interpolated angular translation to our Pacific features. The models of Stock and Molnar (1988) and Engebretson et al. (1985) locate the interior parts of the Pacific and North American plates relative to one another, but tectonic reconstructions in the plate contact zone must also take into account the history of sea-floor spreading near the boundary, the large-scale lateral displacements, rotations, and extension in the coastal California/Baja California region, and movements related to the extensional history of the Basin and Range structural province.

The maps and conclusions of Lonsdale (1991) provided much of the information needed for our analysis of the sea-floor spreading history in the eastern Pacific. Regional maps of Atwater and Severinghaus (1989) were used in areas not covered by him. Reconstructed ridge geometry is based on the magnetic pattern preserved in the Pacific plate, which we regard as rigid over the time period examined. Once formed, anomalies on the Pacific plate assume its motion relative to North America, regardless of spreading rate or ridge orientation.

The lateral displacement that we restore on each of the large strike-slip faults of coastal California and Baja California is summarized in Table 1 and depicted graphically on Figure 3B.

North of the Garlock fault, the largest right-slip displacements are documented on the San Andreas, the Rinconada/Huasna, and the Sur/San Gregorio faults (Stewart and Crowell, 1992). On land the San Andreas displacement is restored based on the model of Stanley (1987), and it totals 330 km, with 240 km occurring since 5 Ma. There is no evidence of San Andreas displacement during the period 30–24 Ma (Stanley, 1987). Rinconada/Huasna displacement, totaling 60 km, is based primarily on Ross's (1984) study. Timing on this segment is poorly constrained. Faults in the Santa Maria River/Santa Ynez River areas (Fig. 3B) were probably active during clockwise rotation of the western Transverse Ranges (Hornafius et al., 1986; Luyendyk, 1991), but little is known of their actual displacement histories. The primary constraints for offshore displacement come from Ross (1984) and Clark et al. (1984), who suggested 115–150 km of displacement on the San Gregorio fault since late Miocene time.

South of the Garlock fault, the San Andreas, San Gabriel, San Jacinto, and Elsinore faults (Fig. 3B) account for most of the right-slip displacement on land (Stewart and Crowell, 1992). It is reasonably agreed that displacement on the San Andreas and closely related faults totals about 270 km since the late Miocene, and that the San Gabriel fault was active between 5 and 11 Ma with a total displacement of about 60 km (Crowell, 1962, 1982; Powell, 1982, 1993; Stewart and Crowell, 1992). Disagreements persist over which parts of the San Andreas system were important when and what the relative positions of individual blocks may have been at any one time, but these are relatively inconsequential for our purposes. Powell (1993) provided the most complete and up-to-date synthesis, and it is used herein. The San Jacinto and Elsinore faults (Fig. 3B) have 28–30 and 5–40 km of displacement, respectively, all occurring in the last 5 Ma (Sharp, 1967; Stewart and Crowell, 1992; Powell, 1993).

Post–30 Ma displacements west of the southern part of the San Andreas fault system are related to the clockwise rotation of the western Transverse Ranges (Luyendyk et al., 1980, 1985; Hornafius et al., 1986; Jackson and Molnar, 1990; Luyendyk, 1991) and to right-slip and crustal extension in the California Continental Borderland (Legg, 1985, 1991; Crouch, 1979, 1981; Crouch and Suppe, 1993). Although several slightly different models have been presented to describe the kinematics of Transverse Range rotation, they are all limited to 90°–105° of total rotation, most of which occurred during Miocene time (Hornafius et al., 1986). The models of Legg (1991), Crouch and Suppe (1993), and Sedlock and Hamilton (1991) provided a kinematic link between large-magnitude extension in the borderland province and the clockwise rotation in the Transverse Ranges. The essential element of the borderland models is the palinspastic "closure" of a large exposure of Catalina Schist that lies between the East Santa Cruz Basin fault and the Newport Inglewood fault zone (Fig. 3B) by a combination of detachment-related extension and right-slip faulting (Crouch and Suppe, 1993). In the Crouch and Suppe (1993) model, the two arrows on Figure 3B, one to the west of the East Santa Cruz Basin fault and the other east of the Newport Ingle-

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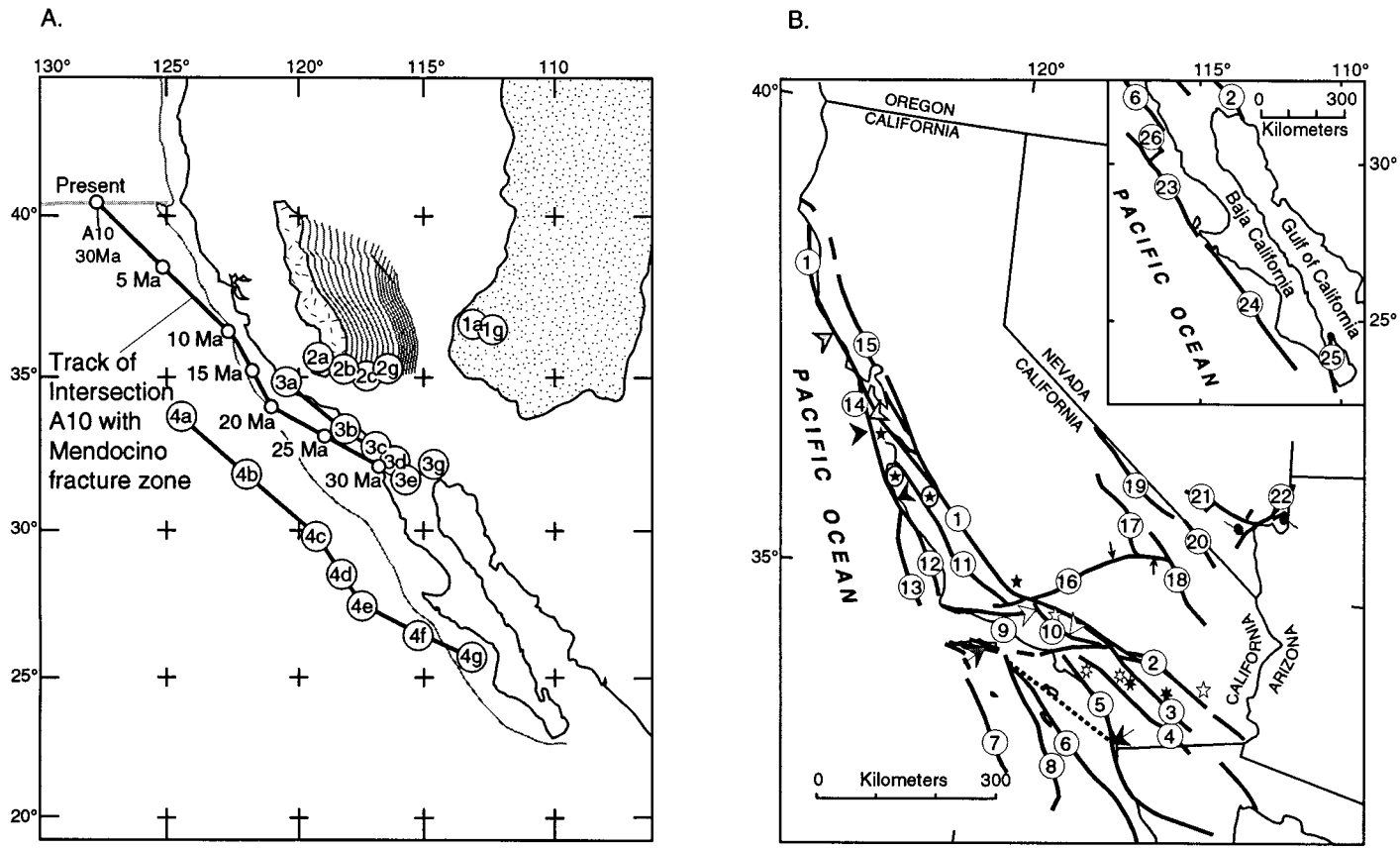

Figure 3. Displacements used in reconstructing Pacific/North American relative positions and configurations at different times.

A. Motion tracks of Pacific plate and selected Cordilleran features relative to frame fixed on interior of North America. Stippled area is Colorado Plateaus and Rocky Mountains; cross-hatch area is Sierra Nevada Range. Ghosted east edge of Sierra Nevadas indicates relative position in 1 m.y. increments since 17 Ma. Track with large open circles, small black dots, and marked ages represents position in 1 m.y. increments of intersection of anomaly 10 (30 Ma) and Mendocino fracture zone, based on Stock and Molnar (1988). Open circles with interior labels represent past and present locations of the following: 1, Saint George, Utah, on Colorado Plateau; 2, Bakersfield, California, west of Sierra Nevada Range; 3, Santa Maria, California, in present coastal zone; and 4, intersection of magnetic anomaly 10 and Murray fracture zone on Pacific plate. These features were chosen because of their present rough east-northeast alignment. Position of each in past is marked by letter designation: a, present; b, 5 Ma; c, 10 Ma; d, 15 Ma; e, 20 Ma; f, 25 Ma; and g, 30 Ma. Comparative track of each shows the way we model strain partitioning for the last 30 Ma in the southwest U.S. Cordillera.

B. Graphic summary of displacement data and interpretations for selected large strike-slip faults, rotated blocks, and extended areas in southwestern United States and coastal Mexico. References and text summary of each, accompanied by appropriate references, can be found in Table 1. 1. Northern part of San Andreas fault—displacement shown by open circles. 2. Southern part of San Andreas fault—displacement shown by open five-point star. 3. San Jacinto fault—displacement shown by filled seven-point star. 4. Elsinore fault—displacement shown by open seven-point star. 5. Newport-Inglewood fault zone—transport of Poway clasts from source marked by filled arrow. 6. San Clemente fault—location of offset Poway clasts shown by filled arrow. 7. Santa Rosa/Cortes Ridge fault. 8. East Santa Cruz Basin fault. 9. Western part of Transverse Ranges—rotated clockwise about 105 . 10. San Gabriel fault—displacement shown by open arrowhead. 11. Rinconada/Huasna/Pine Mountain fault zone—displacement shown by star-within-circle. 12. Hosgri fault offshore and faults of Santa Maria River/Santa Ynez River areas on land. 13. Santa Lucia Bank fault. 14. Sur/San Gregorio fault zone—displacement shown by filled arrowhead. 15. Maacama/Bennet Valley/Hayward/Calavaras fault zones. 16. Garlock fault—displacement shown by small black arrows. 17. Panamint Valley/Hunter Mountain faults. 18. Death Valley/Granite Mountain faults. 19. Furnace Creek fault zone. 20. Pahrump fault. 21. Las Vegas Valley shear zone—displacement shown by lines. 22. Lake Mead fault system—displacement shown by gray ball and bar. 23. San Benito fault. 24. Tosco/Abreojos fault. 25. La Paz fault. 26. Southern borderland rift.

wood fault zone, are juxtaposed whereas the western Transverse Ranges back-rotate into position behind the translated blocks.

The amount and timing of upper crustal displacement across part of the Basin and Range structural province has been defined by the synthesis of Wernicke et al. (1988). Their range-to-range reconstructions, which are based on correlations of pre-Neogene features, cross the southern part of the Great Basin from the stable Colorado Plateaus Province to the southern part of the Sierra Nevada Range, and they provide a "high-end" estimate for both the amount of extension and the right-lateral translation

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TABLE 1. SUMMARY OF DISPLACEMENT HISTORY FOR MAJOR FAULTS OF SOUTHWESTERN CORDILLERA

Number	Fault, fault segment, or region	Magnitude and sense of displacement	Displacement history followed in reconstructions in Figure 15	References*
1	San Andreas fault (north segment)	330 km right slip	240 km displacement since 5 Ma 40 km during period 5–22 Ma 50 km during period 22–24 Ma No displacement 24–30 Ma	Hill and Dibblee,1953; Addicott, 1968; Huffman, 1972; Matthews, 1976; Howell, 1975; Crowell, 1979; Nilsen, 1984; Ross, 1984; **Stanley, 1987**
2	San Andreas fault (south segment)	270 km right slip	Total displacement since late Miocene 90 km between 11 Ma and late Miocene Mostly inactive between 5 and 11 Ma 160 km since 5 Ma	**Crowell, 1962, 1982; Powell, 1981, 1993; Stewart and Crowell, 1992**
3	San Jacinto fault	30 km right slip	Total displacement since late Miocene	Crowell and Ramirez, 1979; **Sharp, 1967**
4	Elsinore fault	40 km right slip	Total displacement in last 5 Ma	Crowell and Ramirez, 1979; Stewart and Crowell, 1992
5	Newport-Englewood, Rose Canyon fault zone	Magnitude not known	Youthful strike-slip fault Mapped trace marks position of deeper and older detachment fault system upon which Catalina Schist was exhumed in early Miocene	Legg, 1985, 1991; Crouch and Suppe, 1993
6	San Clemente fault	Magnitude not known, right-normal	Responsible for some of the northwest displacement of Nicolas-foreland terrane primarily in the last 15 m.y.	Legg, 1985, 1991
7	Santa Rosa–Cortes Ridge (Ferrelo) fault	Total slip not known, probably right slip	Marks western edge of Nicolas-foreland terrane	Howell and others, 1974; Howell and others, 1987
8	East Santa Cruz Basin fault	Large-magnitude, right-normal	Primary breakaway zone of master detachment fault system (Nicolas detachment) beneath Nicolas-foreland terrane	Howell and others, 1974; Crouch and Suppe, 1993; Howell and Vedder, 1981
9	Western Transverse Ranges Malibu Coast/Santa Cruz and Santa Ynez faults	Up to 95° clockwise rotation between Santa Ynez fault and Malibu Coast/Santa Cruz fault	90°–95° clockwise rotation between 14 and 29 Ma, 60°–75° clockwise between 11 and 14 Ma, and 20°–45° clockwise between 5 and 11 Ma in Santa Ynez Range (western part of block). 45°–75° clockwise rotation between 14 and 29 Ma, 70° clockwise between 11 and 14 Ma, and 45° clockwise between 5 and 11 Ma in Santa Barbara region (central part of block). 35°–70° clockwise rotation between 14 and 29 Ma, 35°–50° clockwise between 11 and 14 Ma, and no rotation between 5 and 11 Ma in Ventura Basin region (eastern part of block).	Luyendyk and others, 1980; Luyendyk and others, 1985; Hornafius and others, 1986; Luyendyk, 1991
10	San Gabriel fault	60 km right slip	60 km of displacement between 5 and 11 Ma, active strand of San Andreas fault during this time	Crowell, 1962, 1982
11	Rinconada-Huasna fault zone	60 km right slip	60 km since early Tertiary 18 km since lower Pliocene	Ross, 1984; Sylvester and Crowell, 1992
12	Hosgri fault	115 km right slip	70–115 km of displacement during Miocene Since late Miocene may be reverse fault	Hall, 1975; Graham and Dickinson, 1978; McCulloch, 1989; Crouch and others, 1984
13	Santa Lucia Bank fault	Not known, probably right slip	Right slip during Miocene	McCulloch, 1989
14	Sur–San Gregorio fault zone	115 km right slip	115–150 km of displacement since late Miocene	Ross, 1984; Clark and others, 1984
15	Maacama–Bennet Valley–Hayward–Calaveras fault zone	Magnitude not known, right slip	Most of displacement is Pliocene and younger	Aydin and Page, 1984
16	Garlock fault	48–64 km left slip	Most of slip probably in last 15 m.y., since fault is viewed as intracontinental transform and extension to north is of that age. Slip history not restored on figure	Smith, 1962; Smith and Ketner, 1970; Davis and Burchfiel, 1973; McKenna and Hodges, 1990
17	Panamint Valley–Hunter Mountain fault	8–10 km right-oblique slip	Slip probably late Miocene Slip history not restored on figure	Burchfiel and others, 1987
18	Death Valley–Granite Mountain fault	8–35 km right slip	8 km based on Precambrian stratigraphic trends, 20–35 km of late Cenozoic slip based on provenance studies Slip history not restored on figure	Butler and others, 1988; Stewart and Crowell, 1992
19	Furnace Creek fault zone	40–100 km right slip	Maximum slip, all late Cenozoic, noted only on north end of fault; south end has little offset Extension in Death Valley region thought to be greater in north Slip history not restored on figure	Wernicke and others, 1988; Stewart, 1983

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TABLE 1. *(Continued).*

Number	Fault, fault segment, or region	Magnitude and sense of displacement	Displacement history followed in reconstructions in Figure 15	References*
20	Pahrump fault	16–19 km right slip	Total slip is late Cenozoic in age Slip history not restored on figure	Ligget and Childs, 1977
21	Las Vegas Valley shear zone	40–67 km right slip	Total slip is late Cenozoic in age Fault is primarily result of more extension to north than to south Slip history not restored on figure	Stewart and others, 1968
22	Lake Mead fault system	50–65 km left slip	Slip took place during period 10–15 Ma Slip history not restored on figure	Anderson, 1973; Bohannon, 1979, 1984
23	San Benito fault	Total slip not known, probably right slip	Slip may have occurred during period from 5.5 to 14 Ma	Spencer and Normark, 1989
24	Tosco-Abreojos fault zone	Total slip not known, probably right slip	Slip may have occurred during period from 5.5 to 14 Ma	Spencer and Normark, 1989
25	La Paz fault	Total slip not known, probably left-normal	Normal activity uplifting southern Cabo province prior to Miocene. After about 19 Ma, east-down, left-oblique motion indicated	Hausback, 1984
26	Southern Borderland Rift	Possible 150 km of north-northwest divergence	Inferred extension and possible sea-floor spreading took place in early to middle Miocene time	Crouch and Suppe, 1993

*Primary references are in bold print.

that is possible across this zone. The southeast limit of the displacement track of the east edge of the Sierra Nevada Range that is shown in Figure 3A is from Wernicke et al. (1988, Fig. 13). The displacement history is consistent with west-southwest extension and strike-slip faulting during the period 17–10 Ma in the southeastern part of the Great Basin (Anderson, 1971, 1973; Anderson et al., 1972; Bohannon, 1979, 1984; Wernicke et al., 1985) and west-northwest displacement during the period 10–0 Ma in the southwestern part (Hamilton, 1987; Cemen et al., 1985; Wernicke et al., 1986; Burchfiel et al., 1987; Butler et al., 1988). Wernicke et al. (1988) accounted for all known displacements between the Colorado Plateaus and the Sierra Nevada, and their analysis precludes a large amount of south-to-north displacement through that region. An assumption of distributed right shear that might result in a large north displacement of the Sierras is common to many regional models (e.g., Atwater, 1970), but it is absent in ours.

The extensional histories of the Mojave Desert, the southern Basin and Range province, and the northern Great Basin have been studied by many workers (e.g., Axen et al., 1993; Glazner and Bartley, 1984; Wernicke, 1992; Dokka, 1989; Spencer and Reynolds, 1989; Gans et al., 1989), but there is no synthesis that crosses the entire extended zone in any of these regions. The amount of extension that we show is limited primarily by the longitudinal position of the Pacific and Farallon/Vancouver spreading ridge or the position of the east edge of the Pacific plate. In the case of the northern Great Basin, our estimate of extension is slightly less than that of Hamilton and Myers (1966).

HISTORY OF SEA-FLOOR SPREADING

Magnetic anomaly data (Atwater and Severinghaus, 1989) and the global-circuit model (Stock and Molnar, 1988) were used to construct a plot of the longitudinal drift of different segments of the active spreading-ridge system, as well as extinct spreading ridges that become locked to plates (Fig. 4). Several patterns emerge. The locus of spreading at all latitudes migrated slowly eastward prior to Pacific/North American contact, yielding positive slopes in Figure 4. Following cessation of spreading, the extinct ridges and the last-formed magnetic anomalies immediately assume the steep-to-intermediate negative slope exhibited by the Pacific plate. Pacific/Juan de Fuca spreading, north of the Mendocino fracture zone, and Pacific/Cocos spreading, south of the Molokai/Shirley fracture zone system, each follow subparallel, steep, positive slopes until the Blanco fracture zone was established in the north at about 10 Ma or spreading ceased near the Molokai/Shirley system at about 12.5 Ma. The initial contact zone, between the Mendocino and Farallon fracture zones (exemplified by the "south of Mendocino" track in Fig. 4), is characterized by spreading that drifted east, died upon contact, and abruptly assumed Pacific-west displacement. Between the Farallon and Morro fracture zones (exemplified by the "Davidson" track in Fig. 4), the locus of spreading approached North America prior to 30 Ma then abruptly assumed a rate of west drift exactly half that of the Pacific plate, until spreading ceased at 19 Ma. South of the Murray fracture zone the east-drifting locus was replaced at 25.5 Ma by one with little or no relative longitudinal change until spreading ceased at 17–18 Ma.

Globally Oriented vs. Locally Oriented Spreading Systems

We use the term *globally oriented spreading* to describe spreading centers in the eastern Pacific Ocean whose orientations and rates were controlled by the angular divergence of plates driven by the strong slab-pull and gravitational gliding forces associated with Pacific rim subduction with little or no local interference by North America. The circum-Pacific subduction interface kept the system of plates in the Pacific Ocean isolated from the North American plate during

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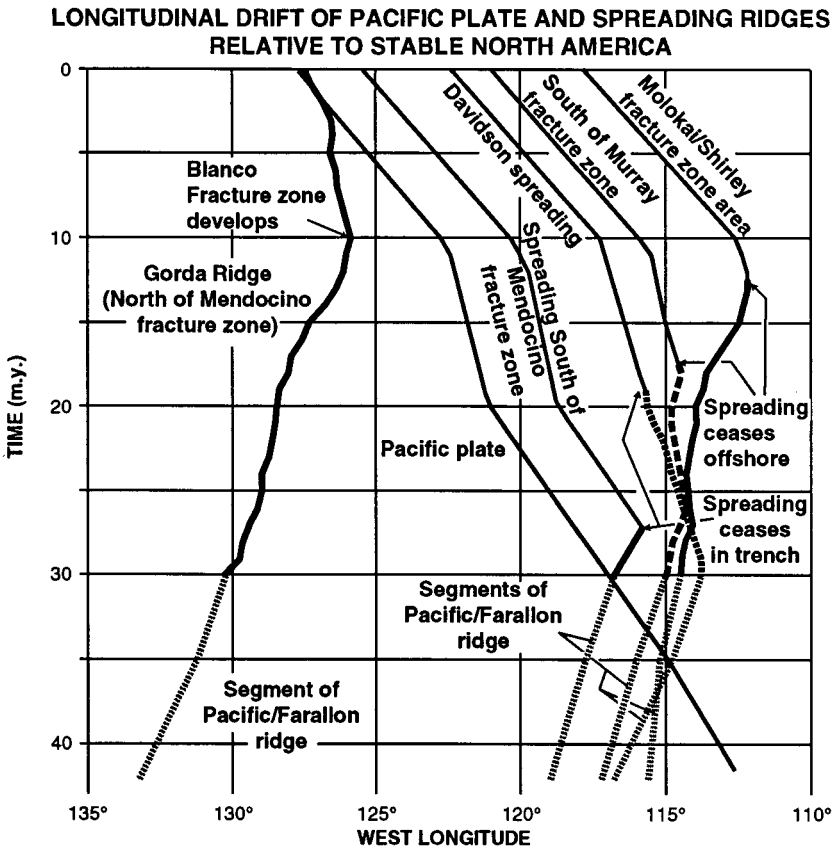

Figure 4. Plot of longitudinal position of selected features related to loci of sea-floor spreading in eastern Pacific Ocean with time. Positive slopes indicate convergence with North America and negative slopes indicate divergence; scale is arbitrary so steepness of slope is not meaningful. For times after 30 Ma, longitude of each feature was noted in North American reference frame at 1 m.y. intervals from a set of drawings similar to those in Figure 5. For times prior to 30 Ma (dotted heavy lines) longitude of features were taken from figures in Stock and Molnar (1988). Heavy lines, both dashed and solid, represent position of active spreading ridges. Medium lines represent position of features that are locked to the Pacific plate. Gorda Ridge track is north-side intersection of ridge and Mendocino fracture zone. South of Mendocino track is south-side intersection of ridge and Mendocino fracture zone. The Davidson track is south end of ridge intersection with Morro fracture zone. South of Murray track is from intersection of spreading south side of Murray fracture zone. Molokai/Shirley track is from intersection of spreading south side of Molokai and Shirley fracture zones. Pacific plate track is representative of any feature older than 30 Ma on plate.

the period 40–30 Ma. Thus, the position, orientation, and spreading rate of globally oriented loci of spreading were products of the instantaneous kinematics of the whole global plate system. Globally oriented systems are characterized by the east-migrating tracks of the loci of spreading in Figure 4 and by fracture zones with east-west to west-northwest trends, ridges oriented more-or-less north-south, and angular rates of divergence commonly >1.0°/m.y. (Stock and Molnar, 1988; Rosa and Molnar, 1988).

The Davidson motion track in Figure 4, prior to cessation of spreading at 19 Ma, typifies what we refer to as a "locally oriented spreading" system. The position, orientation, and spreading rate of these systems are directly controlled by local interaction between the Pacific and North American plates. The full spreading rate is approximately half that of the rate of divergence between the Pacific and North American plates, allowing the ridge to migrate away from North America at half the rate of the Pacific plate. A perfectly developed locally oriented spreading system depends on complete capture of the subducted plate by North America (no subduction) and no deformation in the continental lithosphere, so the ridge geometry is then the product of a simple Pacific and North American two-plate system. Because the ridge geometry is controlled entirely by the divergence of the two plates in a passive system, any slip at the subduction interface or deformation within the continent would cause a corresponding alteration of the spreading rate and reorientation of the ridge geometry such that neither would reflect ideal North American/Pacific relative motion. The ideal locally oriented spreading system would result in spreading ridges with northeast orientations, perpendicular to the Pacific–North American relative motion path of Stock and Molnar (1988), and fracture zones oriented northwest, parallel to the motion path.

There is evidence that the transition from globally to locally oriented spreading is an imperfect one. Motion tracks such as that of the spreading center south of the Murray fracture zone, prior to cessation of spreading at 12.5 Ma, and the Gorda Ridge track after development of the Blanco fracture zone at 10 Ma, seem to lie part way between the ideal global and local systems. This probably reflects imperfect frictional coupling along the subduction interface and deformation within Cordilleran North America.

KINEMATIC CONCERNS—DID NORTH AMERICA REALLY DO WINDOWS?

Dickinson and Snyder (1979), Zandt and Furlong (1982), Furlong et al. (1989), and Severinghaus and Atwater (1990) proposed that some type of a slab-free zone or "asthenospheric window" developed beneath North America following its contact with the Pacific plate. There are two main variants on the slab-free-zone theme. In one, the gap in the slab enlarges with time as the plate contact lengthens, the subducted lithosphere warms, and the slab sinks away (e.g., Severinghaus and Atwater, 1990). In the other, the size of the slab-gap is limited by conductive cooling and the formation of new lithospheric mantle material, but a northward-migrating active zone persists inland and slightly south of the Mendocino triple junction (e.g., Furlong et al., 1989). In either case, a large part of the western North American crust is suggested to be directly underlain by mantle that upwelled in Neogene time. The kinematic requirements for some type of slab-free zone are compelling, but the depth at which one might form is highly debatable. Many of the slab-free-

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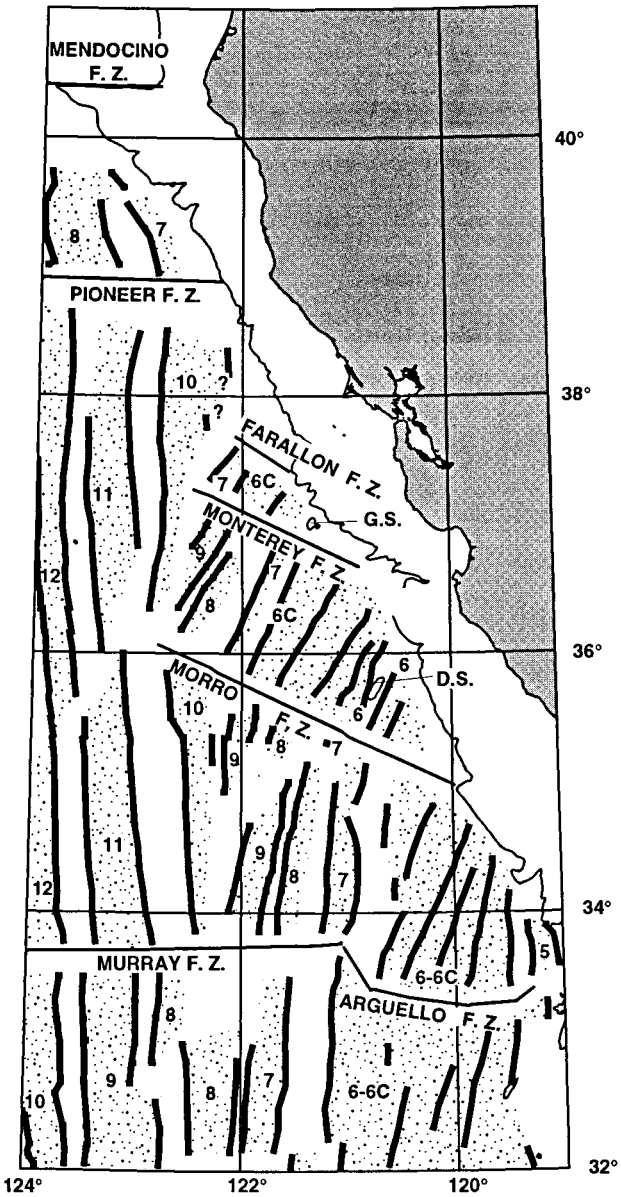

Figure 5. Magnetic anomalies and fracture zones that define the Monterey microplate system. Fracture zone (F.Z.) orientations are parallel to Pacific/North American motion direction provided by Stock and Molnar (1988) for 19–29 Ma time period. Heavy black lines are from actual picks of youngest and oldest positions of each anomaly provided in a digital data set associated with the magnetic anomaly map of Atwater and Severinghaus (1989). Shading simply outlines region within any given chron and does not necessarily indicate polarity. Davidson Seamount (D.S.) and Guide Seamount (G.S.) are shown. They are interpreted to be fossil spreading ridges that became extinct at about 19 and 20 Ma, respectively. Fragments of Farallon plate are preserved to the east of those seamounts (Lonsdale, 1991).

zone models endeavor to explain aspects of North American geology through the upwelling asthenosphere that results from a shallow gap in the slab. Shallow slab gaps are a natural consequence of the migrating-triple-junction paradigm and the assumption that the subducted Farallon slab sank freely at all levels. The documented areas of plate-capture show that this was not the case in some places (Lonsdale, 1991). We argue that it probably was not the case anywhere along the North American margin.

The problem of slab-free zones can be examined kinematically from the perspective of Pacific/North American relative motions. The magnetic anomalies between the Mendocino and Farallon fracture zones (Fig. 5) and the "south of Mendocino" motion track (Fig. 4) indicate that this segment of the Pacific and Farallon/Vancouver spreading ridge operated as a viable globally oriented spreading system until its contact with the continent, at which time it simply ceased to exist. After contact, the last-formed anomaly and the continental edge both appear to have assumed Pacific plate motion. South of the Farallon fracture zone (Fig. 5), during the same time period, a locally oriented spreading system was apparently established as indicated by the Davidson motion track in Figure 4. We present several possible kinematic models to describe what might have happened beneath North America in the critical time surrounding this plate contact.

The two kinematic models that we view as the most realistic are shown in Figure 6. We call them the stall-and-deform and local-ocean models. Two other possibilities, in which asthenosphere rises to shallow levels beneath the continent as a result of slab separation, are shown in Figure 7. In one of these end-member possibilities, upwelling is distributed and all flow is vertical. In the other, upward flow is localized with shallow lateral flow as the gap in the slab widens. A gap in the slab is kinematically required in all of the models. In the stall-and-deform and local-ocean models it forms well beneath normal lithospheric levels. The distributed-flow and localized-flow models are meant to illustrate some of the kinematic consequences of forming the gap at lithospheric levels.

Stall and Deform

The stall-and-deform model (Fig. 6) is characterized by a globally oriented spreading ridge that drifts east until it contacts the continental edge in the trench, where it stalls upon contact. The dead spreading ridge then drifts west with the Pacific plate. The shallow part of the formerly descending plate becomes linked to the Pacific plate and assumes Pacific relative motion (plate capture). The part of the continental lithosphere that is above and to the east of the oceanic slab stretches, widening the distance between the continental edge and the west limit of the coherent continental interior. This allows the continental edge to remain in contact with the last-formed anomaly on

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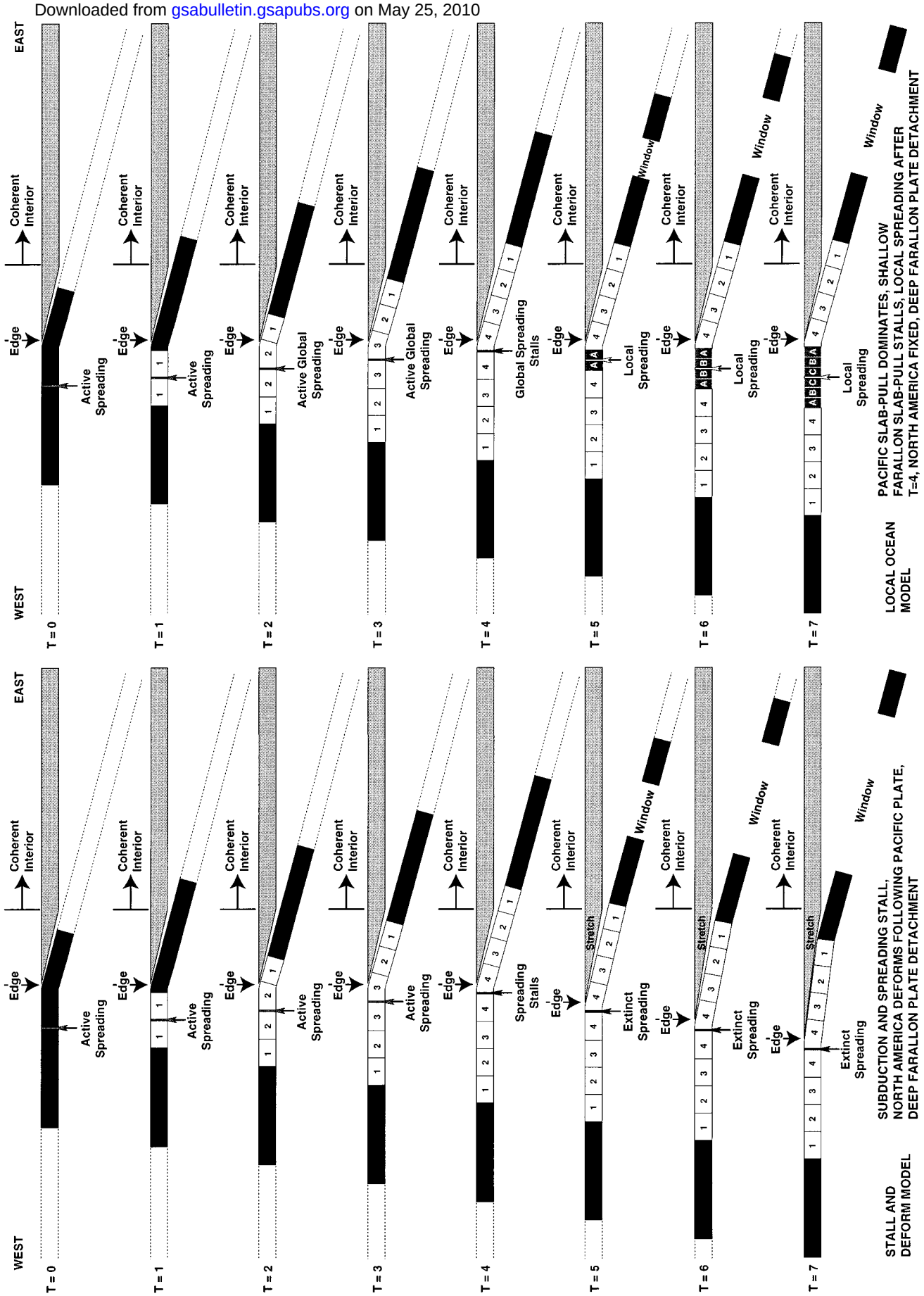

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Figure 6. Stall-and-deform and local-ocean models for encounter of globally oriented oceanic-spreading system with North American continental lithosphere. Shown at eight times (T = 0 to T = 7) separated by even increments. Pacific plate on west side of each diagram drifts to west at uniform rate. Coherent interior of North American plate, shown shaded on east side of each diagram, is fixed through examined time sequence. Black lithosphere on each oceanic plate represents an arbitrary section of lithosphere that formed immediately prior to T = 0. Oceanic lithosphere is formed at uniform rate on each plate, and sections that form during each time interval are shown by a corresponding number. Global-spreading half rate is sufficient to cause locus of spreading to drift east relative to continental edge. Convergence of subducting (Farallon) plate with North America occurs at a rate dictated by west-drift of Pacific plate and full globally oriented spreading rate.

Stall-and-deform: Once globally oriented spreading ridge encounters North American edge, it becomes extinct and immediately assumes west drift of Pacific plate in accordance with south-of-Mendocino motion track in Figure 4. Friction at subduction interface overcomes slab-pull force on part of Farallon plate in contact with North American lithosphere. "Stuck" segment of plate assumes Pacific west drift, taking overlying section of North American lithosphere with it and causing extension along western continental edge. Separation between west edge of continent and last-formed, globally oriented spreading anomaly on Pacific plate remains unchanged with time. Slab window forms at sublithospheric levels beneath North America in response to slab-pull force on part of descending plate not overcome by frictional resistance.

Local-ocean: Encounter of globally oriented spreading ridge results in development of locally oriented spreading system governed by interaction of Pacific and North American plates. Full local spreading rate, shown by shaded boxes of new oceanic lithosphere A through C that develop at T = 5 to T = 7, is dictated by Pacific/North American divergence rate. Friction at subduction interface overcomes slab-pull force on part of Farallon plate in contact with North American lithosphere. "Stuck" segment of plate assumes North American motion. Separation between west edge of continent and last-formed, globally oriented spreading anomaly on Pacific plate increases at full local spreading rate with time. No deformation of continental lithosphere is required. Slab window forms at sublithospheric levels beneath North America in response to slab-pull force on part of descending plate not overcome by frictional resistance.

the Pacific plate. The gap in the descending slab forms well beneath the lithosphere, where it has little or no impact on the overlying continental lithosphere.

The purely kinematic view leaves some things unexplained. It offers no physical explanation for the plate capture event once the spreading ridge is extinguished. Furthermore, it does not treat the force-balance problem between friction at the subduction interface and the body forces acting on the descending slab that is necessary in order to understand the continental deformation. Last, there is no attempt to deal with the thickness, thermal-life, or strength and coherence of the descended plate. These aspects of the problem will be treated in the section on thermal and mechanical considerations.

Local Ocean

The local-ocean model begins, as the stall-and-deform model does, with the east drift of a globally oriented spreading ridge and its extinction upon contact with the continent. In this case, globally oriented spreading is immediately replaced by locally oriented spreading whose locus drifts west at half the rate of the Pacific plate (Fig. 6). In this model, the Pacific and North American plate divergence is precisely compensated by the locally oriented spreading system, so there can be no further shallow subduction and no continental deformation is required. Continued Pacific and Farallon/Vancouver plate divergence is accounted for by the slab window, which develops well beneath the lithosphere, where it has little or no effect on the overlying continent. The part of the descended slab that is upslope of the window is captured by North America, with which it moves. This model requires relatively little additional dynamic explanation because no continental deformation occurs and the Pacific/North American divergence is compensated by normal oceanic spreading. However, it does not explain why the shallow part of the subducted slab becomes linked to North America, rather than descending farther. The problems of thickness, life, strength, and integrity of the deep slab are left untreated by the simple kinematic approach.

Distributed Upwelling

A shallow slab-gap might have formed if the subduction process continued after the contact of the globally oriented spreading ridge with the continent (Dickinson and Snyder, 1979; Severinghaus and Atwater, 1990). In most slab-gap models, the space between the plates is filled by upwelling asthenosphere without the creation of an overlying conducting lid, a mechanical boundary layer and coherent plate, or a compositionally dissimilar layer. We consider here an end-member case of that model in which the upwelling is uniformly distributed between the divergent plates such that all flow is vertical (Fig. 7). In this case, the subducted slab descends uniformly beneath the continent, presumably because friction and buoyancy are insufficient to overcome the collective slab-pull forces and the slab has enough internal strength to remain intact. The main problem with this model is that the Pacific and North American plate divergence causes a widening zone to emerge at the Earth's surface between the continental edge and the last magnetic anomaly formed by normal oceanic spreading (Fig. 7).

How might the Pacific and North American divergence problem be avoided? One way would be to allow North America to deform to cover the hole. But what causes the extension? Since upwelling is vertical and friction at the base of the continental lithosphere is low enough to allow the slab to fall away, what provides the connection between the continental edge and the west-moving Pacific plate? The simple kinematic model that we present is incapable, by itself, of providing any continental extension.

Localized Upwelling

The localized upwelling end member of the shallow slab-window model (Fig. 7) has many characteristics in common with the distributed upwelling model. There is a similar problem created by the Pacific and North American plate divergence. It might be argued, in this case, that continental extension might result from tractional forces set up by the lateral flow of asthenosphere.

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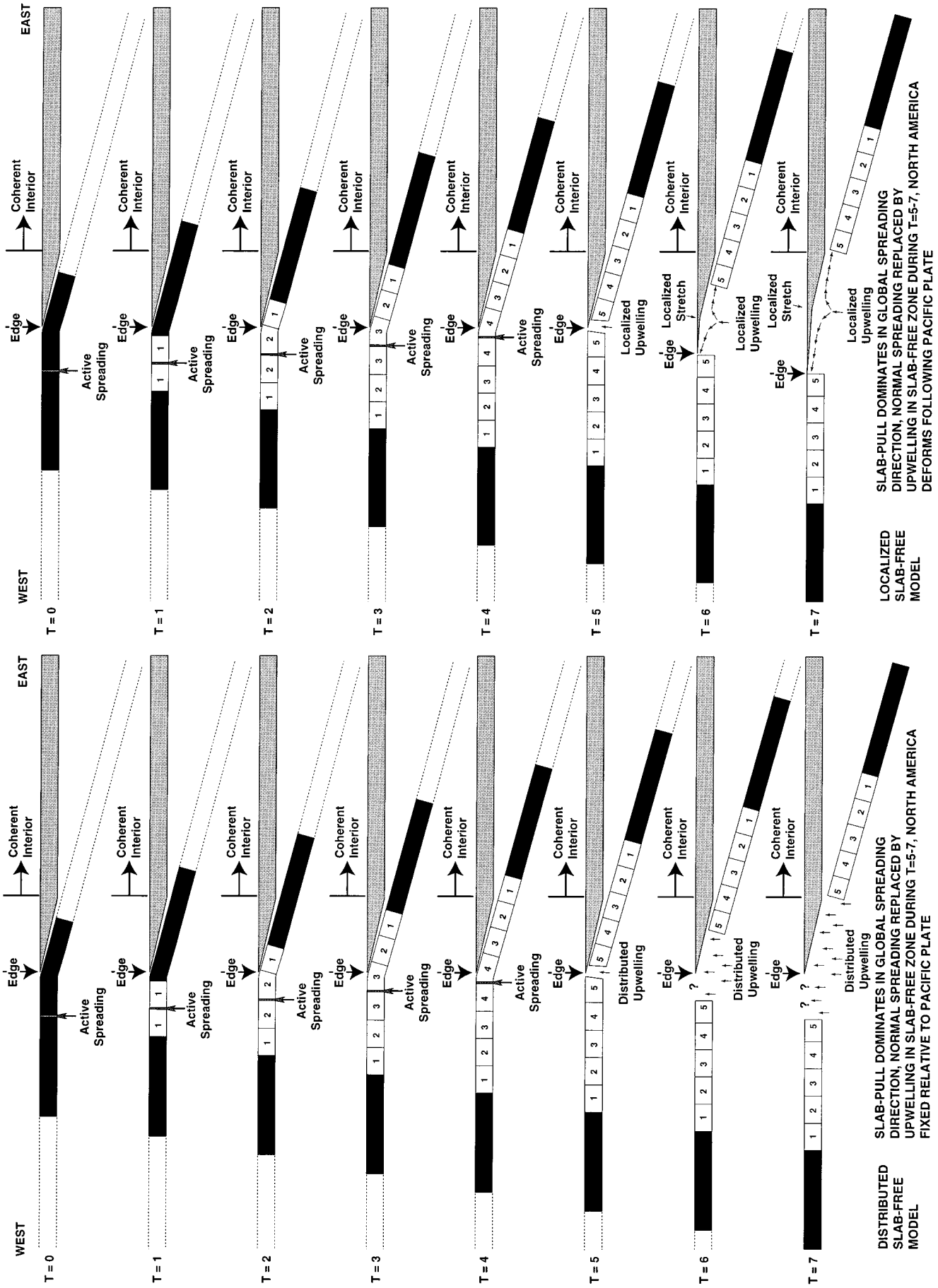

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Figure 7. Two end-member models that illustrate kinematic consequences of developing slab-window at lithospheric levels. Widening gap between oceanic plates results from strong slab-pull forces that are not overcome by frictional resistance with North American lithosphere at subduction interface. Asthenosphere is assumed to fill gap between plates by upwelling without production of conducting lid, mechanical boundary layer, or compositionally dissimilar layer (crustal layer formed by underplating of frozen partial melt) below subduction interface. Upwelling is assumed to be distributed evenly throughout the widening gap between the diverging plates with all flow being vertical in one end-member case. Upwelling is assumed to be highly localized, maintaining a uniform distance halfway between the diverging plates, with the widening gap being filled by lateral asthenospheric flow in the other case. Kinematic development and graphic elements are similar to those in Figure 6.

Distributed upwelling: Once globally oriented spreading ridge encounters North American edge, it gives way to uniformly distributed asthenospheric upwelling. Divergence between Pacific and North American plates causes a widening gap between the continental edge and the last-formed magnetic anomaly on the Pacific plate after the ridge has encountered the continent. The kinematics of this model do not suggest a suitable force that might cause the deformation within the continental lithosphere that would be required in order to "cover the hole." This is because friction at the subduction interface is low enough to allow the cold descending slab to fall completely away and all asthenospheric flow is vertical.

Localized upwelling: Encounter of globally oriented spreading ridge results in continued localized asthenospheric upwelling, but no lithosphere is formed. "Reverse subduction" problem might be avoided by allowing continental edge to deform above gap between diverging Pacific and North American (stable part) plates. This must occur by traction at the base of the continental lithosphere caused by lateral flow within the asthenospheric cell. As such, friction would have a greater influence on the asthenosphere/continent system than it did on the plate/continent system that fell away. A localized zone of extension would migrate inland with time in the continental lithosphere above the spreading locus in the asthenosphere.

However, the model requires the slab, which is strong and cool, to fall away presumably because friction and buoyancy at the subduction interface are insufficient to hold it in place. The hot and fluid asthenosphere that replaces the slab is then required to exert tractional force across the same interface. The result is a localized zone of continental stretching that migrates inland with time. The latter phenomenon is not consistent with the established pattern of Neogene tectonism in California (e.g., Sedlock and Hamilton, 1991).

Implications and Shortcomings

There are several important implications of the stall-and-deform and local-ocean models for Cordilleran deformation. Each model requires a strong coupling at the interface between the descended slab and the overlying continental rocks. The subducted slab and the Pacific plate are mechanically linked in the stall-and-deform model and the subducted slab and the North American plate are linked in the local-ocean model. The stall-and-deform model provides an ideal mechanism for deforming the part of the continental plate above the stalled subduction interface. The local-ocean model eliminates the need for extension in the overlying continental lithosphere. Thus, Cordilleran extension should occur inland of Pacific capture zones, whereas adjacent regions inland of continentally captured plate should remain undeformed. We think this is actually the case (see Discussion section).

There are several shortcomings of the stall-and-deform and local-ocean kinematic models. Neither addresses the actual plate-capture process very well. The stall-and-deform model requires an abrupt change in the relative motion of the subducted plate with respect to North America once the Pacific captures the latter plate, yet the model offers no physical explanation for this. The continental extension, following plate capture in the stall-and-deform model, requires that the stalled spreading ridge assume a greater strength profile than that of the deforming continent. Both models require a cessation of shallow subduction, accompanied by some type of separation of the deep and shallow parts of the subducted plate for reasons that are not apparent. These problems are addressed in the section on thermal and mechanical considerations.

The chief drawback of the two shallow-slab-gap models is that they are not applicable along the 65% of the margin where Lonsdale (1991) has documented extinct spreading ridges offshore. They might be applicable elsewhere, but neither offers a good internal mechanism for keeping the continental edge in contact with the diverging Pacific plate. A possible way to overcome this is to assume some type of specialized precondition in the Cordilleran continental lithosphere that would enable it to extend independently. Coney and Harms (1984), Hamilton (1987), Sonder et al. (1987), Glazner and Bartley (1984), among others, speculated that Cordilleran extension resulted from the gravitational spreading of unstable crust. An unstable continent might spread over the widening zone of mantle upwelling in either of the shallow-slab-gap models. In such a case, continental extension would be controlled by the plate boundary configuration and kinematics, not driven by them. We discuss some problems with this approach in the Discussion section.

THERMAL AND MECHANICAL CONSIDERATIONS

In this section we explain many of our reasons for concluding that the stall-and-deform and local-ocean models offer the best description of the present continental margin of the western Cordillera. Both models rely on the cessation of subduction at shallow levels once very young oceanic lithosphere enters the subduction zone, so we present some relevant force-balance arguments. The gap in the slab develops well below lithospheric levels in each model, and we explain this based on thermal modeling. Finally, strength profiles are examined for an extinct spreading ridge and a cross section across part of the margin in order to substantiate our claim that plate capture is a viable means of explaining continental deformation.

Force Balance

Turcotte and Schubert (1982) summarized the driving forces associated with subduction and ridge elevation as:

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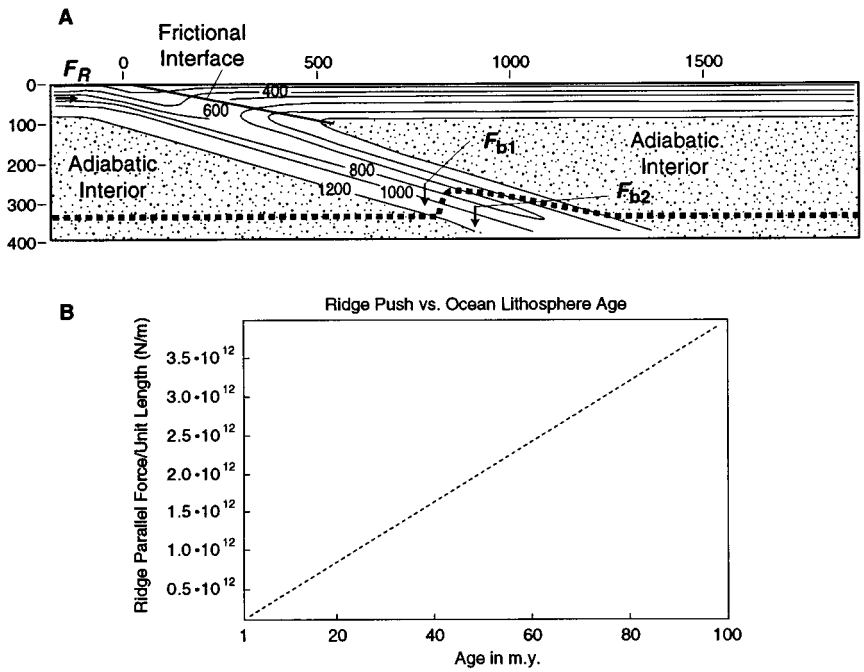

Figure 8. Summary of positive forces acting on subducting plates through the frictional zone along the subduction interface: $F_{b1} + F_{b2} + F_R$.

A. F_{b1} is a strong force related to the vertical thermal boundary layer associated with the thermal deficit of the cool subducting slab. There is also a force due to the elevation of the olivine-spinel phase change (F_{b2}) that is graphically represented by the heavy dashed line at about the level of the 1600 C isotherm in the adiabatic interior. It is elevated in the cooler descending slab. Turcotte and Schubert (1982) reasoned that this force might total 1.6×10^{13} N/m measured along the trench length, but the strength of this force at the frictional interface is probably minimal except in the oldest slabs because of the depth at which it is generated.

B. Plot of ridge-push or gravitational-glide force (F_R) versus age of oceanic lithosphere. The ridge-push gravitational glide is a body force acting on the lithosphere that results from the pressure head caused by the elevational difference at the base of the lithosphere between oceanic ridges and trenches (Turcotte and Schubert, 1982). The ridge push force (F_R) is linearly proportional to the age of the lithosphere in millions of years for most oceanic lithosphere, and it is measured in newtons per meter parallel to ridge length. It abruptly goes to zero once ridge enters trench.

$$F = F_{b1}\,(\text{thermal deficit}) + F_{b2}(\text{Ol/Spl transition}) + F_R\,(\text{ridge push})$$

(Fig. 8). F_{b1} is a strong force related to the vertical thermal boundary layer associated with the thermal deficit of the cool subducting slab. Turcotte and Schubert (1982) reasoned that it is effective to at least 700 km depth and that it probably averages in the range of 3.3×10^{13} N/m, measured parallel to the trench length. F_{b1} is linearly proportional to depth, and the percentage of the total amount of F_{b1} available to overcome retarding forces acting along the subduction interface depends on a measurement of the depth to the 700 °C isotherm below which dunite might support stress of 1 GPa or more. There is also a force due to the elevation of the olivine-spinel phase change (F_{b2}) that is graphically represented by the heavy dashed line at about the level of the 1600 °C isotherm in the adiabatic interior in Figure 8. It is elevated in the cooler descending slab. Turcotte and Schubert (1982) reasoned that this force might total 1.6×10^{13} N/m measured along the trench length, but the strength of this force at the frictional interface is probably minimal except in the oldest slabs because of the depth at which it is generated. The ridge push is a body force acting on the lithosphere that results from the pressure head caused by the elevation difference along the base of the lithosphere between oceanic ridges and trenches (Turcotte and Schubert, 1982). The ridge push force (F_R) is linearly proportional to the age of the lithosphere in millions of years for most oceanic lithosphere. Ridge push is measured parallel to ridge length, and it is commonly an order of magnitude less than the other positive driving forces (Fig. 8). The ridge push force acting on the subducted slab abruptly goes to zero once the ridge enters the trench.

Forces of retardation include the viscous resistance to flow deep in the mantle, positive buoyancy forces (where present), friction along the subduction interface and along transform faults, and tractional drag at the base of the lithosphere. The stall-and-deform and local-ocean models depend on friction along the subduction interface and positive buoyancy associated with young oceanic lithosphere locally overcoming the other forces that promote the subduction of a small plate.

The forces that promote the subduction of the shallow part of a young plate are reduced in several ways. The large forces associated with the thermal deficit and the elevation of the olivine to spinel phase change are severely reduced in young subducting lithosphere. This is primarily because the 700° isotherm is elevated in young slabs and the strong, deeply generated pulling forces cannot be transmitted to the surface. An additional reduction in the thermal deficit arises from the fact that young oceanic lithosphere has a slight positive buoyancy relative to its underlying fluid mantle. The observed relations of sea floor elevation as a function of age indicate that it takes 8–10 m.y. for normal sea-floor elevation to drop below 3.5 km depth (Parsons and Sclater, 1977), which is the calculated depth of the hypothetical free surface of the fluid mantle (Morgan and Burke, 1985; Lachenbruch and Morgan, 1990). The ridge-push body force acting on the subducted slab is diminished or disappears altogether. All this leads us to the conclusion that the downward motion through the subduction interface is severely reduced or eliminated in small, young plates. Frictional drag per unit length of the subduction interface should not decrease significantly as a function of slab age. Thus, we think it is logical to conclude that friction and buoyancy are able to overcome the forces that promote the subduction of the shallow part of a young slab.

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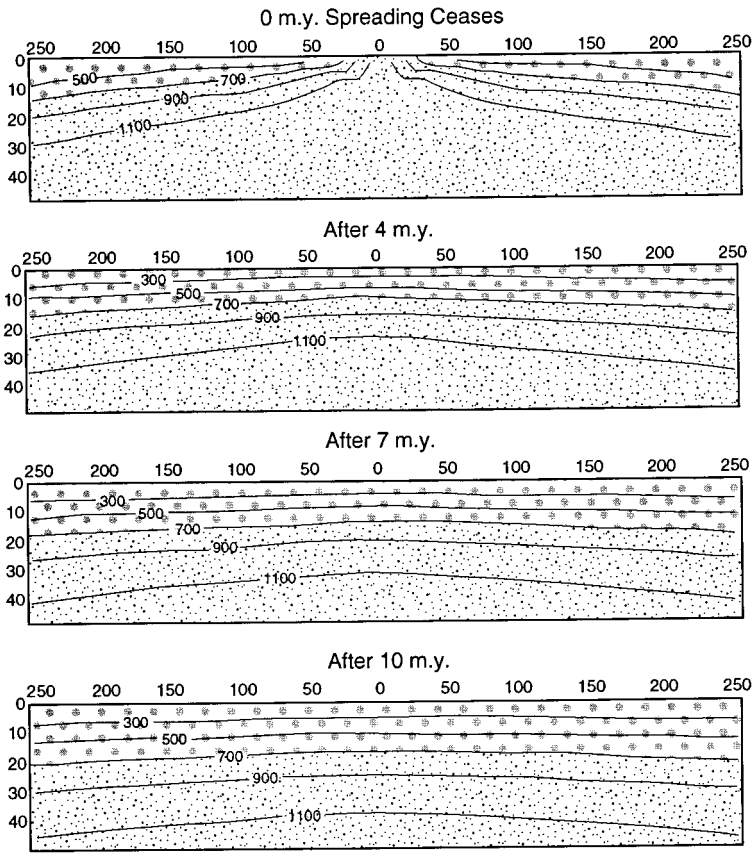

Figure 9. Thermal history of an extinct spreading ridge over a time interval of 10 m.y. After 4 m.y. the spreading center cools to the extent that a strong mechanical mantle layer develops above the 700–800 C contour. It is likely that even after a short time interval (≤4 m.y.) the extinct spreading ridge is stronger than continental crust. Details on the thermal modeling methods and results are given in the text.

Thermal Model

We use two-dimensional thermal modeling to make predictions of where the leading edge of integrity in the subducting Farallon slab was at various times during the Tertiary period, and on the thermal history of a spreading ridge after spreading ceases. A numerical finite difference solution is applied to solve the two-dimensional thermal-conduction equation

$$C\rho(\partial T/\partial t) = \nabla \quad (K\nabla T) + Q$$

where C is specific heat, ρ is density, T is temperature, t is time, K is thermal conductivity, and Q is a combination of internal heat production and frictional heat generated by the descending slab when applicable. Variable heat capacity and conductivity with depth and composition are considered after summaries by Saxena and Shien (1992) and Durham et al. (1987). The frictional heating generated by subduction is approximated using the relation for constant shear-stress heating after Dumitru (1991) and a contact shear stress of 6 MPa after Bird (1984). We apply an iterative technique to establish heat production values for continental crust, continental lithospheric mantle, oceanic crust, and asthenosphere that generates long-term steady-state thermal conditions throughout a test model. The alternating direction implicit method of Douglas (1961) is used to solve the conduction equation over finite time steps. To develop the thermal history of a dying spreading ridge, we apply a static model on an initial temperature field after Denlinger (1992). To insure a stable numeric solution, the thermal calculations are incremented by 10 000 yr time steps. The models are calculated over a far greater width and depth than the region of interest to avoid side and bottom boundary effects.

Thermal Evolution of Stalled Spreading Centers and Subducted Slabs

We treat the thermal evolution of a stalled spreading center precisely like that of normal aging oceanic lithosphere without worrying about the dynamical reasons for it stalling. We predict that the 700 °C isotherm drops below 10 km within 4 m.y. after spreading ceases. Thus, a thick mechanically coherent layer probably develops within 1 or 2 m.y. at the site of the extinct ridge crest (Fig. 9).

The dynamic thermal model of the subducting margin is tailored to conditions off western North America. To establish proper boundary conditions, the model is initiated at 90 Ma, when the North American plate boundary was characterized by the rapid subduction of 150 Ma oceanic lithosphere (Engebretson et al., 1985). An oceanic plate with an initial thermal structure that incorporates a variable thermal diffusivity (Denlinger, 1992) slides incrementally beneath a 100-km-thick North American conducting lid in 1 m.y. time steps. The thermal calculations are incremented by 10 000 yr time steps for numerical stability. Subduction is allowed to progress at a dip of 15° over an 85 m.y. time period (from 90 Ma to 5 Ma) at slowing subduction rates that are averaged from calculations of Engebretson et al. (1985) and Stock and Molnar (1988) to establish the thermal structure at the latitude of Los Angeles during the Tertiary period.

Our models show the slab isotherms receding up the slab to the west due to the reduction in age of the lithosphere entering the trench from 43 to 30 Ma (Fig. 10). From 30 to 10 Ma, the isotherms in the slab south of the Mendocino fracture zone continued to recede back toward the trench after Pacific/North American plate contact, causing the coherent leading edge of the slab to lie only about 400 km east of the trench by 20 Ma (Fig. 10).

A 15° dip is retained for the slab through the modeling period. Teleseismic studies suggest that the present-day dip of the Gorda and Juan de Fuca slabs is about 15° (Nabelek et al., 1993). An east-to-west rollback in volcanism occurred between about 42 and 15 Ma that has been attributed to a steepening of the subducting slab (Coney and Reynolds, 1977; Burke and McKee, 1979; Armstrong and Ward, 1991). At about 43 Ma, there was a slowing in convergence rate (Engebretson et al., 1985; Stock and Molnar, 1988) that might have initiated the change in volcanic patterns. The westward-

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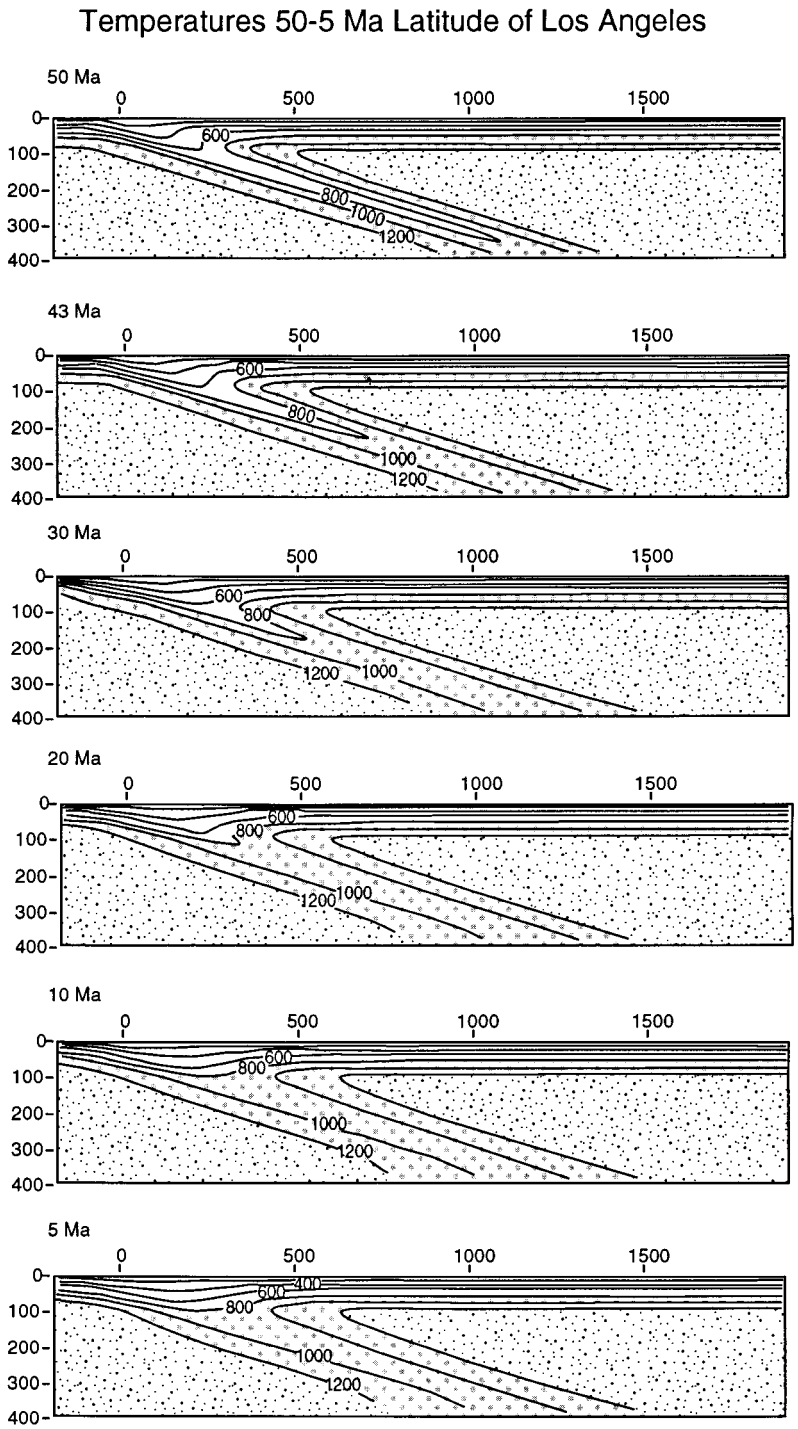

Figure 10. Thermal history of a subducting slab dipping 15 beneath the continental margin at the latitude of Los Angeles, California, during the time interval of 50–5 Ma. The thermal profile of the slab at depth gradually grows warmer over the modeled time interval. This is due to the gradual slowing of subduction and to the subduction of younger (thinner) lithosphere with time. The cold, strong part of the slab recedes back toward the trench by thermal softening faster than new slab can be subducted. Thus a continual reduction in the local slab-pull force exerted by the slab occurs, eventually causing subduction to terminate. See text for details on the modeling method, parameters, and discussion of ramifications.

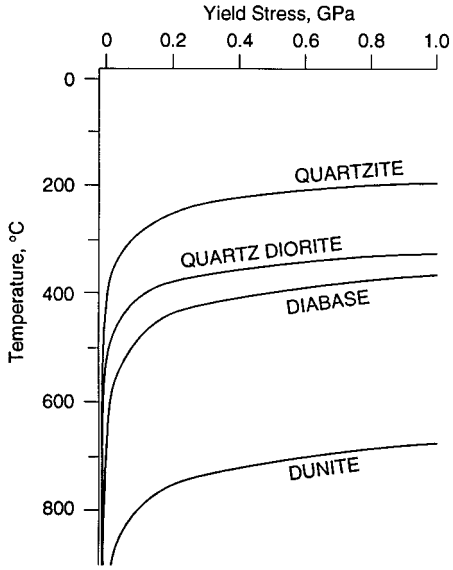

Figure 11. Predicted steady-state strength at $\varepsilon = 10^{-15}$ s^{-1} (steady-state creep rate) shown as differential stress versus temperature for nominally dry rocks. Taken from Kirby (1985, Fig. 5).

receding isotherm marking the coherent edge of the slab in our thermal model ties well with the observed rollback in volcanism until about 30 Ma. Thus it is possible that the slab never got any steeper and that the locus of volcanism simply occurred above the east edge of coherence in the slab. Subduction stopped at about 29 Ma, following Pacific/North American plate contact, at the latitude where the rollback has been documented. After that the locus of volcanism was probably more coincident with changing patterns of extensional tectonism than with a particular slab geometry or characteristic.

Strength Profiles

The rheologic behavior of rocks in the lithosphere is strongly dependent on their temperature and composition (Ranalli, 1987). Thermal models (e.g., Fig. 10) provide a predicted temperature distribution through time. Kirby (1985) studied the strength/temperature relations that characterize several types of crustal rocks and compared them to that of dunite. His findings are summarized in Figure 11. By combining the thermal model with a simple lithologically based cross section we can use the strength and temperature curves to predict where zones of weakness are apt to occur in a two-dimensional section of the lithosphere.

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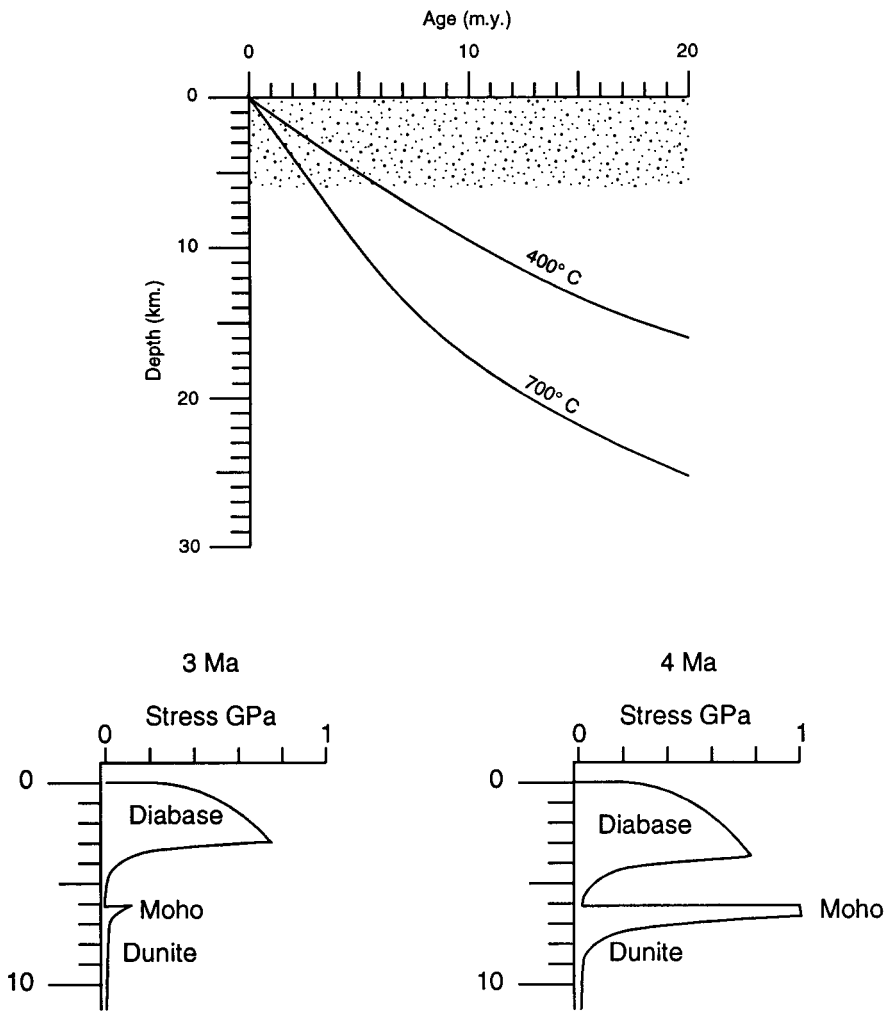

Figure 12. Depth to 400 and 700 C isotherms in oceanic lithosphere that is 0 to 20 m.y. old after Denlinger (1990). Crustal layer 6 km thick is shown by gray pattern. Strength versus depth plotted for 3- and 4-m.y.-old lithosphere based on strength/temperature curves of Kirby (1985) that are shown in Figure 13. Upper limits to strength curves are generalized from Ranalli (1987, Figs. 12.1 and 12.2).

Our initial concern is to establish the strength profile of a stalled spreading ridge, whose cooling pattern is treated herein as if it were normal oceanic lithosphere. Denlinger (1992) modeled the temperature structure of oceanic lithosphere based on a variable, rather than averaged, thermal diffusivity. Kirby's (1985) strength curves suggest that the 400° and 700° isotherms are of primary importance for oceanic lithosphere because of their control on the diabase and dunite rheologies, respectively. Figure 12 shows the depth to those isotherms versus age of the lithosphere and the strength profiles determined for lithosphere that is 3–4 m.y. old.

Young oceanic lithosphere is thin, but Figure 12 suggests that parts of it can be strong. Basalt and gabbro in the upper half of the crustal layer have a yield strength in the 0.6–0.8 GPa range even in oceanic lithosphere whose age is 3 m.y. old or younger (Fig. 12). At 4 m.y. old the strength of oceanic lithosphere increases markedly through the cooling of rocks of the upper mantle. There is a drop in strength just above the Moho in 4-m.y.-old oceanic lithosphere suggesting that the compositional discontinuity might also be a rheological boundary for a short time period. The 400° isotherm drops below 6 km depth after 6 m.y. (Fig. 12), which would eliminate the weak layer above the Moho. Most of the increase in strength after 6 m.y. results from thickening of the lithospheric mantle layer. We assume that the strength profile of a stalled spreading ridge would evolve in approximately the same way.

The 20 Ma thermal cross section from Figure 10 is used to predict the two-dimensional position of relatively strong and weak layers in the lithosphere from ocean to continental interior (Fig. 13). It is during this time that the kinematic models have subduction stalled along the interface and the Pacific plate is linked to the subducted slab. Also at this approximate latitude, upper crustal extension was well underway in the Colorado River extensional corridor and in the continental borderlands of California on both sides of the Peninsular Range batholith (Spencer and Reynolds, 1989; Axen et al., 1993; Legg, 1992; Crouch and Suppe, 1993). Clockwise rotation had also begun in the Transverse Ranges (Hornafius et al., 1986). Some heterogeneity in the continental crust is built into Figure 13 in an attempt to account for the different strength profiles associated with the coastal accretionary prism zone, the Peninsular Range batholith, and the regions inland of there.

The strongest zone in the two-dimensional slice of the lithosphere represented in Figure 13 lies in the uppermost mantle beneath the ocean and in the subducted slab. Even the oceanic crustal layer contributes to the strength of the slab as far inland as 150 km. The slab thickens and picks up strength down dip because of the slightly greater age when it entered the trench and the additional cooling time afforded to it. It continues as a strong zone to depths in excess of 80 km and as far inland as 250 km (Fig. 13). The Pacific plate cannot be distinguished from the subducted slab in either the temperature or strength profiles, so the link between them is strong.

The strongest crustal profiles are at 50, 100, and 250 km inland of the continental edge (Fig. 13). The two western profiles are strong primarily because of the shallow basalt layer at the top of the subducted slab and the absence of a weak layer in the middle and lower crust (Fig. 13). The upper-crustal strength of the accretionary prism rocks in the two western columns is derived from the low geotherm associated with the subduction process. The high upper-crustal strength profile in column F (250 km inland) results from the quartz diorite representation of Peninsular Range batholith rocks (Fig. 13).

Several weak zones are conspicuous (Fig. 13). A continuous zone of low strength cuts the mantle lithosphere following the inclined basaltic layer at the top of the deeper

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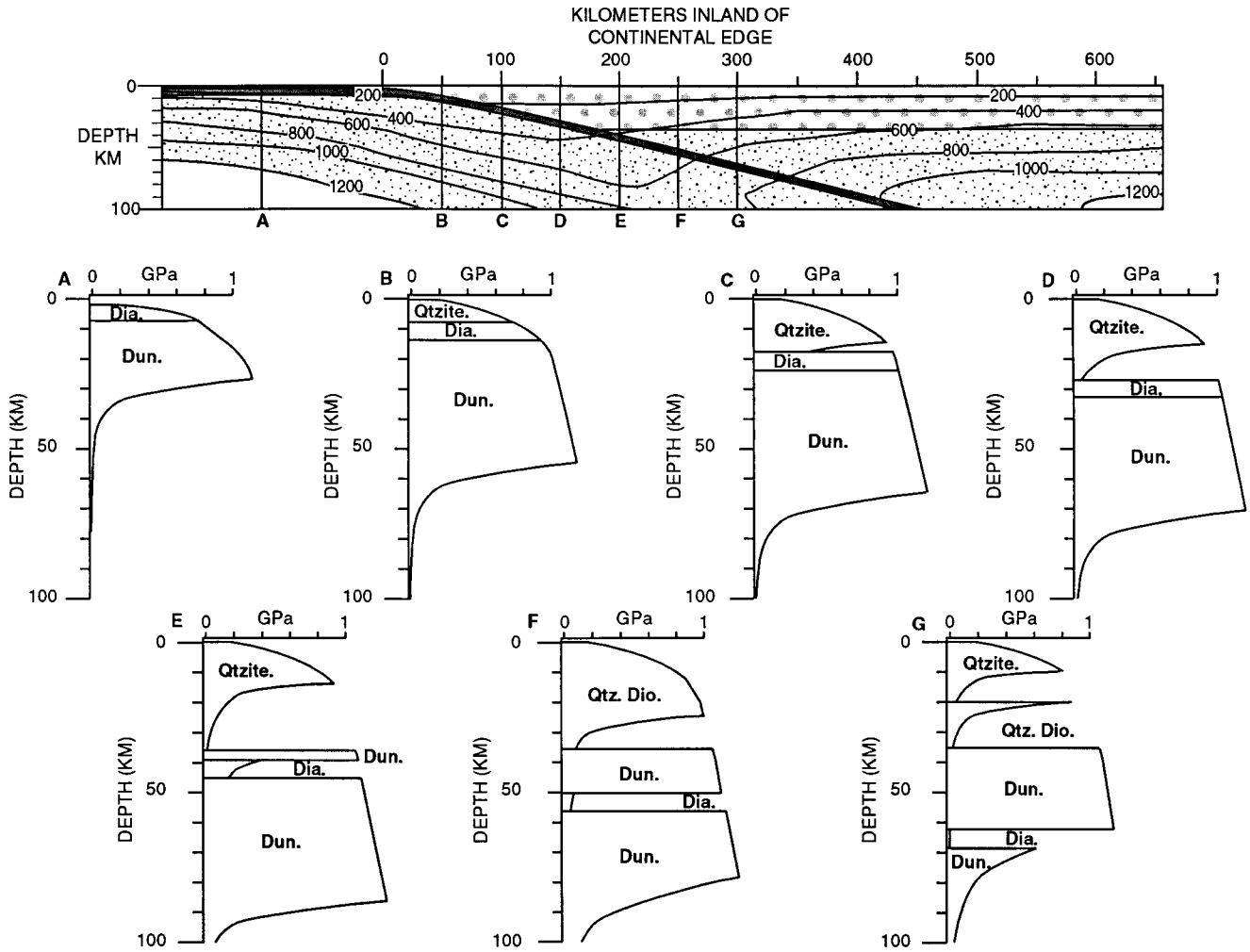

Figure 13. Variability in predicted rock strength versus depth for selected lithospheric columns within the thermal framework of the 20 Ma model from Figure 12 at the latitude of Los Angeles. Strength versus depth plotted for seven columns based on strength/temperature curves of Kirby (1985) that are shown in Figure 13. Upper limits to strength curves are generalized from Ranalli (1987, Figs. 12.1 and 12.2). Thin dark gray zone in cross section represents 6-km-thick layer of basalt/gabbro crust at top of oceanic lithosphere and subducted slab. Dotted layer represents 35-km-thick continental crust. Stippled pattern is conducting lid of mantle lithosphere above 1200 C isotherm. Isotherms shown in degrees Celsius at 200 intervals. Column A is oceanic lithosphere to west of continental edge. Columns B through G are at 50 km intervals, starting at (B) 50 km inland of continental edge to (G) 300 km inland of edge. In all columns, diabase represents oceanic crustal layer and dunite represents mantle lithosphere. At crustal levels quartzite represents rocks of accretionary prism, assumed to make up entire crust down to basalt layer, in columns B through E. Quartz diorite represents rocks of Peninsular Range batholith in column F. Strength of continental crust is considered to be represented by that of quartzite to 20 km depth and by that of quartz diorite at 20–35 km in column G. Dia., diabase; Dun., dunite; Qtzite., quartzite; Qtz. Dio., quartz diorite.

parts of the subducted slab. The lower and middle crust in the accretionary prism are weak because of the quartzite rheology used in the model. The quartz diorite rheology yields a zone of weakness in the deep crust beneath the batholithic belt. Two weak layers occur in the middle and lower crust of the eastern column in the region of the southern Basin and Range province (Fig. 13). It is reasonable to expect that stresses accumulating from the motion of the Pacific plate relative to the North American interior would release first in the weak zones.

Implications

One of the biggest implications of the thermal and mechanical considerations is that they help substantiate the mechanical link between the west-moving Pacific plate and the North American continental edge. The closer a globally oriented spreading ridge is to North America the more likely it is to stall, and once stalled it probably only takes a few million years for it to cool enough to support levels of stress that exceeded the yield strength of many parts of the nearby continental crust. The strong frictional tie across the subduction interface not only terminates the descent of the young subducted slab, but it provides a way to transmit the stresses arising from Pacific plate relative motion into the weakest zones in the continental crust.

We think the arguments made regarding

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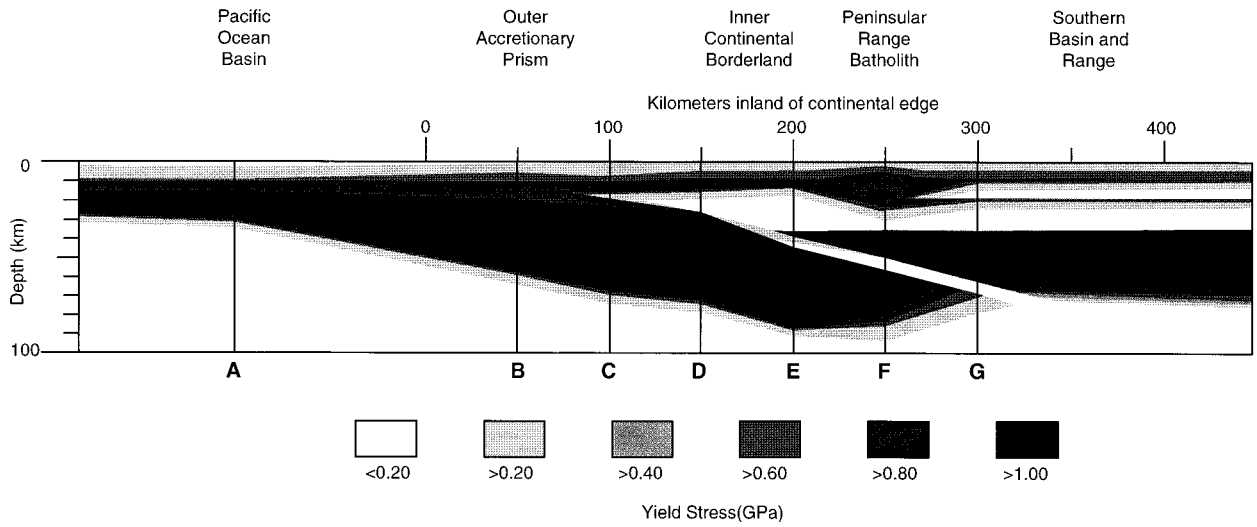

Figure 14. Cross section at 20 Ma of predicted yield stress in lithosphere from Pacific Ocean basin to southern part of Basin and Range structural province. Yield stress interpolated from columns and thermal model shown in Figure 15.

the reduction of slab-pull and ridge-push forces relative to the retarding forces associated with young lithosphere answer the other questions left open by the stall-and-deform and local-ocean kinematic models. At the same time they amplify concerns regarding the continued descent of the shallow part of the slab that is an integral part of all shallow slab-gap models. Thus, the stall-and-deform and local-ocean kinematic models offer the best ways to view the evolution of the Pacific/North American contact zone.

Controlled-source seismic investigations along the western margin of North America have consistently documented the presence of high velocity layers at the base of the continental crust. Miller et al. (1992) and Howie et al. (1993) modeled an oceanic slab beneath the margin at least as far east as the San Andreas Fault at the latitude of Santa Maria, California. There are also thick high-velocity layers beneath San Francisco Bay (Brocher et al., 1993) and beneath Monterey Bay (Page and Brocher, 1993) that have been modeled as oceanic crust. Fuis and Mooney (1990) interpreted a basaltic layer at depth beneath central California east of the San Andreas fault to the Sierran block. High velocities characteristic of basalt are also present under the southern Californian shoreline to the west of the Peninsular Range batholith (D. A. Okaya and Y. G. Li, 1992, written commun.).

The thermal model has important implications. Younger slabs rapidly reheat at sublithospheric levels, and the subducting slab beneath the western Cordillera was much younger during the Neogene than it was during Laramide times. Thus, there was a profound reduction in its coherency that was accompanied by a west recession of the edge of the coherent slab with time. At shallow levels the young slab remains in contact with the continental lithosphere, where it cools and strengthens rather than sinking away to be replaced by upwelled asthenosphere or thermally recycling to become "indistinguishable from asthenosphere" (Severinghaus and Atwater, 1990). No part of the slab is completely assimilated during the times and over the range of depths that we model it. At places its strength and coherency are reduced, but it everywhere maintains a well-developed thermal deficit. Once it is at sublithospheric depths, there is no reason to doubt that the cool, although probably incoherent, material continues its downward flow. Thus the slab-gap forms deep and is probably better envisioned as a zone of slab attenuation.

The inclined zone of weakness in the lithospheric mantle that is associated with the basaltic layer at the top of the descended oceanic slab offers an ideal break between the Pacific and North American plates at the level of the mantle lithosphere (Fig. 14). The relative northwest drift of the underlying Pacific plate could be accommodated across this zone. We view the resultant motion as a type of reverse-subduction. There would also be a strong component of lateral (horizontal) shear.

The modeled strength profiles of the continental crust help explain why deformation might nucleate in some areas over others. The two weakest crustal columns are E and G (Figs. 13 and 14), on either side of the Peninsular Range batholith. These columns roughly correspond to the continental borderlands and southern Basin and Range, both highly deformed areas. A surprisingly strong link between the subducted plate and the overlying continental crust is predicted near the continental edge in columns B and C (Figs. 13 and 14). Such strength is required if the continental edge is to follow the last-formed magnetic anomaly on the rapidly moving Pacific plate. The uniform Pacific motion of the strong layer of dunite in the subducted slab should impart a strain gradient through viscous flow within the weak zones in the middle and deep continental crust, particularly in columns D through G (Figs. 13 and 14). In the deepest crust, much of that flow should be localized between columns D and F where the weak basalt layer emerges from the mantle lithosphere (Figs. 13 and 14). Strong upper crustal layers would essentially float and deform differentially on this flowing deep crustal layer.

DISCUSSION AND CONCLUSIONS

Many aspects of the tectonic development of the plate margin and the western Cordillera are consistent with our model. Figure 15 summarizes the sequential development of the Pacific/North American system in 5 m.y. increments.

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Timing of Events

Present. The modern San Andreas fault system is active and the Gulf of California is fully developed as a new ocean basin within which Pacific/North American relative motion is entirely accommodated. As such, the Gulf of California is a well-developed local ocean system (Fig. 15A[1]). Upper crustal shortening is occurring in much of the Coast Range region of central California (Page, 1992) and in parts of southern California. Regional extension is active along the margins of the Great Basin (e.g., Smith, 1978).

5 Ma. The modern San Andreas fault system was active and the entire coastal strip from the southern tip of Baja Peninsula to the Mendocino fracture zone was in contact with the Pacific plate (Fig. 15B). Relatively intense extension was occurring in the southwestern Great Basin (Wernicke, 1992). The Gulf of California had undergone or was undergoing a transition from continental rifting to a local sea-floor spreading system (Sawlan, 1991; Lyle and Ness, 1991) with Pacific/North American orientations.

10 Ma. Sea-floor spreading had stalled offshore as far south as the southern tip of Baja California (Lonsdale, 1991), so the Pacific/North American contact encompassed most of the length of the Californias (Fig. 15C). The bulk of the Transverse Range rotation and borderland extension was over by this time (Hornafius et al., 1986; Crouch and Suppe, 1993), but intense extension was underway inland at the Gulf of California rift (Stock and Hodges, 1989, 1990; Lyle and Ness, 1991). Extension was strong in the southwestern part of the Great Basin and moderate extension was occurring throughout much of the remainder of that province (Wernicke, 1992).

15 Ma. The northern terminus of Pacific/Farallon spreading was located near the Patton fracture zone and the southern borderland rift of Crouch and Suppe (1993) (Fig. 15D). North of that point to the Mendocino fracture zone, the continental rocks were in contact with the Pacific plate or microplates that were locked to it. The southern borderland rift and the inner continental borderland were actively extending while the western Transverse Ranges were undergoing clockwise rotation (Crouch and Suppe, 1993; Legg, 1991; Sedlock and Hamilton, 1991). Core-complex and Mojave Desert extension were over by this time (Axen et al., 1993), but parts of the southeastern Great Basin were actively extending (Bohannon, 1984; Wernicke et al., 1988).

20 Ma. The microplates between the Farallon and Murray fracture zones were captured by the Pacific plate or about to become so (Lonsdale, 1991), so a longer segment of the North American edge was gradually exposed to Pacific motion (Fig. 15E) than had been before. On the continent, inland of the contact zone, Transverse Range rotation and extension in the inner continental borderland had been initiated by this time (Hornafius et al., 1986; Legg, 1991; Crouch and Suppe, 1993). Core-complex extension was winding down (Spencer and Reynolds, 1989; Axen et al., 1993; Gans et al., 1989), but parts of the Mojave Desert may have been actively extending (Dokka, 1989).

25 Ma. There was contact of the Pacific and North American plates between the Mendocino and Farallon fracture zones. South of there, the Farallon plate broke into microplates between the Farallon and Morro fracture zones (Fig. 15F). Spreading ridge and fracture zone orientations were governed by Pacific/North American relative motion in the microplate system (Fig. 5). Core-complex extension became active by this time due east of the short Pacific/North American contact zone at the continental margin (Spencer and Reynolds, 1989; Axen et al., 1993; Gans et al., 1989). The timing of the initiation of core-complex extension corresponds spatially with the northward displacement of the Mendocino fracture zone (Glazner and Bartley, 1984; Glazner and Supplee, 1982; Axen et al., 1993; Gans et al., 1989). The timing of the cessation of core-complex extension (Axen et al., 1993) corresponds with the northward migration of the active microplate system (Fig. 15F).

30 Ma. The initial contact of the Pacific and Farallon/Vancouver spreading center was about to occur (29 Ma) south of the Pioneer fracture zone. The ridge approached the part of the continent now known as the borderlands and western Transverse Ranges (Fig. 15G). Once achieved, the contact completely isolated the Juan de Fuca (formerly Vancouver) plate from the Farallon plate (Fig. 15G). Prior to 30 Ma, the Farallon and Vancouver plates were separated only by a linear sawtooth pattern of anomalies and pseudofaults between the Pioneer and Murray fracture zones (Atwater, 1989).

Pacific/North American Interaction and Continental Extension

The globally oriented spreading ridge between the Pacific and Farallon plates contacted the continental edge between the Pioneer and Farallon fracture zones at about 29 Ma (Fig. 15G). Core-complex extension began at about the same time, directly inland in what is now Arizona (e.g., Axen et al., 1993). At this early stage both phenomena were spatially limited features, but coincidental nonetheless. The plate contact grew in length once the ridge segment between the Mendocino and Pioneer fracture zones also contacted the continent shortly after 27 Ma, and so did the extent of core-complex extension (e.g., Axen et al., 1993). This is consistent with the stall-and-deform model, and it suggests that the plate-capture phenomenon began soon after the ridge contacted the continent. Pacific plate relative motion was imparted to the base of the continental crust through the part of the subducted Farallon plate that was captured by the Pacific plate.

The spatial and temporal correspondence of the cessation of the core-complex activity with the northward migration of the active microplate system is consistent with the application of the local-ocean model to the part of the margin between the Farallon and Morro fracture zones (Figs. 5 and 15F). That part of the subducted Farallon plate was apparently captured by North America. North American capture explains the close correspondence in orientations of fracture zones and ridges with Pacific/North American motion direction, a spreading rate that is exactly half the rate of Pacific/North American divergence, and the cessation of continental deformation inland.

The stall-and-deform model becomes immediately applicable once a local-ocean system stops being active or once a ridge dies offshore. This is suggested by the close spatial correspondence of the initiation of Transverse Range rotation and intense crustal extension in the inner continental borderland with the death of local-ocean spreading between the Farallon and Morro fracture zones between 20 and 18 Ma (Figs. 15E and 15D). A similar situation occurred when spreading ceased along the Magdalena/Guadalupe spreading ridge at 12.5 Ma. Extension began almost immediately in the region that is now the Gulf

[1]Loose insert: Figure 15 is on a separate sheet accompanying this issue.

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of California inland of the dead ridge (Figs. 15D and 15C).

Extension in the Great Basin is also consistent with the stall-and-deform model. Most of the extension in this part of the Basin and Range structural province took place after 15 Ma as the Mendocino fracture zone migrated from the southern to northern ends of the Great Valley and Sierra Nevada belts (Figs. 15D through 15A). We think the strong tractional forces at the base of the continental crust, exerted by the captured Farallon slab fragment, caused the Sierran batholithic belt to migrate due west as a unit, following the diverging Pacific plate throughout this time period. The strength and coherency of the batholith provided a strong control over the distribution of extension in the Great Basin.

Could another mechanism of continental deformation explain the same observations using one of the shallow-slab-gap models? In our kinematic analysis, it was suggested that a gravitationally unstable crust might expand westward following the diverging Pacific plate. The gravitational instability might be a precondition imposed by excessive regional elevations (e.g., Coney and Harms, 1984), structural over thickening (e.g., Sonder et al., 1987), or high thermal gradients (e.g., Hamilton, 1987). Possibly no cause is needed if crust is simply unstable on its own (e.g., Glazner and Bartley, 1984). We think that it is hard to explain the close correspondence in position and breadth of the Pacific/North American contact with zones of active extension using these concepts. By necessity, any type of precondition would have been present over the entire breadth of the modern extended area including Mexico and the northern Great Basin (Fig. 15A). However, only that part of the unstable continent that is adjacent to an oceanic plate with Pacific motion actually did undergo active extension at any given time. Those regions inland of active subduction zones and actively forming microplate systems were not extending even though the suggested gravitational instability would have had to have been present. No Pacific-plate drive is advocated by these models, only Pacific-plate control. We think the stall-and-deform model is a better way to drive the extensional deformation.

Temperature, Strength, and Deformation

The rock strength at 20 Ma (Fig. 14) offers an explanation of why extensional deformation began inland of the Peninsular Ranges. That is where the weakest upper crustal rocks may have been. The 20 Ma geotherm in the coastal region in Figure 10 is low because it had not equilibrated from the pre–30 Ma subduction. The 30 Ma geotherm was even lower (Fig. 10), which resulted in a high relative yield strength of each rock column west of the Peninsular Ranges. The lower subduction-related temperatures did not extend to the east side of the batholith (Fig. 10). Thus the southern part of the Basin and Range structural province may have had the weakest crustal column, primarily because the strongest layers are relatively thin there because of its higher thermal gradient (Figs. 13 and 14).

The geotherm in the coastal region (Fig. 10) steepened between 30 and 20 Ma, weakening the upper crust there (Fig. 14). This allowed extension to begin in the inner part of the California Continental Borderland (Figs. 15D and 15E) once the local-ocean system that formed the Monterey microplates became extinct at 19 Ma (Lonsdale, 1991). The spatial position of the rotating southwest tip of the Transverse Ranges corresponds, even at the 1 m.y. resolution level, to that of the extinct spreading ridge at Davidson Seamount during the period between 18 and 10 Ma (Figs. 15C through 15E). This indicates a strong link between the Pacific plate, the subducted slab, and the upper crust near the continental edge following the extinction of spreading at 19 Ma. Presumably this link was the product of the high-strength profile that is predicted for the lithospheric columns near the continental edge (Fig. 14).

Strike-Slip Faulting and the Stall-and-Deform Model

The San Andreas fault system was fully developed in the last 15 m.y., and the main fault strand is associated with greatly accelerated displacements in the last 5–10 m.y. (Stewart and Crowell, 1992). In southern California, the fault developed in the weak rocks to the east of the Peninsular Ranges. In our view, the ideal location for the main strand of the San Andreas in northern California is near column C in Figures 13 and 14. The crust and subducted oceanic lithosphere are firmly linked by friction and the high-strength profile of the rock columns to the west of there (Fig. 14), so that region ought to move uniformly with the Pacific plate. The middle and deep crust should behave with a fluid rheology to the east of column C (Fig. 14). The 15–18 km depth of the top of the weak crustal layer approximately coincides with the observed depth to the base of the seismogenic zone along the fault (Hill et al., 1990). The weak layer in the crust isolates the strong upper crust to the east of column C from the Pacific plate relative motions imparted to the deepest crust by the underlying strong oceanic plate (Fig. 14). Regions in which the zone of active faulting migrated inland with time might be explained by a mechanical thinning of the weak layer caused by a rise of the deep strong layer as it pulls out from beneath the lithosphere to the east. This would pinch the wedge of weak rocks to the east, placing strong upper crust onto strong deep lithosphere with Pacific motions. Regions in which the zone of active faulting migrated west with time might be explained by a gradual steepening of the thermal gradient as the ancient subduction-related gradient equilibrated. This would weaken the crustal column, causing the wedge of weak rocks in the middle crust to migrate to the west. These phenomena should occur interactively.

Pacific/North American plate divergence has had the broad effect of extending the western Cordillera, but there has also been obvious convergence across parts of the San Andreas fault system. Argus and Gordon (1991) compared very long baseline interferometry (VLBI) data for motions of subelements within the western Cordillera with the NUVEL-1 model (DeMets et al., 1990) for Pacific/North American motion. They compared San Andreas fault motion vectors, which differ slightly from ideal Pacific/North American relative plate motions, with those originating from Basin and Range extension, including the westward drift of the Sierra Nevada block. Argus and Gordon (1991) concluded that compression across the San Andreas fault absorbs the difference in these relative motions. The problem is essentially one of strain partitioning. The coastal region adjacent to the initial Pacific/North American contact has shifted with nearly 100% of each of the north and west components of Pacific plate drift. A large percentage of the west drift, but a very small percentage of the north drift, has been imparted to the Sierran block. The Cascadia margin seems to have received almost 100% of the west drift, but little or none of the north drift. The Peninsular Ranges have moved with equal percentages of north and west drift, but Peninsular Range totals are far less than that of the initial contact region because of the influence of the microplate systems offshore. Local zones of compres-

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sion and fault bending have developed in between these larger, semicoherent upper-crustal blocks that have exhibited differential movement histories.

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REFERENCES CITED

Addicott, W. O., 1968, Paleontologic evidence for large post–early Miocene lateral slip on the San Andreas fault, California: Geological Society of America Special Paper 115, p. 307.

Anderson, R. E., 1971, Thin skin distension in Tertiary rocks of southeastern Nevada: Geological Society of America Bulletin, v. 82, p. 43–58.

Anderson, R. E., 1973, Large magnitude late Tertiary strike-slip faulting north of Lake Mead, Nevada: U.S. Geological Survey Professional Paper 794, 18 p.

Anderson, R. E., Longwell, C. R., Armstrong, R. L., and Marvin, R. F., 1972, Significance of K-Ar ages of Tertiary rocks from the Lake Mead region, Nevada-Arizona: Geological Society of America Bulletin, v. 83, p. 273–287.

Argus, D. F., and Gordon, R. G., 1991, No-net-rotation model of current plate velocities incorporating plate motion model NUVEL-1: Geophysical Research Letters, v. 18, p. 2039–2042.

Armstrong, R. L., and Ward, P. L., 1991, Evolving geographic patterns of Cenozoic magmatism in the North American Cordillera; the temporal and spatial association of magmatism and metamorphic core complexes, *in* Mid-Tertiary Cordilleran magmatism; plate convergence versus intraplate processes: Journal of Geophysical Research, v. 96, p. 13201–13224.

Atwater, T., 1970, Implications of plate tectonics for the Cenozoic tectonic evolution of western North America: Geological Society of America Bulletin, v. 81, p. 3513–3535.

Atwater, T., 1989, Plate tectonic history of the Northeast Pacific and western North America, *in* Winterer, E. L., Hussong, D. M., and Decker, R. W., eds., The eastern Pacific Ocean and Hawaii: Boulder, Colorado, Geological Society of America, Geology of North America, v. N, p. 21–72.

Atwater, T., and Molnar, P., 1973, Relative motion of the Pacific and North American plates deduced from sea-floor spreading in the Atlantic, Indian, and South Pacific oceans, *in* Proceedings of Conference on tectonic problems of the San Andreas fault system: Stanford University Publications in the Geological Sciences, v. 13, p. 136–148.

Atwater, T., and Severinghaus, J., 1989, Tectonic map of the northeast Pacific Ocean, *in* Winterer, E. L., Hussong, D. M., and Decker, R. W., eds., The eastern Pacific Ocean and Hawaii: Boulder, Colorado, Geological Society of America, Geology of North America, v. N, plate 3-A-C.

Axen, G. J., Taylor, W. J., and Bartley, J. M., 1993, Space-time patterns and tectonic controls of Tertiary extension and magmatism in the Great Basin of the western United States: Geological Society of America Bulletin, v. 105, p. 56–76.

Aydin, A., and Page, B. M., 1984, Diverse Pliocene-Quaternary tectonics in a transform environment, San Francisco Bay region, California: Geological Society of America Bulletin, v. 95, no. 11, p. 1303–1317.

Berggren, W. A., Kent, D. V., Flynn, J. J., and Van Couvering, J. A., 1985, Cenozoic geochronology: Geological Society of America Bulletin, v. 96, no. 11, p. 1407–1418.

Bird, P., 1984, Laramide crustal thickening event in the Rocky Mountain foreland and Great Plains: Tectonics, v. 3, p. 741–758.

Bohannon, R. G., 1984, Nonmarine sedimentary rocks of Tertiary age in the Lake Mead region, southeastern Nevada and northwestern Arizona: U.S. Geological Survey Professional Paper 1259, p. 72.

Bohannon, R. G., 1979, Strike-slip faults of the Lake Mead region of southern Nevada, *in* Armentrout, J. M., Cole, M. R., TerBest, H., Jr., eds., Cenozoic paleogeography of the western United States: Pacific Coast Paleogeography Symposium, v. 3, p. 129–139.

Brocher, T. M., Holbrook, W. S., ten Brink, U. S., Hole, J. A., and Klemperer, S. L., 1993, Results from wide-angle seismic reflection and refraction profiling of the San Francisco Bay area during BASIX: Geological Society of America Abstracts with Programs, v. 25, p. A-310.

Burchfiel, B. C., Hodges, K. V., and Royden, L. H., 1987, Geology of Panamint Valley–Saline Valley pull-apart system, California; palinspastic evidence for low-angle geometry of a Neogene range-bounding fault: Journal of Geophysical Research, v. 92, p. 10422–10426.

Burke, D. B., and McKee, E. H., 1979, Mid-Cenozoic volcano-tectonic troughs in central Nevada: Geological Society of America Bulletin, v. 90, p. 1181–1184.

Butler, P. R., Troxel, B. W., and Verosub, K. L., 1988, Late Cenozoic history and styles of deformation along the southern Death Valley fault zone, California: Geological Society of America Bulletin, v. 100, p. 402–410.

Cemen, I., Wright, L. A., Drake, R. E., and Johnson, F. C., 1985, Cenozoic sedimentation and sequence of deformational events at the southeastern end of Furnace Creek strike-slip fault-zone, Death Valley region, California, *in* Biddle, K. T., and Christie-Blick, N., eds., Strike-slip deformation, basin formation, and sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 37, p. 127–139.

Clark, J. C., Brabb, E. E., Greene, H. G., and Ross, D. C., 1984, Geology of Point Reyes Peninsula and implications for San Gregorio fault history, *in* Crouch, J. K., and Bachman, S. B., eds., Tectonics and sedimentation along the California margin: Society of Economic Paleontologists and Mineralogists, Pacific Section, v. 38, p. 67–86.

Coney, P. J., 1987, The regional tectonic setting and possible causes of Cenozoic extension in the North American Cordillera, *in* Coward, M. P., Dewey, J. F., and Hancock, P. L., eds., Continental extensional tectonics: Geological Society of London Special Publication 28, p. 177–186.

Coney, P. J., and Harms, T. A., 1984, Cordilleran metamorphic core complexes: Cenozoic relicts of Mesozoic compression: Geology, v. 12, p. 550–554.

Coney, P. J., and Reynolds, S. J., 1977, Cordilleran Benioff zones: Nature, v. 270, p. 403–406.

Crouch, J. K., 1979, Neogene tectonic evolution of the California continental borderland and western Transverse Ranges: Geological Society of America Bulletin, v. 90, p. 338–345.

Crouch, J. K., 1981, Northwest margin of California continental borderland; marine geology and tectonic evolution: American Association of Petroleum Geologists Bulletin, v. 65, no. 2, p. 191–218.

Crouch, J. K., and Suppe, J., 1993, Late Cenozoic tectonic evolution of the Los Angeles Basin and inner California borderland: A model for core complex–like crustal extension: Geological Society of America Bulletin, v. 105, p. 1415–1434.

Crouch, J. K., Bachman, S. B., and Shay, J. T., 1984, Post-Miocene compressional tectonics along the central California margin, *in* Crouch, J. K., and Bachman, S. B., eds., Tectonics and sedimentation along the California margin: San Diego, California, Pacific Section of the Society of Economic Paleontologists and Mineralogists Field Trip Guidebook 38, 37 p.

Crowell, J. C., 1962, Displacement along the San Andreas fault, California: Geological Society of America Special Paper 71, 61 p.

Crowell, J. C., 1979, The San Andreas fault system through time: Journal of the Geological Society of London, v. 136, part 3, p. 293–302.

Crowell, J. C., 1982, The Violin Breccia, Ridge Basin, southern California, *in* Crowell, J. C., and Link, M. H., eds., Geologic history of Ridge Basin, southern California: Santa Barbara, California, University of California, p. 89–98.

Crowell, J. C., and Ramirez, V. R., 1979, Late Cenozoic faults in southeastern California, *in* Crowell, J. C., and Sylvester, A. G., eds., Tectonics of the juncture between the San Andreas fault system and the Salton Trough, southeastern California; a guidebook: Santa Barbara, California, University of California, p. 27–39.

Davies, G. F., 1988, Role of the lithosphere in mantle convection: Journal of Geophysical Research, v. 93, p. 10451–10466.

Davis, G. A., and Burchfiel, B. C., 1973, Garlock fault; An intracontinental transform structure, southern California: Geological Society of America Bulletin, v. 84, p. 1407–1422.

DeMets, C., Gordon, R. G., Argus, D. F., and Stein, S., 1990, Current plate motions: Geophysical Journal International, v. 101, p. 425–478.

Denlinger, R. P., 1992, A revised estimate for the temperature structure of the oceanic lithosphere: Journal of Geophysical Research, v. 97, p. 7219–7222.

Dickinson, W. R., and Snyder, W. S., 1979, Geometry of triple junctions related to San Andreas transform: Journal of Geophysical Research, v. 84, no. B2, p. 561–572.

Dokka, R. K., 1989, The Mojave extensional belt of southern California: Tectonics, v. 8, no. 2, p. 363–390.

Douglas, J., 1961, A survey of numerical methods for parabolic differential equations: Advances in Computers, v. 2, p. 1–54.

Dumitru, T. A., 1991, Effects of subduction parameters on geothermal gradients in forearcs, with an application to Franciscan subduction in California: Journal of Geophysical Research, v. 96, p. 621–641.

Durham, W. B., Mirkovich, V. V., and Heard, H. C., 1987, Thermal diffusivity of igneous rocks at elevated pressure and temperature: Journal of Geophysical Research, v. 92, p. 11615–11634.

Elston, W. E., 1984, Subduction of young oceanic lithosphere and extensional orogeny in southwestern North America during mid-Tertiary time: Tectonics, v. 3, p. 229–250.

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

Fuis, G. S., and Mooney, W. D., 1990, Lithospheric structure and tectonics from seismic-refraction and other data, *in* Wallace, R. E., ed., The San Andreas fault system, California: U.S. Geological Survey Professional Paper 1515, p. 207–236.

Furlong, K. P., Hugo, W. D., and Zandt, G., 1989, Geometry and evolution of the San Andreas fault zone in northern California: Journal of Geophysical Research, v. 94, no. 3, p. 3100–3110.

Gans, P. B., Mahood, G. A., and Schermer, E. R., 1989, Synextensional magmatism in the Basin and Range province; a case study from the eastern Great Basin: Geological Society of America Special Paper 233, 53 p.

Glazner, A. F., and Bartley, J. M., 1984, Timing and tectonic setting of Tertiary low-angle normal faulting and associated magmatism in the southwestern United States: Tectonics, v. 3, p. 385–396.

Glazner, A. F., and Supplee, J. A., 1982, Migration of Tertiary volcanism in the southwestern United States and subduction of the Mendocino fracture zone: Earth and Planetary Science Letters, v. 60, p. 429–436.

Graham, S. A., and Dickinson, W. R., 1978, Apparent offsets of on-land geologic features across the San Gregorio–Hosgri fault trend, *in* San Gregorio–Hosgri fault zone, California: California Division of Mines and Geology Special Report 137, p. 13–23.

Hall, C. A. J., 1975, San Simeon–Hosgri fault system, coastal California; economic and environmental implications: Science, v. 190, p. 1291–1294.

Hamilton, W. B., 1987, Crustal extension in the Basin and Range province, southwestern United States, *in* Coward, M. P., Dewey, J. F., and Hancock, P. L., eds., Continental extensional tectonics: Geological Society [London] Special Publication 28, p. 155–176.

Hamilton, W., and Myers, W. B., 1966, Cenozoic tectonics of the western United States: Reviews of Geophysics, v. 4, p. 509–549.

Hausback, B. P., 1984, Cenozoic volcanic and tectonic evolution of Baja California Sur, Mexico: Pacific Section of the Society of Economic Paleontologists and Mineralogists Field Trip Guidebook 39, p. 219–236.

Hill, D. P., Eaton, J. P., and Jones, L. M., 1990, Seismicity, 1980–86, *in* Wallace, R. E., ed., The San Andreas fault system, California: U.S. Geological Survey Professional Paper 1515, p. 115–151.

Hill, M. L., and Dibblee, T. W. J., 1953, San Andreas, Garlock, and Big Pine faults, California—A study of the character, history, and tectonic significance of their displacements: Geological Society of America Bulletin, v. 64, p. 443–458.

Hornafius, J. S., Luyendyk, B. P., Terres, R. R., and Kamerling, M. J., 1986, Timing and extent of Neogene tectonic rotation in the western Transverse Ranges, California: Geological Society of America Bulletin, v. 97, p. 1476–1487.

Howell, D. G., 1975, Early and middle Eocene shoreline offset by the San Andreas fault, southern California, *in* San Andreas fault in southern California; a guide to San Andreas fault from Mexico to Carrizo Plain: California Division of Mines and Geology Special Report 118, p. 69–74.

Howell, D. G., and Vedder, J., 1981, Structural implications of stratigraphic discontinuities across the southern California borderland, *in* Ernst, W. G., ed., The geotectonic development of California (Rubey Volume I): Los Angeles, University of California, p. 535–558.

Howell, D. G., Stuart, C. J., Platt, J. P., and Kamerling, M. J., 1974, Possible strike-slip faulting in the southern California Borderland: Geology, v. 2, p. 93–98.

Howell, D. G., Moore, G. W., and Wiley, T. J., 1987, Tectonics and basin evolution of western North America; an overview, *in* Scholl, D. W., Grantz, A., and Vedder, J. G., eds., Geology and resource potential of the continental margin of western North America and adjacent ocean basins, Beaufort Sea to Baja California: Menlo Park, California, Circum Pacific Council for Energy and Mineral Resources, p. 1–15.

Howie, J. M., Miller, K. C., and Savage, W. U., 1993, Integrated crustal structure across the south central California margin; Santa Lucia Escarpment to the San Andreas fault: Journal of Geophysical Research, v. 98, p. 8173–8196.

Huffman, O. F., 1972, Lateral displacement of upper Miocene rocks and the Neogene history of offset along the San Andreas fault in central California: Geological Society of America Bulletin, v. 83, p. 2913–2946.

Ingersoll, R. V., 1982, Triple-junction instability as a cause for late Cenozoic extension and fragmentation of the western United States: Geology, v. 10, p. 621–624.

Jackson, J., and Molnar, P., 1990, Active faulting and block rotations in the western Transverse Ranges, California: Journal of Geophysical Research, v. 95, p. 22073–22087.

Kent, D. V., and Gradstein, F. M., 1985, A Cretaceous and Jurassic geochronology: Geological Society of America Bulletin, v. 96, p. 1419–1427.

Kirby, S. H., 1985, Rock mechanics observations pertinent to the rheology of the continental lithosphere and the localization of strain along shear zones, *in* Carter, N. L., and Uyeda, S., eds., Collision tectonics; deformation of continental lithosphere: Tectonophysics, v. 119, p. 1–27.

Lachenbruch, A. H., and Morgan, P., 1990, Continental extension, magmatism and elevation; formal relations and rules of thumb, *in* Lucchitta, I., and Morgan, P., eds., Heat and detachment in continental extension: Tectonophysics, v. 174, no. 1-2, p. 39–62.

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Legg, M. R., 1985, Geologic structure and tectonics of the inner continental borderland, offshore northern Baja California, Mexico [Ph.D. dissert.]: Santa Barbara, University of California, 410 p.

Legg, M. R., 1991, Developments in understanding the tectonic evolution of the California continental borderland, *in* Osborne, R. H., ed., From shoreline to abyss; contributions in marine geology in honor of Francis Parker Shepard: Los Angeles, California, Society of Economic Paleontologists and Mineralogists Special Publication 46, p. 291–312.

Liggett, M. A., and Childs, J. F., 1977, An application of satellite imagery to mineral exploration: U.S. Geological Survey Professional Paper 1015, p. 253–270.

Lonsdale, P., 1991, Structural patterns of the Pacific floor offshore of Peninsular California, *in* Dauphin, J. P., and Simoneit, B. R. T., eds., Gulf and Peninsular Province of the Californias: American Association of Petroleum Geologists Memoir 47, p. 87–143.

Luyendyk, B. P., 1991, Tectonic events in southern California during Neogene time suggested by paleomagnetic data: Geological Society of America Abstracts with Programs, v. 23, no. 5, p. 22.

Luyendyk, B. P., Kamerling, M. J., and Terres, R., 1980, Geometric model for Neogene crustal rotations in southern California: Geological Society of America Bulletin, v. 91, p. 211–217.

Luyendyk, B. P., Kamerling, M. J., Terres, R. R., and Hornafius, J. S., 1985, Simple shear of southern California during Neogene time suggested by paleomagnetic declinations: Journal of Geophysical Research, v. 90, p. 12454–12466.

Lyle, M., and Ness, G. E., 1991, The opening of the southern Gulf of California, *in* Dauphin, J. P., and Simoneit, B. R. T., eds., Gulf and Peninsular Province of the Californias: American Association of Petroleum Geologists Memoir 47, p. 403–424.

Matthews, V., 1976, Correlation of Pinnacles and Neenach volcanic formations and their bearing on San Andreas fault problems: American Association of Petroleum Geologists Bulletin, v. 60, p. 2128–2141.

McCulloch, D. S., 1989, Evolution of the offshore central California margin, *in* Winterer, E. L., Hussong, D. M., and Decker, R. W., eds., The eastern Pacific Ocean and Hawaii: Boulder, Colorado, Geological Society of America, Geology of North America, v. N, p. 439–470.

McKenna, L. W., and Hodges, K. V., 1990, Constraints on the kinematics and timing of late Miocene–Recent extension between the Panamint and Black Mountains, southeastern California, *in* Wernicke, B. P., ed., Basin and Range extensional tectonics near the latitude of Las Vegas, Nevada: Geological Society of America Memoir 176, p. 363–376.

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

McKenzie, D. P., and Morgan, W. J., 1969, Evolution of triple junctions: Nature, v. 224, p. 125–133.

McKenzie, D. P., and Parker, R. L., 1967, The north pacific—An example of tectonics on a sphere: Nature, v. 216, p. 1276–1280.

Miller, K. C., Howie, J. M., and Ruppert, S. D., 1992, Shortening within underplated oceanic crust beneath the central California margin: Journal of Geophysical Research, v. 97, p. 19961–19980.

Morgan, P., and Burke, K. C., 1985, Collisional plateaus, *in* Carter, N. L., and Uyeda, S., eds., Collision tectonics; deformation of continental lithosphere: Tectonophysics, v. 119, no. 1-4, p. 137–151.

Mount, V. S., and Suppe, J., 1992, Present-day stress orientations adjacent to active strike-slip faults; California and Sumatra, *in* Zoback, M. L., ed., The World Stress Map Project: Journal of Geophysical Research, v. 97, p. 11995–12013.

Nabelek, J., Li, X.-Q., Trehu, A., Azevedo, S., Braunmiller, J., Fabritius, A., Leitner, B., and Zandt, G., 1993, A seismic image of the Cascadia subduction zone: Part II: Geological Society of America Abstracts with Programs, v. 25, no. 6, p. A-312–A-313.

Nicholson, C., Sorlien, C. C., Atwater, T., Crowell, J. C., and Luyendyk, B. P., 1994, Microplate capture, rotation of the western Transverse Ranges, and initiation of the San Andreas transform as a low-angle fault system: Geology, v. 22, p. 491–495.

Nilsen, T. H., 1984, Oligocene tectonics and sedimentation, California, *in* Nilsen, T. H., ed., Fluvial sedimentation and related tectonic framework, western North America: Sedimentary Geology, v. 38, no. 1-4, p. 305–336.

Page, B. M., 1992, Tectonic setting of the San Francisco Bay region, *in* Borchardt, G., ed., Proceedings, Second Conference on earthquake hazards in the eastern San Francisco Bay area: California Division of Mines and Geology Special Publication 113, p. 1–7.

Page, B. M., and Brocher, T. M., 1993, Thrusting of the central California margin over the edge of the Pacific Plate during the transform regime: Geology, v. 21, p. 635–638.

Parsons, B., and Sclater, J. G., 1977, An analysis of the variation of ocean floor bathymetry and heat flow with age: Journal of Geophysical Research, v. 82, p. 803–827.

Powell, R. E., 1993, Balanced palinspastic reconstruction of pre–late Cenozoic paleogeology, southern California; geologic and kinematic constraints on evolution of the San Andreas fault system, *in* Powell, R. E., Weldon, R. J., II, and Matti, J. C., eds., The San Andreas fault system; displacement, palinspastic reconstruction, and geologic evolution: Geological Society of America Memoir 178, p. 1–106.

Powell, R. E., 1982, Crystalline basement terranes in southern eastern Transverse Range, California, *in* Cooper, J. D., ed., Geologic excursions in the Transverse Ranges, southern California; Field trip no. 11: Boulder, Colorado, Geological Society of America, p. 109–136.

Ranalli, G., 1987, Rheology of the Earth: Boston, Massachusetts, Allen and Unwin, 366 p.

Rosa, J. W. C., and Molnar, P., 1988, Uncertainties in reconstructions of the Pacific, Farallon, Vancouver, and Kula plates and constraints on the rigidity of the Pacific and Farallon (and Vancouver) plates between 72 and 35 Ma, *in* Orcutt, J. A., and Sharman, G. F., eds., Special section; tribute to H. W. Menard: Journal of Geophysical Research, v. 93, p. 2997–3008.

Ross, D. C., 1984, Possible correlations of basement rocks across the San Andreas, San Gregorio–Hosgri, and Rinconada–Reliz–King City faults, California: U.S. Geological Survey Professional Paper 37, 64 p.

Sawlan, M. G., 1991, Magmatic evolution of the Gulf of California rift, *in* Dauphin, J. P., and Simoneit, B. R. T., eds., Gulf and Peninsular Province of the Californias: American Association of Petroleum Geologists Memoir 47, p. 301–370.

Saxena, S. K., and Shien, G., 1992, Assessed data on heat capacity, thermal expansion, and compressibility for some oxides and silicates: Journal of Geophysical Research, v. 97, p. 19813–19825.

Sedlock, R., and Hamilton, D. M., 1991, Late Cenozoic tectonic evolution of southwestern California: Journal of Geophysical Research, v. 96, p. 2325–2351.

Severinghaus, J., and Atwater, T., 1990, Cenozoic geometry and thermal state of the subducting slabs beneath western North America, *in* Wernicke, B. P., ed., Basin and Range extensional tectonics near the latitude of Las Vegas, Nevada: Geological Society of America Memoir 176, p. 1–22.

Sharp, R. V., 1967, San Jacinto fault zone in the Peninsular Ranges of California: Geological Society of America Bulletin, v. 78, p. 705–729.

Smith, G. I., 1962, Large lateral displacement on Garlock fault, California, as measured from offset dike swarm: American Association of Petroleum Geologists Bulletin, v. 46, p. 85–104.

Smith, G. I., and Ketner, K. B., 1970, Lateral displacement on the Garlock fault, southeastern California, suggested by offset sections of similar metasedimentary rocks, *in* Geological Survey Research 1970, Chapter D: U.S. Geological Survey Professional Paper 700-D, p. D1–D9.

Smith, R. B., 1978, Seismicity, crustal structure, and intraplate tectonics of the interior of the western Cordillera, *in* Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 111–144.

Sonder, L. J., England, P. C., Wernicke, B. P., and Christiansen, R. L., 1987, A physical model for Cenozoic extension of western North America, *in* Coward, M. P., Dewey, J. F., and Hancock, P. L., eds., Continental extensional tectonics: Geological Society [London] Special Publication 28, p. 187–201.

Spencer, J. E., and Normark, W. R., 1989, Neogene plate-tectonic evolution of the Baja California Sur continental margin and the southern Gulf of California, Mexico, *in* Winterer, E. L., Hussong, D. M., and Decker, R. W., eds., The eastern Pacific Ocean and Hawaii: Boulder, Colorado, Geological Society of America, Geology of North America, v. N, p. 489–497.

Spencer, J. E., and Reynolds, S. J., 1989, Middle Tertiary tectonics of Arizona and adjacent areas, *in* Jenney, J. P., and Reynolds, S. J., eds., Geologic evolution of Arizona: Arizona Geological Society Digest 17, p. 539–574.

Stanley, R. G., 1987, New estimates of displacement along the San Andreas fault in central California based on paleobathymetry and paleogeography: Geology, v. 15, p. 171–174.

Stewart, J. H., 1983, Extensional tectonics in the Death Valley area, California; transport of the Panamint Range structural block 80 km northwestward: Geology, v. 11, p. 153–157.

Stewart, J. H., and Crowell, J. C., 1992, Strike-slip tectonics in the Cordilleran region, western United States, *in* Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., eds., The Cordilleran orogen: Conterminous United States: Boulder, Colorado, Geological Society of America, Geology of North America, v. G-3, p. 609–628.

Stewart, J. H., Albers, J. P., and Poole, F. G., 1968, Summary of regional evidence for right-lateral displacement in the western Great Basin: Geological Society of America Bulletin, v. 79, p. 1407–1413.

Stock, J. M., and Hodges, K. V., 1989, Pre-Pliocene extension around the Gulf of California, and the transfer of Baja California to the Pacific plate: Tectonics, v. 8, p. 99–115.

Stock, J. M., and Hodges, K. V., 1990, Miocene to Recent structural development of an extensional accommodation zone, northeastern Baja California, Mexico: Journal of Structural Geology, v. 12, p. 315–328.

Stock, J., and Molnar, P., 1988, Uncertainties and implications of the Late Cretaceous and Tertiary position of North America relative to the Farallon, Kula, and Pacific plates: Tectonics, v. 7, p. 1339–1384.

Turcotte, D. L., and Schubert, G., 1982, Geodynamics; applications of continuum physics to geological problems: New York, John Wiley & Sons, 450 p.

Wernicke, B., 1992, Cenozoic extensional tectonics of the U.S. Cordillera, *in* Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., eds., The Cordilleran orogen; conterminous United States: Boulder, Colorado, Geological Society of America, Geology of North America, v. G-3, p. 553–581.

Wernicke, B., Walker, J. D., and Beaufait, M. S., 1985, Structural discordance between Neogene detachments and frontal Sevier thrusts, central Mormon Mountains, southern Nevada: Tectonics, v. 4, p. 213–246.

Wernicke, B. P., Hodges, K. V., and Walker, J. D., 1986, Geologic evolution of Tucki Mountain and vicinity, Central Panamint Range, *in* Dunne, G. C., ed., Mesozoic and Cenozoic structural evolution of selected areas, east-central California: Northridge, California State University, p. 67–80.

Wernicke, B. P., Axen, G. J., and Snow, J. K., 1988, Basin and Range extensional tectonics at the latitude of Las Vegas, Nevada: Geological Society of America Bulletin, v. 100, p. 1738–1757.

Wilson, J. T., 1965, A new class of faults and their bearing on continental drift: Nature, v. 207, p. 343–347.

Zandt, G., and Furlong, K. P., 1982, Evolution and thickness of the lithosphere beneath coastal California: Geology, v. 10, p. 376–381.

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