interplanetary magnetic field polarity is toward the sun the daily mean of the horizontal component is less than average. The physical cause for this relationship has not yet been determined, but it may be related to an increased drag on the tail of the geomagnetic field from the northern polar cap during an interplanetary field polarity away from the sun as compared with a polarity toward the sun, or perhaps to plasma flows in the polar cusp regions of the magnetosphere (*6*).

These considerations suggest that we might expect to find in the observed interplanetary field polarity a 26⅞-day recurrence pattern with the field polarity predominantly toward the sun (negative) during approximately the first half of the period and predominantly away from the sun (positive) during approximately the second half of the period. To investigate this possibility we have taken an integral number of 26⅞-day periods from the starting date of 1 July 1926 used by Olsen. Thus period number 505 began on 1 August 1963. The result of such a superposed epoch analysis of the polarity of the observed interplanetary magnetic field from November 1963 through December 1968 is shown in Fig. 2. The polarity pattern expected from the above considerations is clearly evident, and it is very nearly in phase with Olsen's recurrence system. If we assign a 2-day uncertainty in the phase for Fig. 2, then the period as determined over the approximately 550 rotations since 1926 is 26⅞ $\pm$ 0.003 days. A schematic drawing of the rotating solar magnetic "dipole" inferred from these considerations is shown in Fig. 3.

In addition to the two basic polarity intervals per period shown in Fig. 2, there is evidence in the range from 9 days to 19 days for a modulation that may be related to a structure of four polarity changes per period, such as has been observed in the form of four interplanetary magnetic sectors per solar rotation during a portion of the interval from 1963 to 1968 observed with spacecraft (*7*). It thus appears that the basic pattern in the rotating solar layer is two polarity regions per rotation, but that effects perhaps related to solar activity and to the Babcock (*8*) model of solar magnetism cause the observed photospheric (and interplanetary) magnetic fields to display considerable variation about this basic pattern. A relatively unchanging rotating magnetic dipole may be lurking within (figuratively or literally) the highly variable sun that we see in day-by-day observations.

Note added in proof: Yearbooks for most of the years from 1926 to 1964 (*9*) containing observations on the horizontal component at Godhavn have recently become available to us. The recurrence pattern of Olsen shown in Fig. 1 is present in each individual sunspot cycle during this interval, except during the cycle from 1944 to 1954, which appears to be anomalous (*10*).

JOHN M. WILCOX*
WALTER GONZALEZ*

Space Sciences Laboratory,
University of California,
Berkeley 94720

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* Present address: Institute for Plasma Research, Stanford University, Stanford, California.

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Evolving Subduction Zones in the Western United States, as Interpreted from Igneous Rocks

Abstract. *Variations in the ratio of K_2O to SiO_2 in andesitic rocks suggest early and middle Cenozoic subduction beneath the western United States along two subparallel imbricate zones dipping about 20 degrees eastward. The western zone emerged at the continental margin, but the eastern zone was entirely beneath the continental plate. Mesozoic subduction apparently occurred along a single steeper zone.*

Sea floor magnetic anomalies (*1, 2*) and andesitic continental margin volcanism (*3, 4*) indicate that until middle Cenozoic time a spreading rise system was continuous west of North America, and a subduction zone was continuous along the western continental margin. We here attempt to interpret the late Mesozoic and Cenozoic evolution of this subduction system from variations in the composition and distribution of igneous rocks in the western United States.

Modification of the subduction system in late Cenozoic time has produced plate boundaries in the western United States (Fig. 1) that consist of an oblique transform system separating two remaining segments of the spreading ocean rise (*1, 2, 5*). Initiation of extensional faulting through a wide region inland from this transform boundary (Fig. 1) followed termination of the subduction system (*1, 6*). Miocene and younger igneous rocks associated with the extensional faulting are fundamentally basaltic; they include basalt fields, differentiated alkalic basaltic sequences, and bimodal basalt-rhyolite suites (*6*). This late Cenozoic basaltic association is not discussed further.

Lower and middle Cenozoic continental lavas (Fig. 2) are predominantly calc-alkalic intermediate compositional types (andesite-rhyodacite), commonly with associated, more silicic ash-flow sheets. Compositional zonations in some individual ash-flow sheets, from rhyolite upward into quartz latite, record magmatic differentiation in underlying batholithic source chambers (*7*). The intermediate lavas probably are samples of the bulk of these batholiths, with the ash-flow tuffs representing their differentiated tops (*3*). Similar igneous rocks continued to form in later Cenozoic time but were restricted to a shrinking zone opposite the continental margin trench (*6*). By Quaternary time predominantly intermediate-composition igneous activity was restricted to the Cascade Range and to central Mexico (Fig. 1).

These calc-alkalic intermediate igneous suites, although compositionally varied, are within the range observed in active volcanic arcs that are associated with plate convergence (*3, 4*), where chains of andesitic volcanoes are aligned above Benioff seismic zones (*8, 9*). Lower and middle Cenozoic volcanic rocks of the western United States display generally increasing alkali contents toward the continental interior (*4, 10*); these are similar to the systematic chemical variations transverse to individual active arcs of the Pacific basin (*11*). These similarities suggest that the lower and middle Cenozoic andesitic rocks of the western United States, like modern andesites of island arcs, were genetically related to a subduction system.

In regions of active subduction-related volcanism, the ratio of K_2O to SiO_2 in andesitic rocks increases systematically with increasing depth to the Benioff seismic zone (*8, 12*). This correlation seems empirically valid, whatever the mechanism of magma generation. Accordingly, we have estimated, from SiO_2-variation diagrams, the K_2O contents (within ± 0.2 percent K_2O) at 60 percent SiO_2 for about 70 dated middle Cenozoic volcanic suites (*4*). We have converted these figures into depth values by means of graphs of K_2O content plotted against depth to Benioff zone for active andesitic arcs (*8, 12*). We think that most differences greater than 30 km among the individual depth values are significant. The actual depth values may be less significant because of possible differences between K_2O-depth correlations for island arc and continental margin environments, scatter in the K_2O-depth graph for modern andesites (*8, 12*), and our extrapolation to about 100 km greater depths from that graph.

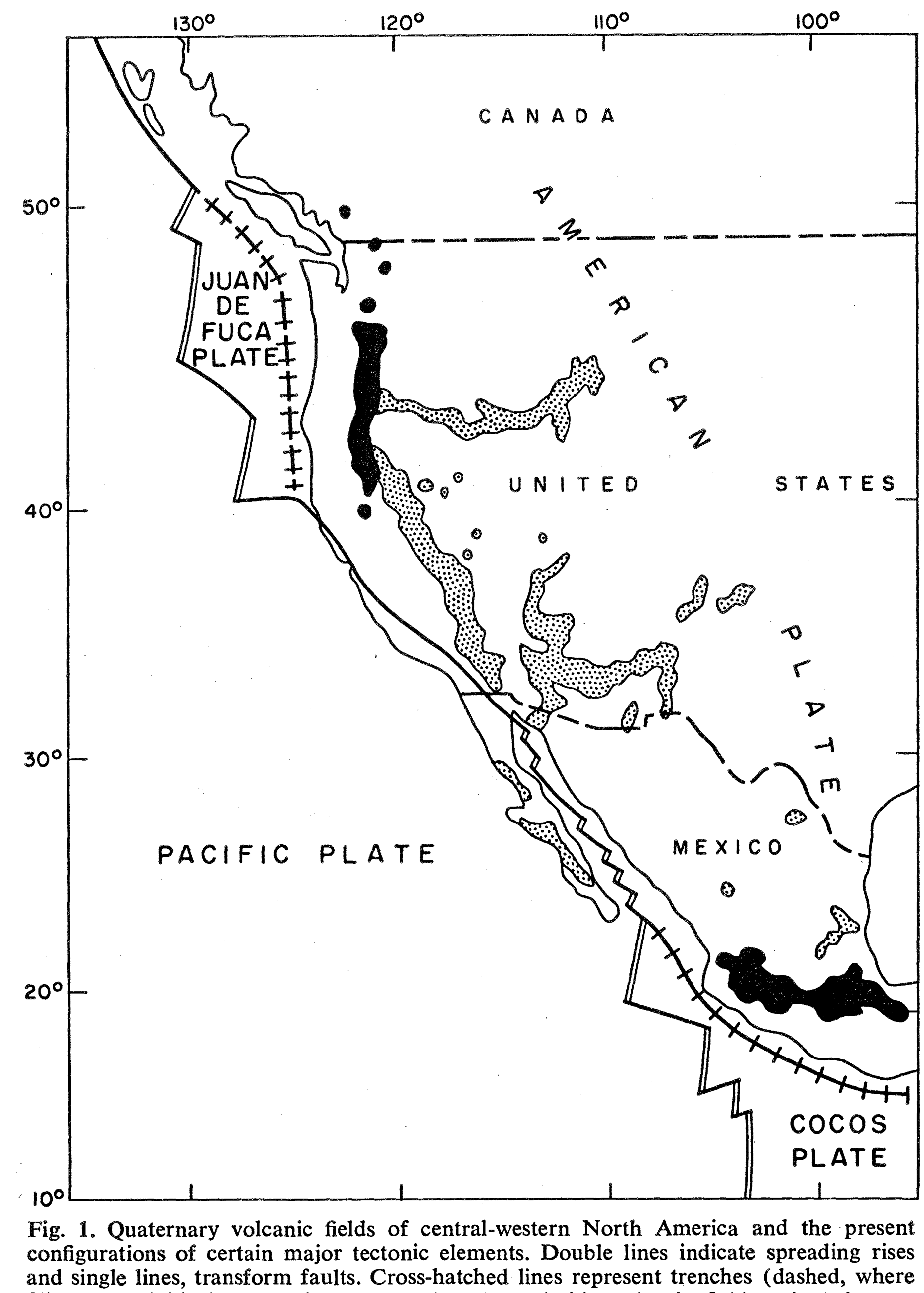


Fig. 1. Quaternary volcanic fields of central-western North America and the present configurations of certain major tectonic elements. Double lines indicate spreading rises and single lines, transform faults. Cross-hatched lines represent trenches (dashed, where filled). Solid black areas show predominantly andesitic volcanic fields; stippled areas show fundamentally basaltic fields (*6*).

The depth values were plotted and contoured at 50-km intervals (Fig. 3a); only two points (asterisks) fall clearly outside the general trends. The plot shows two well-defined regions of eastward-increasing depths, separated by a north-trending discontinuity across which contours cannot easily be connected. This discontinuity extends from the northern Rocky Mountains, south along the Wasatch front, to the western Colorado Plateau (Fig. 3a); it coincides approximately with the east edge of the belt of late Mesozoic–early Tertiary thrusting (*13, 14*). The discontinuity in depth values is at a maximum in the west-central United States and becomes less pronounced toward the Canadian border. In southwestern Canada, Cenozoic igneous rocks are absent east of the projected trend of the discontinuity (approximately coincident with the Rocky Mountain Trench). Petrographic descriptions—chemical data are sparse—suggest that the discontinuity persists southwest into or through Arizona but not far into Mexico.

Cross sections show that the depth values define two parallel zones that dip eastward at an angle of about 15° to 20° (Fig. 3b). Because the western United States has been extended by late Cenozoic basin-range faulting and related extensional deformation (*15*), middle Cenozoic dips of the subduction zones that are defined by these points would have been slightly steeper, probably 20° to 25°. These are gentler than the 45° to 75° dips of active Benioff zones of oceanic arcs (*8, 12, 16*) but may be typical of continental margin seismic zones, which commonly are less steep (*8, 17*).

The western subduction zone of mid-Cenozoic North America should have emerged at the surface as a trench, much like the Peru-Chile trench along the coast of South America today. The western margin of the eastern subduction zone, however, had no recognizable middle Cenozoic surface expression, other than volcanism, that would suggest a continental analog of an oceanic trench. For this reason, we think that the eastern descending plate must have been decoupled from the overlying continental plate and that the eastern subduction zone never emerged at the surface. The shallowest depths of the inferred eastern subduction zone to which volcanism seems to have been related are 175 to 200 km (Fig. 3). These depths correspond with the low-velocity layer of the asthenosphere (*18*), a plausible horizon along which the eastern descending plate might have been decoupled from the continental plate. The two anomalously shallow depths referred to earlier (Fig. 3a, asterisks) are close to the low-velocity layer in the region between the two dipping zones (Fig. 3b, *A-A′*), and the low K_2O contents of these volcanic rocks may reflect the zone of decoupling. The cross section at the bottom of Fig. 3b schematically illustrates our interpretation of the plate boundaries, with decoupling of the eastern zone in the low-velocity layer. The southward shift of igneous activity from Eocene to Oligocene time (Fig. 2) indicates that the imbricate system was an unstable configuration, but the inferred cross-sectional geometry remained similar as the distribution of activity shifted (Fig. 3).

The presence of an imbricate, gently dipping subduction system under the western United States in middle Cenozoic time provides a plate tectonic interpretation for the conspicuous widening to as much as 1500 km across, of the cordilleran belts of igneous and tectonic activity that elsewhere in North and South America are about 500 km wide. Similar reconstructions can be made of paleosubduction zones formed along the western margin of North America through late Mesozoic and Cenozoic time.

Jurassic to Late Cretaceous igneous activity was confined to a narrow region of the western United States (*19*). Granitic rocks as young as about 80 million years are found in the Sierra Nevada batholith (*20*), but Mesozoic igneous rocks of this age and older are sparse east of Nevada. A steep subduction zone (about 50°) has been inferred for the Sierra Nevada region in late Mesozoic time (*9, 21, 22*). Although this interpretation is based on batholithic rocks whose ages range from Early Jurassic to Late Cretaceous, it is supported by the narrowness and linearity of granitic sequences of more restricted age ranges within the composite batholith (*20*). In contrast, middle Cenozoic igneous activity associated with gently dipping subduction zones extends over a broad region (Fig. 2).

The eastward shift of igneous activity in latest Cretaceous and early Cenozoic time was noted long ago by Lindgren (*23*). In the western United States, Upper Cretaceous igneous rocks older than about 80 million years are found mainly in California and western Nevada; uppermost Cretaceous (Laramide) igneous rocks (70 to 65 million years) occur in north-central Nevada (*24*) and as far east as the southern Rocky Mountains (*25*). Our attempts to reconstruct the geometry of the latest Cretaceous-Paleocene subduction zone from the chemistry of Laramide igneous rocks have been hampered by a paucity of chemical data and by extensive alteration of many of the analyzed rocks. The few Laramide igneous suites for which reliable compositional trends can be drawn fit the trends defined by middle Cenozoic igneous suites (Fig. 3a). This suggests that the major shift in the geometry of subduction occurred between about 80 and 70 million years ago.

Concurrently with this shift in igneous activity, an eastward shift in orogenic activity marked the initiation of Laramide foreland deformation. The beginning of Laramide uplift in the central Rocky Mountains, as marked by a change in sedimentation from marine shale to marine and continental sandstone, occurred about 70 million years ago (*26*). These changes in igneous and tectonic patterns, which would be enigmatic in terms of a steeply dip-

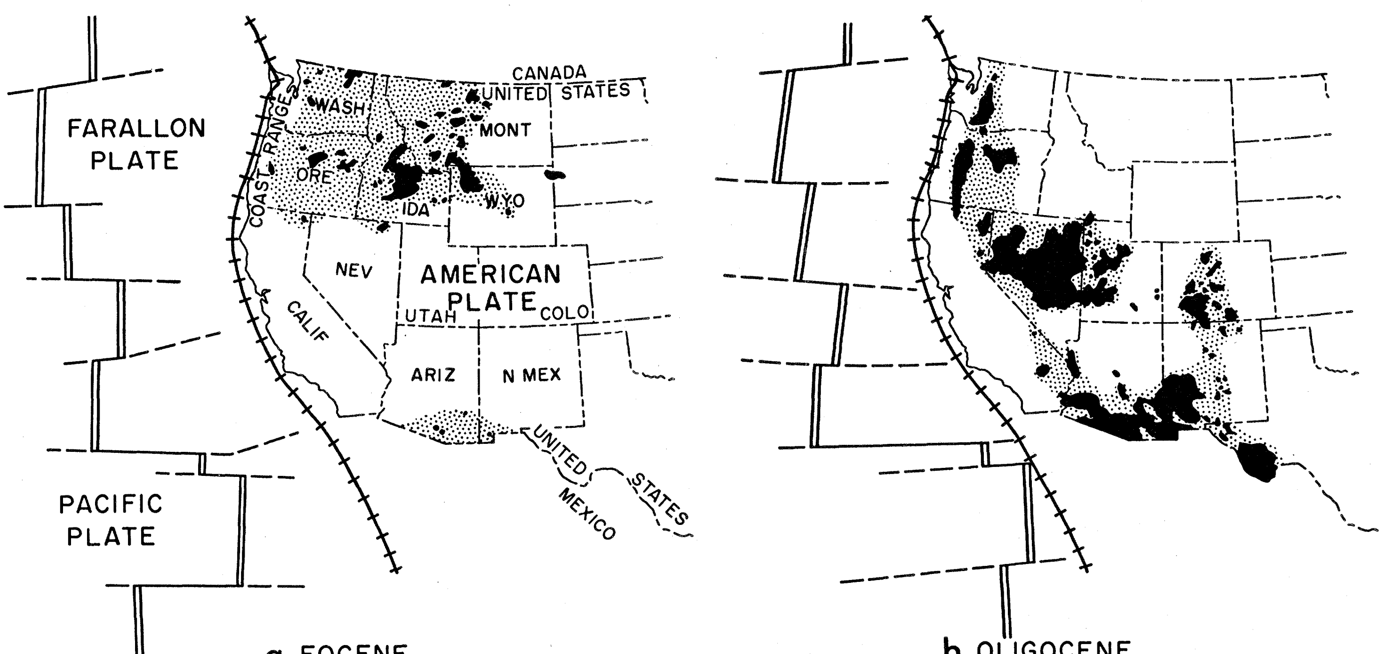
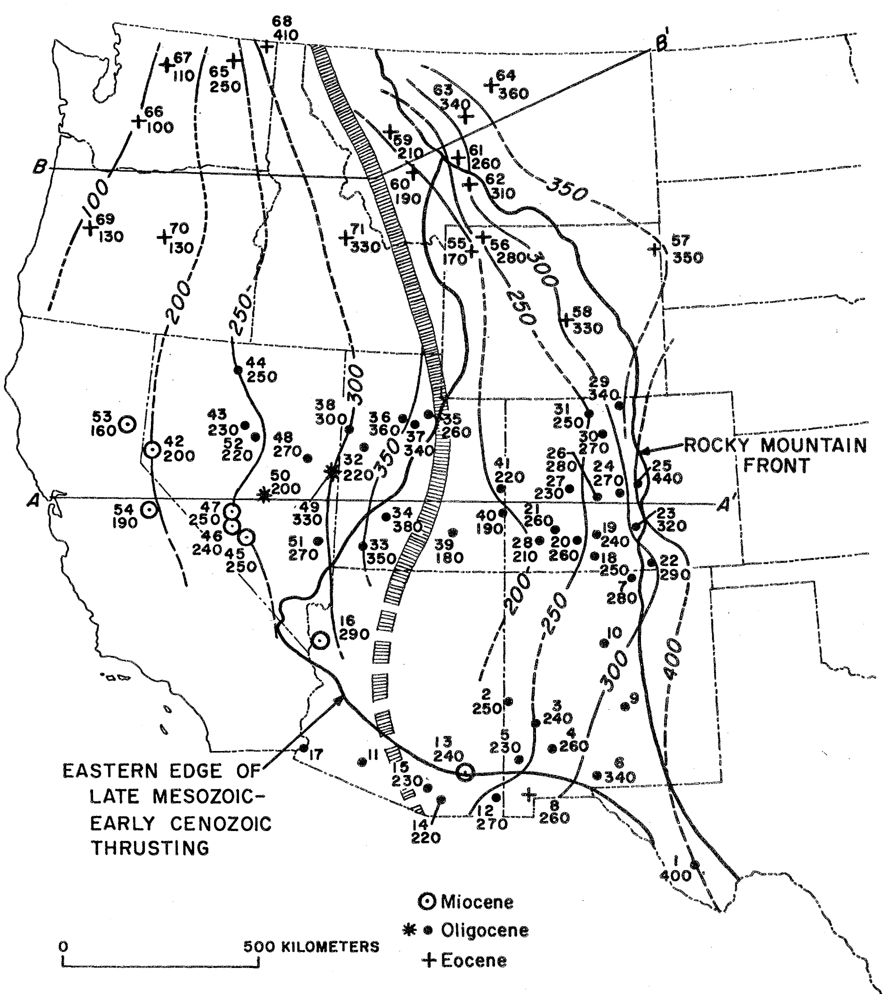


Fig. 2. Inferred early and middle Cenozoic geometry of major tectonic plates and the approximate distribution of continental igneous rocks in the western United States (present distribution, solid black; inferred original extent, stippled pattern). Marine basaltic rocks in the Oregon-Washington coast ranges are not shown. Plate geometry is inferred from magnetic anomaly maps and interpretations of plate motions in the northeast Pacific (*1*) and from our inferences based on the distribution of Cenozoic igneous suites. Double lines indicate spreading rises; single lines, transform faults; and cross-hatched lines, trenches. (a) Distribution of Eocene igneous rocks (40 to 55 million years) and the plate geometry about 50 million years ago. (b) Distribution of Oligocene igneous rocks (25 to 40 million years) and the plate geometry about 30 million years ago.

ping subduction zone (*27*), are plausible consequences of imbrication and flattening of the subduction system.

The imbricate subduction model provides a possible interpretation for otherwise puzzling features of the Colorado Plateau. The plateau was a region of tectonic stability during the deformation and uplift of the surrounding region in latest Cretaceous and early Tertiary time, and also during the basin-range extensional faulting of adjacent areas in late Cenozoic time (*14*). This stable area occurs above shallow parts of the eastern zone of the inferred imbricate subduction system, and it may be an intracontinental analog of the nonvolcanic region of tectonic stability between oceanic trench and andesitic volcanic front that occurs above shallow parts of the Benioff zone in active plate-convergence systems of the Pacific basin (*28*).

The changes in subduction geometry could also explain the pattern of reduced heat flow values, which are believed to indicate middle Cenozoic lower crustal or mantle thermal gradients, in the western United States. Abnormally low Cenozoic heat flow in the Sierra Nevada region, which must have been the locus of high thermal gradients during Mesozoic batholith emplacement, is a plausible consequence of latest Cretaceous flattening of the subduction zone, with cooler shallower parts of the descending plate acting as a heat sink (*29*). Probable low Cenozoic heat flow in the Colorado Plateau province, between belts of high heat flow over deeper parts of the two subduction zones in the Basin and Range



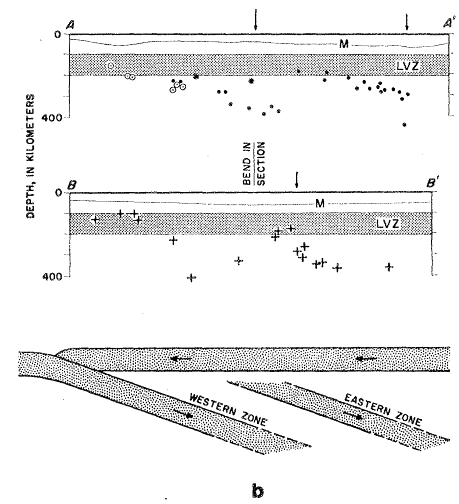


Fig. 3. (a) Contoured depths (in kilometers) to inferred middle Cenozoic subduction zones of western United States. The sources of chemical data for the localities listed below are given in (*4*). The upper number at each point identifies the locality; the lower number gives the depth to subduction zone, obtained from K_2O-depth plots for modern arc volcanics (*8*, *12*). The hachured line indicates discontinuity in the contours. Points for Oligocene rocks in the Cascade Range (Fig. 2b) have not been plotted because the subduction geometry in this region appears to have changed at about the end of the Eocene time, as discussed in the text. (b) Sections showing inferred early and middle Cenozoic subduction zones. The horizontal scale is the same as the vertical scale; surface topography is not shown. All Oligocene (●) and Miocene (⊙) points within 100 km north or south of section *A-A'* are projected into this section; asterisk points (*) are discussed in the text. All points (+) for Eocene igneous rocks in the northwestern United States are projected into section *B-B'*. *M* is the Mohorovicic discontinuity (*33*), *LVZ* is the approximate location of a low-velocity zone (*18*), and the arrows are structural province boundaries from (a). The bottom section is our schematic interpretation of the lithospheric plate geometry suggested by sections *A-A'* and *B-B'*. The localities plotted in (a) are as follows. Texas: 1, Davis Mountains. New Mexico: 2, Blue Range; 3, Gila Wilderness; 4, Black Range area; 5, Santa Rita area; 6, Organ Mountains; 7, southern Sangre de Cristo Mountains; 8, Little Hatchet Mountains; 9, Carrizozo area; 10, Cerrillos area. Arizona: 11, Ajo area; 12, Cochise County; 13, Galiuro Mountains; 14, Santa Rita Mountains; 15, Sierrita Mountains; 16, Oatman district; 17, Yuma area. Colorado: 18, southeast San Juan Mountains; 19, northeast San Juan Mountains; 20, central San Juan Mountains; 21, west San Juan Mountains; 22, Spanish Peaks; 23, Wet Mountain Valley; 24, Thirtynine Mile volcanic field; 25, Cripple Creek; 26, Mount Princeton area; 27, Elk and West Elk Mountains; 28, San Miguel Mountains; 29, Never Summer Range; 30, Middle Park; 31, North Park. Utah: 32, Thomas Range; 33, Iron Springs district; 34, Marysvale area; 35, Park City-Alta area; 36, Bingham area; 37, Tintic area; 38, Gold Hill district; 39, Henry Mountains; 40, Abajo Mountains; 41, La Sal Mountains. Nevada: 42, Virginia City area; 43, Shoshone Range; 44, Osgood Mountains; 45, Cactus Range; 46, Goldfield area; 47, Tonopah area; 48, Eureka area; 49, Ely area; 50 Hot Creek-Pancake Ranges; 51, Delamar district; 52, Roberts Mountains. California: 53, northern Sierra Nevada; 54, central Sierra Nevada. Wyoming: 55, southwest Absaroka volcanic field; 56, northeast Absaroka volcanic field; 57, Black Hills; 58, Rattlesnake Hills. Montana: 59, Garnet Range; 60, Lowland Creek area; 61, Castle and Little Belt Mountains; 62, Crazy Mountains; 63, Highwood Mountains; 64, Bearpaw Mountains. Washington: 65, Republic graben area; 66, Mount Rainier area; 67, northern Cascade Range; 68, Coryell batholith. Oregon: 69, western Cascade Range; 70, John Day basin. Idaho: 71, Challis volcanic field.

and southern Rocky Mountain provinces (*29*), can be similarly related to the shallow part of the inner eastern subduction zone.

Reconstruction of the inferred Pleistocene paleoseismic zone under the Oregon and Washington Cascade volcanic chain (*21*) indicates dips of 40° to 50°. We suggest that the low-angle imbricate Eocene subduction system in the Pacific Northwest (Fig. 3) was replaced in early Oligocene time by a steeper subduction zone that was active until recently. This shift is reflected by the termination of andesitic volcanism in the continental interior (Montana, Wyoming, and Idaho) about 40 million years ago (Fig. 2) and its continuation in the Cascade region through Quaternary time (*30*).

In the western and southwestern United States the low-angle imbricate subduction system operated through Oligocene time, as indicated by the distribution of Oligocene andesitic volcanic rocks in these regions (Fig. 2b). Predominantly calc-alkalic intermediate-composition volcanism terminated in the southwestern United States at about the end of the Oligocene, but it continued in Miocene and Pliocene time in parts of western Nevada and eastern California and through the Quaternary in the Cascade Range (*6*). Initial intersection of the Pacific and American plates (Fig. 2b) probably also occurred at about the end of Oligocene time (*1*). We interpret the enlarging gap in the belt of late Cenozoic andesitic volcanism in western North America as reflecting the growing zone of contact between the American and Pacific plates (Fig. 1), where the intervening plate had been consumed and the trench replaced by a transform boundary system (*1, 6*).

This analysis leaves unresolved problems such as the cause of the abrupt shift in subduction geometry at about the end of Cretaceous time, the persistence of the potentially unstable imbricate geometry over much of the Tertiary, and the periodicity of subduction-related igneous activity—with well-defined Laramide and middle Tertiary peaks. We suspect that these features reflect factors such as changes of subduction rate with time, velocity differences between the two imbricate subduction zones, mobility of the asthenosphere between the two zones, and changes in absolute motion of the main American plate relative to the underlying asthenosphere. In particular, analysis of plate motions with respect to mantle "hot-spots" (*31*) suggests that the American plate moved westward especially rapidly from about 80 to 40 million years ago (*32*). A rapidly moving American plate could have recurrently overridden and entrapped gently dipping lobes of Farallon plate, which would have become attenuated and separated from the main Farallon plate and could then descend slowly, because of density differences, into the asthenosphere, as the main subduction boundary was reestablished farther west.

PETER W. LIPMAN
HAROLD J. PROSTKA
ROBERT L. CHRISTIANSEN
United States Geological Survey, Federal Center, Denver, Colorado 80225

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Precambrian Columnar Stromatolite Diversity: Reflection of Metazoan Appearance

Abstract. *Columnar stromatolites (organosedimentary structures built by blue-green algae) show a marked decrease in diversity in the Late Precambrian; this decrease in diversity occurs at approximately the same time as the appearance of metazoans, 600 to 700 million years ago.*

Stromatolites constitute the most abundant fossils in the Precambrian. Their widespread occurrence in carbonates continues into the early Lower Paleozoic, but by Middle Ordovician time they begin to decline (*1*). Garrett (*2*) attributes this decline in the Phanerozoic to the evolution and diversification of grazing and burrowing metazoans. In effect, these animals restricted stromatolites to one or a combination of the following environments: (i) intertidal-supratidal zones, (ii) hypersaline regions, and (iii) environments where strong current and high sediment movement exclude burrowers and grazers (*2*).

Before the rise of metazoans, the only ecologic restrictions on stromatolite growth probably were (i) depth of