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EVOLUTION OF A LATE MESOZOIC BACK-ARC FOLD AND THRUST BELT, NORTHWESTERN GREAT BASIN, U.S.A.

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ABSTRACT

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The Luning-Fencemaker fold and thrust belt was active from the Middle or Late Jurassic through the Early Cretaceous and involves rocks of the Mesozoic marine-province of the northwestern Great Basin. Rocks of the marine-province were deposited in a back-arc basin bound on the west by the Sierran arc. They are underlain by simatic crust formed either in a Paleozoic marginal-basin formed on the west coast of North America or in an exotic oceanic-arc accreted to North America in the Permo-Triassic. Deposition in the marine-province was localized by the underlying simatic crust, which later controlled the locus of back-arc thrusting. To the south, in areas of the back-arc region overlying continental crust, Mesozoic rocks are terrestrial volcanic and volcanogenic deposits. Rocks of this area are not disrupted by thrusts of significant magnitude until the southwestern end of the Sevier thrust belt is encountered.

During contraction, the Luning-Fencemaker thrust belt underwent several hundred kilometers of NW–SE shortening which was not developed in the Sierra Nevada to the west. The arc and back-arc regions were separated by a regionally extensive left-lateral fault system that was active during back-arc thrusting. The marine-province was partially coupled with the Sierra Nevada during transpressional motion on the left-lateral fault, resulting in imbrication and emplacement of the Luning-Fencemaker thrust belt. In the continental region to the south, decoupling was more complete and only minor thrusting was associated with the southerly displacement of the Sierran arc. Farther south, where the Paleozoic miogeoclinal hingeline was encountered, conditions apparently changed and the Sevier thrust belt developed. Timing relations suggests that the Andean-type margin of the western United States had paired back-arc thrust belts (Luning-Fencemaker and Sevier) for much of the late Mesozoic. The belts were separated by a region that underwent relatively minor internal shortening. The intervening region probably was decoupled at depth and the thrust belts may have shared a décollement.

Development of the Luning-Fencemaker thrust belt and the associated left-lateral fault system extended over a period of 40–50 m.y. and is best explained as a product of plate convergence. The NW–SE shortening axis for the thrust belt and the inferred left-slip on the strike-slip fault are not compatible with the generally accepted right-oblique subduction proposed for the Sierran arc since the Early Triassic. They suggest an interval of left-oblique convergence for the Middle to Late Jurassic and Early Cretaceous. Regionally developed structures indicate a return to right-oblique convergence in the late Early Cretaceous.

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INTRODUCTION

The Mesozoic marine-province of the northwestern Great Basin consists of basinal, volcanic arc, and continental shelf rocks deposited in and around a back-arc marginal basin lying east of the Mesozoic Sierran arc. The Mesozoic succession ranges in age from late Early Triassic through Early Cretaceous and is structurally dismembered and divisible into several discrete lithotectonic assemblages. Regional stratigraphic relations, however, allow the entire succession to be tied to North America during its deposition. Rocks of the Mesozoic marine-province generally have no exposed lower depositional contact, but where depositional contacts are observed, the Mesozoic units overlie various Paleozoic successions with angular unconformity. The origin and history of the underlying Paleozoic rocks are controversial as are the origin and tectonic significance of the crust underlying the region. The western edge of sialic North America, created by rifting in the Precambrian (Stewart, 1972) or possibly later (Kistler, 1978), is generally accepted as lying in central and western Nevada. The location of the continental margin is estimated on stratigraphic grounds (Roberts et al., 1958) and more recently by geophysical data (Cogbill, 1979) and the Sr 0.706 (initial $^{87}Sr/^{86}Sr$) contour (Kistler and Peterman, 1973; Kistler, 1978; Kistler et al., 1981; Leeman, 1982). West of the inferred sialic margin, the crust is simatic and formed either in a marginal basin associated with North America (Burchfiel and Davis, 1972, 1975) or in an exotic, oceanic-arc accreted to North America in the Permo-Triassic (Speed, 1979).

The eastern and southern margins of the marine-province roughly coincide with the proposed join between crustal types demarked by the Sr = 0.706 line. The locus of marine deposition was apparently controlled by subsidence of the underlying simatic crust (Speed, 1978a). Farther south, in areas underlain by continental crust, marine conditions did not exist and post-Lower Triassic rocks are dominated by subaerial volcanic and volcanogenic deposits (Dunne et al., 1978).

There is little doubt that the distribution and mechanical properties of the simatic basement fundamentally controlled marine deposition in the Sierran back-arc region, and data presented here strongly suggest that it also localized the development and areal extent of a back-arc fold and thrust belt active from the Middle or Late Jurassic through the Early Cretaceous. The configuration of the simatic basin had a significant impact on the structural history of the region. In west-central Nevada, the change from a southerly- to a westerly-trending boundary between sialic and simatic basement (Fig. 1) is generally cited as tectonic in origin and the product of oroclinal bending (Albers, 1967) or crustal flexure (Wetterauer, 1977). However, recently recognized relations in rocks of the mid-Paleozoic Roberts Mountains allochthon (Oldow, 1984a) and paleomagnetic investigation of Mesozoic rocks (Oldow and Geissman, 1982) around the bend support an earlier interpretation of Ferguson and Muller (1949) that it is not the product of rotation. Rather, the irregular shape of the boundary is a preserved feature formed during rifting which has profoundly affected subsequent deposition and deformation.

LITHOTECTONIC ASSEMBLAGES

Mesozoic rocks of the marine province are divided into eight lithotectonic assemblages (Fig. 2). Five dismembered assemblages, each composed of numerous thrust sheets, are recognized and are bound on the east and south by two autochthonous or para-autochthonous successions which were subjected to relatively minor thrusting. On the west, an assemblage which has not undergone significant

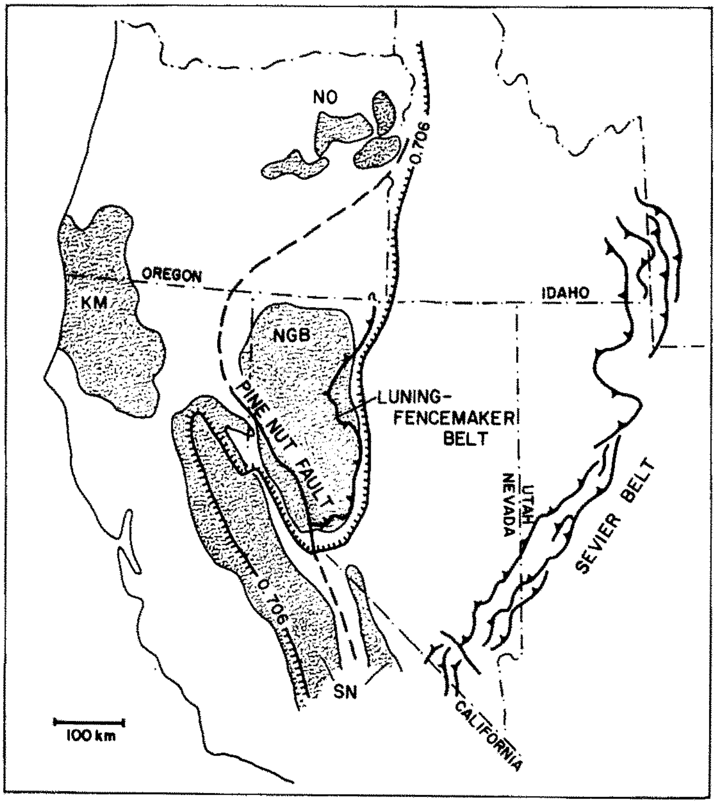


Fig. 1. Generalized map of part of the western United States illustrating major tectonic features. Shaded areas are: northeastern Oregon (*NO*): the Klamath Mountains (*KM*): the Sierra Nevada (*SN*); and the Mesozoic marine-province of the northwestern Great Basin (*NGB*). The hachured line represents the approximate western limit of sialic crust deduced from the initial $^{87}Sr/^{86}Sr = 0.706$ contour. Bold lines are known or inferred faults: the Pine Nut fault is an inferred strike-slip fault system formed in the Jura–Cretaceous; and the Luning-Fencemaker Belt and Sevier Belt are thrust fault systems (teeth on upper plate).

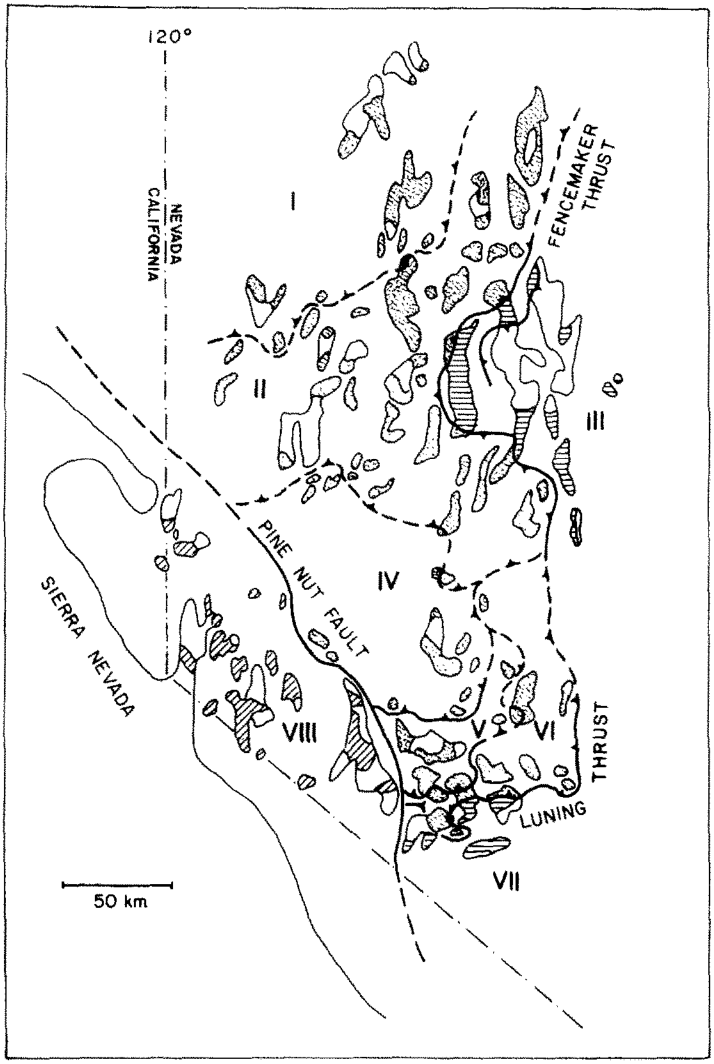


Fig. 2. Generalized map of pre-Tertiary rocks of the northwestern Great Basin. Bold lines are known faults, dashed where inferred; teeth are on the upper plate of thrust faults. Roman numerals delineate lithotectonic assemblages: *I*—Black Rock assemblage; *II*—Lovelock assemblage; *III*—Humboldt assemblage; *IV*—Sand Springs assemblage; *V*—Pamlico assemblage; *VI*—Luning assemblage; *VII*—Gold Range assemblage; and *VIII*—Pine Nut assemblage. Shaded areas represent Mesozoic and Paleozoic rocks constituting dismembered lithotectonic assemblages, horizontally ruled areas represent Mesozoic successions of autochthonous or para-autochthonous assemblages, and diagonally ruled areas are Mesozoic exposures of an internally coherent allochthonous assemblage (*VIII*).

thrust imbrication is recognized and is interpreted as being displaced with respect to the para-autochtonous assemblages to the east. The term lithotectonic assemblage is used here simply to segregate coeval or partly coeval packages of rocks which accumulated in separate, though commonly related, settings and/or which had different structural histories. The voluminous formational nomenclature of the region is omitted in the brief descriptions below. A more detailed rendering of the lithologies and their distributions is given by Speed (1978a).

Black Rock assemblage (I)

Rocks of the Black Rock assemblage (Fig. 2) have been studied only locally and the relations among exposures in several mountain ranges are poorly understood. Nevertheless, available data indicate that it is a complexly deformed pile of thrust sheets composed of Mesozoic and Paleozoic rocks. Few depositional contacts between the Mesozoic and Paleozoic successions are known, but basinal Paleozoic rocks appear to form the substratum for Mesozoic deposition (Russell, 1981, 1984). Mesozoic rocks (Fig. 3) are basin and basin-margin deposits of Late Triassic and Jurassic age that become progressively coarser and more volcanogenic upsection until they are conformably overlain by a constructional magmatic-arc. The arc consists of intermediate volcanic and plutonic rocks dated by whole rock Rb/Sr at 164 ± 35 m.y. B.P., but it apparently began erupting earlier, in the Late Triassic (Russell, 1981). Disconformably overlying the arc sequence are coarse clastic rocks which grade upward into fine-grained clastic rocks and freshwater limestones containing an Early Cretaceous fauna (Willden, 1958, 1964). The clastic rocks are derived from underlying assemblages and exhibit variability in thickness suggestive of local uplift during deposition.

Where studied (Russell, 1981, 1984), Mesozoic structures consist of an early set of weak, northwesterly trending flexures which locally controlled deposition of the magmatic-arc rocks, and a pervasive set of second-phase folds (Fig. 4). Second folds are tight to isoclinal with NE-striking axial planes and axial-plane cleavage and were formed during and after deposition of the post-arc clastic succession. Major thrusts post-date and possibly were synchronous with development of NE-trending folds and were caused by regional NW–SE compression in the Late Jurassic and Early Cretaceous. Earlier NW-trending folds do not reflect significant shortening and formed between the Late Triassic and Late Jurassic. All structures pre-date intrusion of Late Cretaceous Plutons.

Lovelock Assemblage (II)

A thick succession of Triassic basinal rocks, locally overlain by Jurassic shallow marine and subaerial sedimentary rocks, forms a northeasterly trending belt in central western Nevada (Fig. 2). Nowhere is a lower depositional contact exposed; to the east a tectonic lower contact, which is part of the regionally extensive Fencemaker thrust (Speed, 1978a), is occasionally exposed. The basinal rocks grade upward into interbedded carbonate and clastic rocks of slope or outer shelf facies

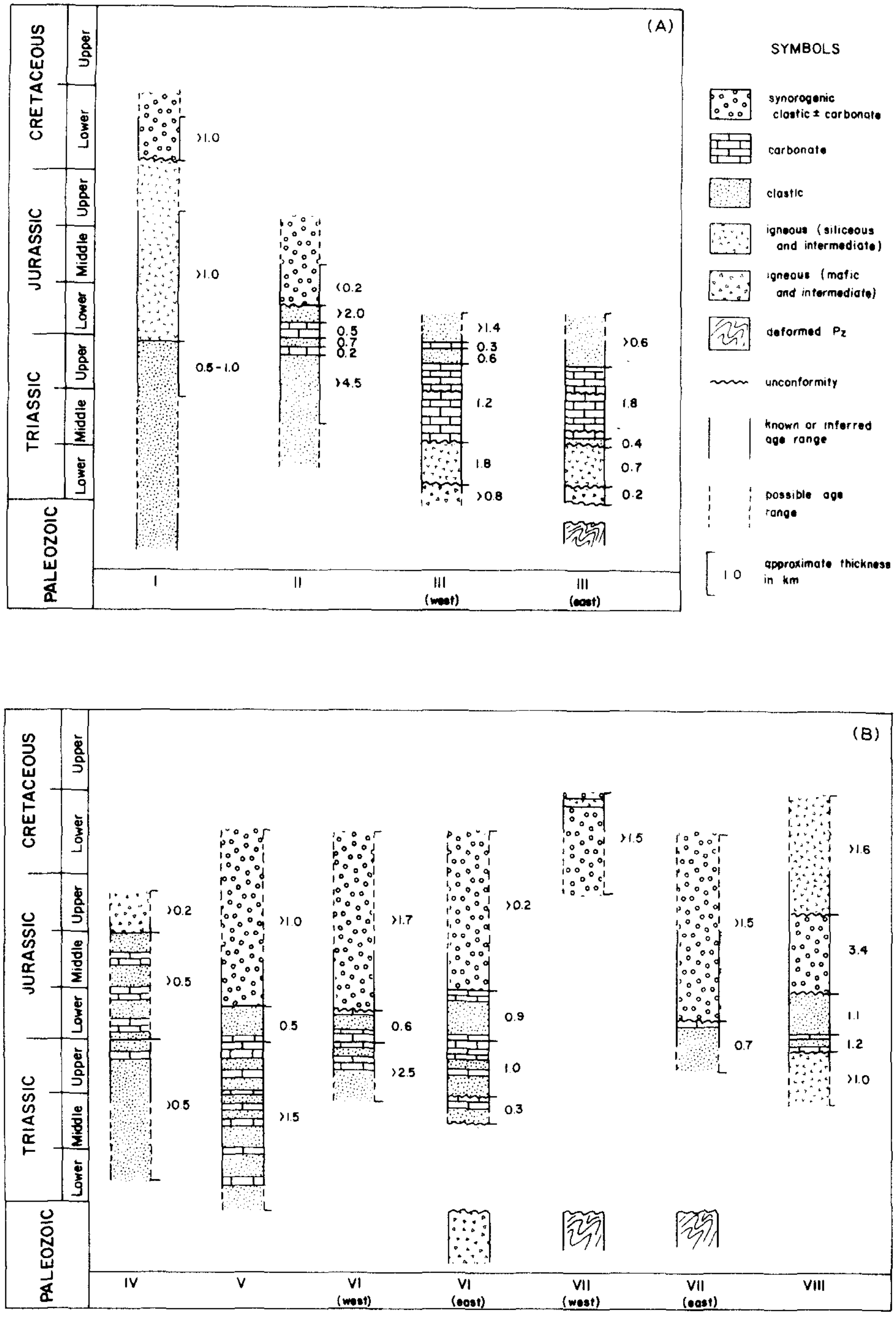


Fig. 3. Composite time-stratigraphic columns for the lithotectonic assemblages delineated in Fig. 2. Data sources are compiled in Table I.

TABLE I

Source of data for composite time-stratigraphic columns depicted in Fig. 3.

Lithotectonic assemblage	Source of data	Lithotectonic assemblage	Source of data
I	Russell (1981, 1984)	V	Oldow (1978, 1984b,
	Willden (1963, 1964)		unpublished data)
			Oldow and Steuer (1984)
II	Burke and Silberling (1973)		Oldow and Speed (1984)
	Compton (1960)		Steuer (1978)
	Johnson (1977)		
	Oldow and Speed (1974)	VI	Ferguson and Muller (1949)
	Oldow (unpublished data)		Mottern (1962)
	Speed (1974, 1978a, 1978b)		Oldow (1981b)
	Willden (1964)		Oldow and Dockery (in prep.)
	Willden and Speed (1974)		Oldow and Meinwald (in prep.)
			Silberling (1959, 1981,
III	Burke and Silberling (1973)		written communication, 1982)
	Johnson (1977)		Wetterauer (1977)
	Nichols and Silberling (1977)		
	Silberling and Wallace, (1967)	VII	Ferguson and Muller (1949)
	Willden and Speed (1974)		Garside (1979)
			Oldow (unpublished data)
IV	Banazak (1969)		Speed (1977b)
	Ekren and Byers (1978)		Speed and Kistler (1980)
	Hardyman (1980)		Wetterauer (1977)
	Oldow and Hardyman (in prep.)		
	Silberling (1981)	VIII	Bingler (1977, 1978)
	Willden and Speed (1974)		Hudson and Oriel (1979)
			Noble (1962)
			Oldow (unpublished data)
			Silberling (1981)

(Oldow and Speed, 1974; Speed, 1978b) which in many areas are the highest stratigraphic horizons preserved (Fig. 3). The distribution of preserved younger rocks is restricted as apparently was the original lateral continuity of many depositional environments. Early Jurassic rocks consist of shallow marine clastics and locally of interbedded carbonate–gypsum deposits accumulated in a restricted shallow-marine basin (Speed, 1974). An areally restricted sequence of quartz arenite and locally derived conglomerate overlies some basinal rocks disconformably; it is interpreted as a synorogenic deposit associated with mid-Jurassic plutonism and volcanism (Speed and Jones, 1969).

The rocks are deformed in polyphase folds and dismembered by numerous thrust faults. First-generation structures are tight to isoclinal NNE- to NE-trending folds developed during NW–SE regional compression that resulted in southeasterly di-

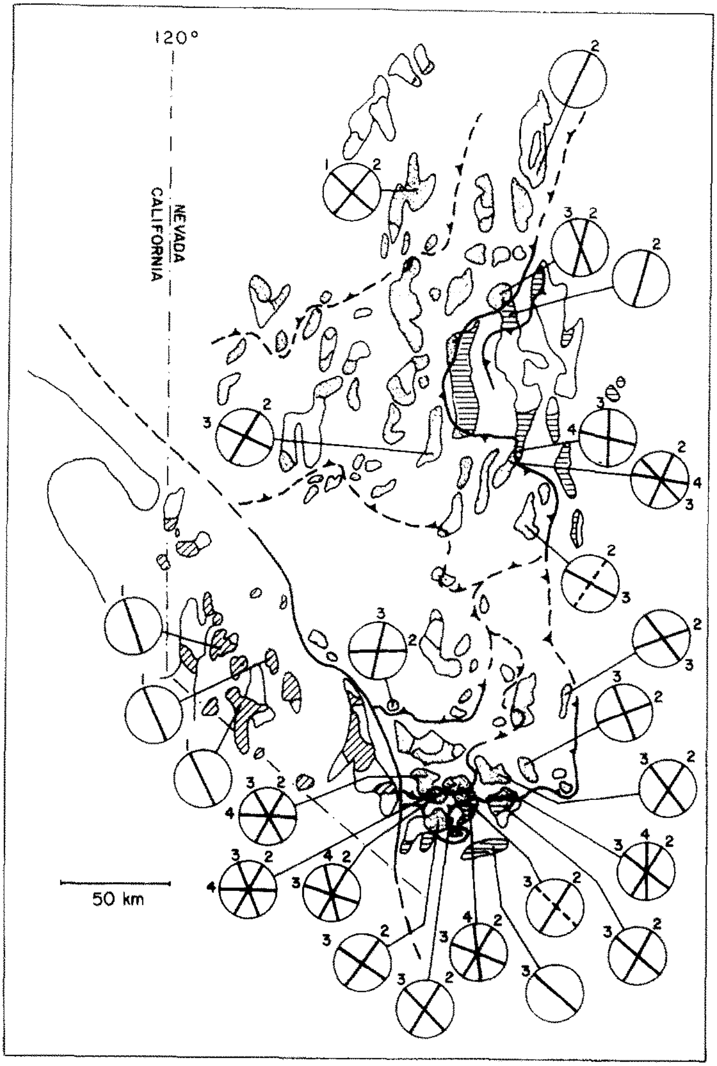


Fig. 4. Composite structure-diagrams of the orientation of folds, D_1-D_4, discussed in the text. Bold lines correspond to the mean orientation of axial planes of folds. Numerals *1–4* refer to phases of deformation D_1-D_4, respectively.

rected thrusting of basinal rocks over coeval platform assemblages. Locally, some platform rocks are incorporated into the allochthonous succession and lie above the Fencemaker thrust (Speed et al., 1982). Later folding varied in intensity (open to tight) and orientation. Nevertheless, a generalized sequential development of northwesterly-trending folds followed by E–W-trending folds is recognized, at least locally (Fig. 4). The upper age limit of deformation is constrained by Late Cretaceous plutons, and the age of inception of NE-trending folds and associated thrusts

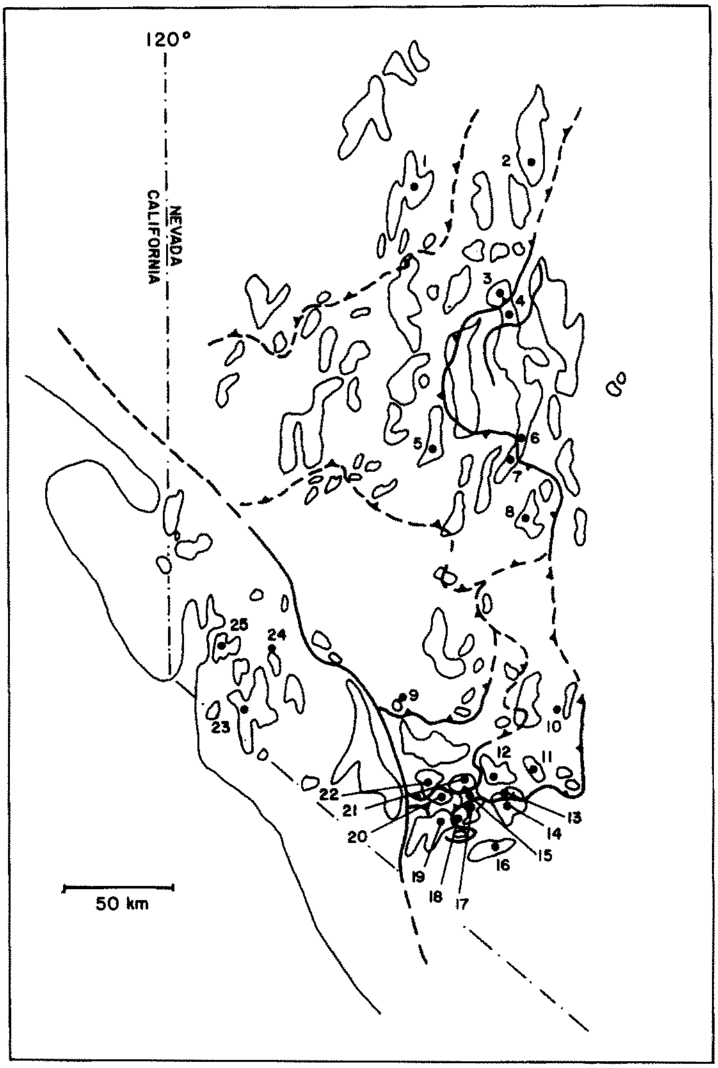


Fig. 5. Location map of structural data depicted in Fig. 4 and compiled in Table II.

is based on a syntectonic pluton dated by K-Ar on a hornblende–biotite pair and biotite at 165 ± 5, 145 ± 5, and 145 ± 5 m.y. B.P., respectively (Willden and Speed, 1974). In a few areas, additional thrusting associated with the development of the NW-trending folds contributed to the shelfward transport of the basinal rocks.

Humboldt assemblage (III)

In the Humboldt assemblage, an early Mesozoic platform succession depositionally overlies upper Paleozoic rocks and borders the allochthonous basinal rocks to

TABLE II

Compilation of sources of structural data; locations keyed to Fig. 5

Location	Source of data
1	Russell (1981, 1984)
2	Weins et al. (1982)
3	Elison et al. (1982)
4	Elison et al. (1982)
5	Speed (1974)
6	C.H. Dodge (written communication, 1975) Oldow and Speed (1974)
7	Oldow (unpublished data) Oldow and Speed (1974)
8	Speed (1978b) Speed and Jones (1969)
9	Oldow and Hardymann (in prep.)
10	Oldow (unpublished data) Wetterauer (1977)
11	Mottern (1962) Wetterauer (1977) S. Brown (unpublished data) Oldow (unpublished data)
12	Ferguson and Muller (1949) Oldow (unpublished data) Oldow and Dockery (in prep.) Wetterauer (1977)
13	Oldow (1981a, 1981b) Oldow and Dockery (in prep.) Oldow and Meinwald (in prep.)
14	Oldow (unpublished data) Wetterauer (1977)
15	Oldow (in prep.) Oldow and Speed (1984)
16	Oldow (1984a) Speed (written commun., 1981)
17	Garside, 1979 Oldow (unpublished data) Speed (1977b)
18	Bowen (1982) Bowen and Oldow (1980) Oldow (unpublished data)
19	Bowen (1982) Bowen and Oldow (1980) Oldow (unpublished data)
20	Oldow (1984b) Oldow and Steuer (1984)
21	Oldow (1978) Oldow and Steuer (1978, 1984) Oldow and Speed (1984)
22	Oldow (1978, in prep.)
23	Noble (1962) Oldow (unpublished data)
24	Hudson and Oriel (1979) Oldow (unpublished data)
25	Bingler (1977) Oldow (unpublished data)

the west (Fig. 2). Platform carbonates dominate the lower section and overlie a thick pile of siliceous volcanic rocks which unconformably overlie greenstones of probable Permo-Triassic age (Speed, 1978a; Johnson, 1977). The greenstones rest on highly deformed upper Paleozoic rocks and may be related to the emplacement of the Golconda allochthon during the earlier Sonoma orogeny (Speed, 1978a). The siliceous volcanic rocks are of probable Early Triassic age and are overlain in the east by clastic and volcanogenic rocks grading upward into carbonates (Fig. 3). In the west, on the other hand, they are directly overlain by carbonate rocks. The platform carbonates persisted from Early Triassic (Spathian) through Late Triassic (Karnian), and record deposition in shallow marine to intertidal environments characterized by significant differential uplift through time (Nichols and Silberling, 1977). Carbonate deposition was interrupted by at least two pulses of clastic sedimentation which occurred in the Late Triassic (Norian) (N.J. Silberling, pers. commun., 1983).

Rocks are deformed in two and locally three phases of folds (Fig. 4). Tight to isoclinal first-folds are developed in areas overlain by regional thrusts and have NNE- to NE-striking axial planes. In the west, the folds are associated with southeasterly directed thrusts of the Fencemaker system carrying basinal rocks over the platform, whereas in the east and central portions of the assemblage, Mesozoic rocks are overthrust by Paleozoic rocks (N.J. Silberling, pers. commun., 1982; Speed et al., 1982). The Paleozoic thrust sheets were emplaced toward the northwest prior to intrusion of a 150 m.y. B.P. pluton (D. Whitehead, personal commun., 1982). The northwest thrusting apparently pre-dates emplacement of the upper-plate of the Fencemaker thrust (R.C. Speed, personal commun., 1982) and both thrust systems are related to NW–SE regional compression. Additional shortening followed the major thrusting, resulting in locally developed northwesterly-trending second-folds and E–W-trending third-folds. The age of inception of first-generation structures is not well constrained in the Humboldt assemblage but post-dates the deposition of Upper Triassic rocks and apparently predates emplacement of the 150 m.y. B.P. pluton. The minimum age of folds is constrained by post-tectonic Cretaceous plutons.

Sand Springs assemblage (IV)

Stratigraphic and structural relations in the Sand Springs assemblage are poorly understood (Fig. 2), but the stratigraphic framework that emerges suggests basinal deposition supplied by a volcanic source to the west. The site of accumulation shoaled with time, resulting in deposition of interbedded volcanic and volcanogenic rocks and shallow-marine carbonates. No lower contact is known. Where sufficient structural control exists, the succession is seen to consist of numerous thrust sheets. No Paleozoic rocks are recognized, but numerous successions are undated and could be at least partially Paleozoic. The presumably oldest Mesozoic rocks are volcanogenic and carbonate turbidites interbedded with mudstone which grade upward into interbedded basinal carbonates and volcanogenic rocks containing Late Triassic faunas (Fig. 3). Elsewhere, interbedded carbonate, volcanic, and volcanogenic rocks are assigned an Early to Middle Jurassic age (Banazak, 1969; Willden and Speed, 1974) and represent relatively shallow-marine to subaerial deposition.

In the Sand Springs assemblage, structural relations are locally complicated by later, Cenozoic deformation. Nevertheless, where recognized, isoclinal first-folds have ENE-striking axial planes (Fig. 4) and are interpreted to be related to major NW–SE thrusting (Oldow and Hardyman, in prep.). Later folds are open to tight and have spatially variable axial planes striking from NE to NW. The age of deformation is constrained between Early Jurassic and 80 m.y. B.P. on the basis of youngest rocks deformed and cross-cutting plutons.

Pamlico assemblage (V)

The Pamlico assemblage (Fig. 2) is a complexly deformed stack of imbricate thrust sheets composed exclusively of either Mesozoic or Paleozoic rocks. No depositional contacts exist between Mesozoic and Paleozoic successions. The Paleozoic rocks constitute a small number of the thrust sheets and are parts of a structurally dismembered Permian volcanic-arc (Speed, 1977a). The oldest known Mesozoic rocks consist of an Early(?) to Late Triassic volcanic, volcanogenic, and carbonate sequence deposited in and around a volcanic archipelago (Oldow, 1978). The uppermost part of the volcano-carbonate sequence is a regionally extensive shelf carbonate conformably overlain by distal shelf clastic and carbonate rocks (Fig. 4). Unlike the archipelago clastics, which are derived entirely from a volcanic source, the shelf clastics are largely derived from a continental source to the east. Conformably overlying the shelf rocks are quartz arenites and poorly sorted, coarse-clastic rocks deposited in areally restricted environments supplied with sediments derived from local and continental sources. The quartz-rich coarse-clastic rocks are faunally dated as Early Jurassic (Hettangian–Pliensbachian) and grade upward into volcanogenic sedimentary and volcanic rocks. The volcanic rocks, increasing in abundance toward their inferred westerly source, are intermediate to siliceous lavas and breccias.

Polyphase folding (Fig. 4) and thrust imbrication are constrained between Early Jurassic and 100 m.y. B.P. (Oldow, 1978). First folds have NE-striking axial planes and axial-plane cleavage, are tight to isoclinal, and were formed contemporaneously with thrust transport and imbrication during major NW–SE shortening. Thrusting resulted in displacement of tens of kilometers toward the southeast (Oldow, 1976). Later folds, accompanied by minor thrusting, have spatially variable orientations. Trends of second folds range from NNW to WNW, whereas two spatially segregated sets of third folds with N–S and E–W axial planes are formed. The variability in orientations of second and third folds is a primary feature and not due to later deformation (Oldow, in prep.).

Luning assemblage (VI)

The Luning assemblage (Fig. 2) is underlain by the regionally extensive Luning thrust (Oldow, 1975, 1981a, 1981b; Speed, 1977a, b) and lies structurally below the Pamlico assemblage. Permian volcanic-arc rocks and other Paleozoic(?) clastics occur in small exotic thrust-slices in the predominantly Mesozoic assemblage; few depositional contacts between Mesozoic and known Paleozoic rocks are recognized. The lower Mesozoic sedimentary rocks (Fig. 3) were deposited in a continental shelf environment and in one locality overlie volcanic and volcanogenic rocks of possible Pennsylvanian or Permian age. The shelf sequence consists of platform carbonates and shallow-marine to deltaic-clastic rocks derived from continental sources. Minor amounts of volcanogenic rocks are interbedded with terrigenous-clastic rocks near the western margin of the terrane (Reilly et al., 1980). The upper part of the clastic-carbonate succession is a regionally extensive carbonate which is conformably overlain by distal shelf clastics and carbonates, like those in the Pamlico assemblage. These are conformably overlain by Lower to Middle Jurassic (Pliensbachian) quartz arenite and coarse-clastic rocks which grade upward into volcanogenic rocks.

Rocks of the Luning assemblage are involved in a complex deformational history with two and locally three phases of folds (Fig. 4) and significant stratal shortening by thrusting that is constrained between Middle Jurassic and 90 m.y. B.P. (Oldow, 1981a, 1981b). First folds, with NE-striking axial planes and axial-plane cleavage, are uniformly developed tight to isoclinal structures; they are associated with major NW–SE thrusting (Oldow, 1981a, 1981b; Seidensticker et al., 1982). Second folds vary spatially in degree of development, ranging from gentle to isoclinal, and in orientation, with axial planes striking from NNW to WNW. Open third-folds with both N–S- and E–W-striking axial planes are sporadically developed and found only where second folds are intensely developed.

Gold Range Assemblage (VII)

The Gold Range assemblage (Fig. 2) is composed predominantly of interbedded Mesozoic clastic and volcanic rocks overlying a Permian basinal assemblage with angular unconformity. The oldest Mesozoic rocks are interbedded, subaerial and shallow-marine terrigenous clastic, volcanoclastic, and minor carbonate rocks overlain by shelf carbonates containing Lower Jurassic pelecypods. Unfossiliferous quartz arenite and coarse-clastic rocks disconformably overlie the shelf carbonate and grade upward into poorly sorted volcanogenic sandstone and coarse-clastic rocks (Fig. 3). Locally, siliceous volcanic rocks have yielded a whole rock Rb/Sr date of 147 ± 17 m.y. B.P. (Speed and Kistler, 1980), and elsewhere, intercalated clastic and siliceous volcanic rocks dated by Rb/Sr at 103 ± 5 m.y. B.P. (Speed and Kistler, 1980) overlie Permian rocks with angular unconformity. Although some stratigraphic relations are as yet unclear, overall they suggest localized, probably diachronous, deposition in a tectonically active environment.

The rocks are locally disturbed by thrust faults of small displacement associated with NE-trending folds (Fig. 4). The NE-trending folds are most intense immediately below the overlying Luning thrust; downwards they become less tight and disappear within a few kilometers from the thrust (Oldow, 1981a, 1981b). Second structures are open to close NW-trending folds developed uniformly throughout the lithotectonic assemblage. Timing of deformation is constrained by Cretaceous (103 ± 5 m.y. B.P.) rocks folded in NE- and NW-trending structures which are truncated by undeformed 90 m.y. B.P. plutons (Garside and Silberman, 1978; Speed and Kistler, 1980).

Pine Nut assemblage (VIII)

The Pine Nut assemblage has a linear northwesterly trending boundary which lies at a high angle to the regional trends to the east (Fig. 2). The western boundary is less clear but appears to parallel the Sierra Nevada. No known Paleozoic rocks are exposed and no depositional lower-contact is exposed for the lowest Mesozoic unit (Fig. 3), which consists of intermediate volcanic rocks dated by Rb/Sr at 215 m.y. B.P. (Einaudi, 1977). Above the volcanic succession lie Late Triassic (Norian) volcanic, volcanogenic sedimentary, and carbonate rocks accumulated in shallow marine to subaerial environments. They are overlain by fine-grained clastic rocks deposited in a distal-shelf setting. The clastics contain a Lower Jurassic (Toarcian) fauna near their top where they grade into siliceous volcanic rocks. Disconformably overlying the shelf succession are unfossiliferous quartz arenites and siltstones which themselves are disconformably(?) overlain by a thick succession of intermediate to siliceous volcanic flows, breccias, and volcanogenic sedimentary rocks.

Structurally (Fig. 4) the rocks are involved in a single phase of tight to isoclinal folds with NNW-striking axial planes (Noble, 1962; Oldow, unpublished data). No major thrust faults are known, and the lateral continuity of the stratigraphy throughout the assemblage argues against significant structural disruption. The minimum age of deformation is equivocal because the age and structural relation of overlying volcanic rocks is not established. Known lower and upper limits are Early Jurassic and 90 m.y. B.P.

REGIONAL RELATIONS

Coeval rocks of different depositional settings are juxtaposed along lithotectonic assemblage boundaries, and at several locations the boundaries are known thrust faults, whereas at others the existence of a fault is inferred. Dismembered assemblages, composed of numerous thrust sheets, are bound on the east and south by the Luning-Fencemaker thrust system and on the west by the northwesterly-trending eastern-boundary of the Pine Nut assemblage (Fig. 2). The Fencemaker thrust forms the western boundary of the platformal Humboldt assemblage (Oldow and Speed, 1974; Speed, 1978a; Speed et al., 1982) and is interpreted to extend south and merge with the Luning thrust (Oldow, 1975, 1978, 1981a, 1981b, 1983; Speed, 1977a, b), which defines the northern boundary of the Gold Range assemblage. These thrusts are generally suballuvial and are exposed only as isolated segments in several mountain ranges, but abrupt changes in regional facies patterns call for their lateral continuity. The eastern boundary of the Pine Nut assemblage is interpreted as a fault, with significant strike-slip motion, and its location is constrained by the western termination of the Gold Range, Luning, Pamlico, and Sand Springs assemblages. In addition to the juxtaposition of rocks of various depositional settings across this boundary, it also separates regions with significantly different structural histories (Fig. 4).

The structural history of the rocks east of the Pine Nut assemblage is remarkably similar with regard to orientation, sequential development, and timing. Timing relations are not established with equal certainty everywhere, but the correspondence in orientation and sequential development of structures allow several regional generalizations. The earliest known structures (D_1) are broad NW-trending flexures developed during Jura–Triassic volcanism in the Black Rock assemblage but are important only locally where they controlled deposition. They are not found elsewhere in the region and are not associated with significant shortening. In the dismembered assemblages, the first regionally extensive deformation (D_2) results in great NW–SE shortening and is expressed by NE-trending folds and associated thrusts formed during imbrication and emplacement of the Luning-Fencemaker thrust system. In the relatively intact Humboldt and Gold Range assemblages, D_2 folds are found only in the vicinity of major thrust faults. Rocks of these two assemblages overlie Paleozoic rocks with angular unconformity and are autochthonous or para-autochthonous with respect to the suprajacent dismembered assemblages.

Later NW-trending folds (D_3) in the autochthonous or para-authochthonous rocks exhibit little spatial variability, but in the overlying allochthonous successions, a wide variation in orientation and intensity is observed. In the authochthon, the NW-trending folds (D_3) are found superimposed on NE-trending folds (D_2) near regional thrusts, but they extend into areas where D_2 folds are not developed. The variability of D_3 folds in the allochthonous terranes is demonstrably a primary feature in several areas (Oldow, 1981a, 1981b, unpubl. data) and not due to later deformation.

D_4 folds occur locally and have either N–S- or E–W-striking axial planes. N–S- and E–W-trending D_4 folds are associated with only minor shortening and are never found superimposed. They are spatially segregated, commonly within the same mountain range, and probably developed contemporaneously.

The relative ages of folding are well established, but of more importance are the absolute ages of inception and cessation of the various folding events. The earliest folds recognized (D_1) are developed during deposition of arc rocks in the Black Rock assemblage and are constrained as Late Triassic to Late Jurassic in age (Russell, 1981). Later NE-trending folds and associated thrusts (D_2) are overprinted by younger NW-trending folds and thrusts (D_3), which in turn are followed by N–S-trending and E–W-trending folds (D_4) of the final stage of deformation. The upper age limit for D_2 and D_3 folds is established regionally as pre-Late Cretaceous (100–90 m.y. B.P.) by cross-cutting plutons. Later D_4 folds may also pre-date plutonism, but due to their local development and slight limb appression, their relationship to plutons is unclear. The age of inception and the duration of D_2 and D_3 folds are more difficult to establish, but a regional pattern appears to exist. NE-trending folds and associated thrusts (D_2) began, at least locally, during the Jurassic (Speed, 1974). In the Lovelock assemblage, a syntectonic pluton associated with D_2 deformation is dated by K-Ar as between 165 to 145 m.y. B.P. (Speed, 1974; Willden and Speed, 1974) suggesting that the onset of deformation occurred during or before Middle or Late Jurassic. In the same area, Lower to Middle Jurassic rocks were deposited before plutonism and deformation. Elsewhere, in the Humboldt assemblage, early stages of NW–SE compression resulted in emplacement of thrust sheets of Paleozoic rocks over Mesozoic platform rocks before intrusion of a 150 m.y. old pluton. The areal extent of Jurassic age D_2 deformation is unknown, but the longevity of the event is bracketed by deformed Cretaceous rocks in the Black Rock and Gold Range assemblages. The Cretaceous rocks in both areas were deposited in tectonically active settings, and they are probably analogous to syntectonic clastic rocks incorporated in other fold and thrust belts. Third-phase (D_3) NW-trending folds are locally (Gold Range assemblage) constrained as post- 103 ± 5 and pre-90 m.y. B.P. by the youngest rocks deformed and cross-cutting plutons. A controversy concerning the timing of D_2 and D_3 folds exists, however, and arises from the proposal of Speed and Jones (1969) that deposition of Lower to Middle Jurassic rocks was controlled locally by NW-trending flexures (D_3). NE-trending folds that predate the NW structures are recognized in the same area but are interpreted as syndepositional with Triassic rocks by Speed (1978b). Although the possibility cannot be ruled out, the arguments for syndepositional development of both fold sets are not compelling, particularly when viewed in light of regional relations.

The deformational scenario that emerges for the region east of the Pine Nut assemblage starts with minor NW-trending folds developed locally (D_1) followed by major NW–SE stratal shortening resulting in NE-trending folds, thrust imbrication, and large transport on regionally extensive thrusts (D_2). Earliest deformation (D_1) began in the Late Triassic or Middle Jurassic and was followed by the main phase (D_2), active from the Middle or Late Jurassic through the Early Cretaceous. The locus of deformation during major NW–SE shortening is poorly understood, but it is unlikely that the entire region was involved in folding and thrusting during the whole time interval. Rather, deformation probably occurred in a migrating belt or possibly belts with coarse-clastic deposition continuing in adjacent areas. No systematic timing relations exist among dismembered assemblages to suggest any younging trends in structural involvement and may reflect inadequate data or the lack of a spatially systematic structural development.

The spatial correspondence of the fold and thrust belt and the eastern and southern margins of the marine basin and the lack of comparable deformation beyond the bounds of the basin suggests that deformation, like earlier deposition, was strongly influenced by the underlying simatic basement. The basement structure did not, however, exert the same degree of control on later folds. D_3 folds have correlative structures in regions far exceeding the limits of the marine province (Oldow et al., 1982; Oldow and Avé Lallemant, 1982).

The structural evolution of the Pine Nut assemblage differs significantly from that of the region to the east and appears to be more closely allied with events in the Sierran arc. The lack of NE-trending folds (D_2) and the lateral continuity of Mesozoic stratigraphy among mountain ranges argue against involvement in the fold-thrust belt to the east. The observed NNW-trending folds are of the same orientation and have been correlated with Jurassic Nevadan structures in the Sierra Nevada (Noble, 1962; R.A. Schweickert, oral commun., 1982). With the lack of age control, however, correspondence with Cretaceous NW-trending folds (D_3) further east cannot be discounted although the weight of data on hand makes this unlikely. It is possible, however, that some later NW-trending folds (D_3) are present and are as yet unrecognized.

To adequately understand the regional significance of the structural history of the marginal basin, comparison must be drawn with events further west in the Sierra Nevada. The Mesozoic structural history of the Sierra Nevada is complex and much disagreement exists among various workers. However, it is generally accepted that the last pre-Cretaceous structures developed in post-Late Triassic rocks were formed during the Late Jurassic (Nevadan orogeny) and characteristically consist of NNW-trending folds and penetrative structures. Schweickert et al. (1983) propose the existence of a two stage development of Nevadan structures with an early main-phase forming NNW-trending folds and foliations, and a locally developed late-phase with NE-trending folds; an interpretation supported by unpublished data of H.G. Avé Lallemant (pers. commun., 1982). The age of both sets of Nevadan structures is tightly constrained as post-158 and pre-150 m.y. B.P. (Schweickert et al., 1983). Post-Nevadan folds are recognized in Cretaceous rocks in the eastern Sierra Nevada (Tobisch and Fiske, 1976, 1982; Nokleberg and Kistler, 1980) and by the reorientation of Nevadan and earlier structures in the northern and western Sierra Nevada (Schmidt, 1982; H.G. Avé Lallemant, unpubl. data). The Cretaceous age folds have NW- to WNW-striking axial planes and range from minor folds to regional flexures. No data exist to suggest that during the interval of the Late Jurassic and Early Cretaceous, between development of Nevadan and the Cretaceous structures, that the Sierra Nevada was undergoing major internal shortening by folding.

Pre-Nevadan structures of mid-Jurassic age are reported in the southern Sierra

Nevada (Nokleberg and Kistler, 1980) and may exist in the north (Warren Sharp, personal commun., 1982). In the western and northern Sierra Nevada and Sierra foothills, increasing evidence indicates the existence of a major post-Middle Triassic deformation characterized by NNW-striking folds (Avé Lallement et al., 1977; Avé Lallement, 1979; Saleeby, 1981). Folds of this set pre-date deposition of Late Triassic (Norian) carbonates, and in many areas exposing Paleozoic rocks, folds of this set may be misidentified as Nevadan.

Pronounced differences exist in the histories of the Sierran arc and the marginal basin. Deposition continued in the marginal basin during early Mesozoic tectonism in the Sierra Nevada, but regional patterns were influenced by the tectonic events to the west. Depositional relations in the marine province are quite clear and are reflected in the present distribution of lithotectonic assemblages even though significant foreshortening has occurred along a NW–SE axis. Following the Permo-Triassic Sonoma orogeny, Mesozoic deposition around the eastern and southern margins of the marginal basin was dominated by siliceous to intermediate volcanism that waned with time, as deposition of clastic rocks from an easterly source and platform-carbonate deposition became dominant. Apparently, in the basinal environments to the west, no late Early or Middle Triassic volcanism preceded deposition, because volcanic rocks of the proper age are lacking in the allochthonous assemblages. In the Black Rock assemblage, Mesozoic basin sediments are inferred to have been deposited directly on Paleozoic basinal rocks (Russell, 1981, 1984) with no evidence of intervening volcanism (Fig. 3). Farther west, in the Shasta region of northern California, basinal Mesozoic rocks (Arlington Fm.) reportedly overlie Paleozoic volcanics conformally (D'Allura et al., 1977). Basinal sites shoaled with time and accumulated shallow-marine deposits derived either from western volcanic or eastern continental source-regions. Along the southern margin, a general westerly progradation of terrigeneous clastic and carbonate rocks over volcano-carbonate environments occurred, and the westerly prograding depositional environments extended into the Sierra Nevada. Both the shoaling and western progradation continued until the Early to Middle Jurassic at which time depositional patterns abruptly changed. During the Jurassic, mature quartz sand from an easterly cratonal source (Speed and Jones, 1969; Stanley et al., 1971) was introduced into the marginal basin and is found in all but those regions (Sand Springs and Black Rock assemblages) interpreted to lie on the western edge of the basin. Quartz arenite deposition followed, and in several areas, was synchronous with local uplift causing the development of local coarse-clastic sources. Later, during continued orogenic deposition, the site of deposition of volcanogenic sedimentary and volcanic rocks migrated from west to east. Where the few Cretaceous rocks are preserved, they consist of coarse clastic and volcanic rocks suggesting deposition in a tectonically active region.

IMPLICATIONS AND TECTONIC MODEL

During pre-Nevadan deformation in the Sierran arc, continued deposition in the marginal basin resulted in shoaling of basinal environments and westerly migration of carbonates and continentally derived clastic rocks. The role of volcanism diminished and the marine province became more "stable" through time. The stabilizing of environments in the back-arc region occurred while repeated deformation was in progress in the Sierran arc. Apparently, the Sierran arc shielded the back-arc environment from the interaction of the plates of the Pacific basin and North America. Synorogenic deposition in the back-arc region, however, preceded and heralded the advent of the Nevadan deformation in the Sierran arc. NNW-trending folds in the Pine Nut assemblage are probably related to the Nevadan deformation, but east of the Pine Nut assemblage the major manifestation of Middle or Late Jurassic tectonism is found in vertical instability during deposition. Syndepositional NW-trending folds in the Black Rock assemblage may be related to the Nevadan deformation, but they are not associated with significant stratal shortening. Elsewhere, in the Gold Range assemblage, limited data (Oldow, 1981a, 1981b, unpub. data) suggest the development of a northerly trending basin active since the Early Jurassic (?) that may have been related to Nevadan or earlier tectonism. Late-phase Nevadan structures recognized in the Sierra Nevada correspond in orientation and in approximate timing of inception with the regional NE-trending folds (D_2) in the marginal-basin rocks, and they may be related. Development of both early- and late-phase Nevadan structures in the Sierra Nevada occurred in a short time ending by 150 m.y. B.P. (Schweickert et al., 1983), whereas in the marine province, deformation was protracted and continued through the Early Cretaceous.

The local development of NE-trending structures in the Sierran arc during the Late Jurassic can account for only limited shortening, but in the back-arc region folding and thrusting resulted in substantial shortening, on the order of several hundred kilometers (Oldow, unpubl. data). The axis of shortening in the back-arc fold-thrust belt (NW–SE) is subparallel to the regional trend of the Sierran arc and requires decoupling of the arc and the marginal basin. If the shortening axis of the Luning-Fencemaker thrust system was more easterly and subperpendicular to the trend of the Sierran arc, detachment at depth and A-type subduction could be called on to accommodate differences in shortening. The lack of this configuration, however, necessitates the development of either a regional strike-slip fault or a tear fault system between the thrusted assemblages and the arc. The eastern boundary of the Pine Nut assemblage is the best candidate for the fault and is referred to here as the Pine Nut fault. Regional development of the structures is probably best explained as a result of plate convergence along the active-margin of the western United States.

Existence of a right-lateral tear-fault (Fig. 6) between the Sierran arc and the fold and thrust belt is compatible with the commonly held belief in long standing

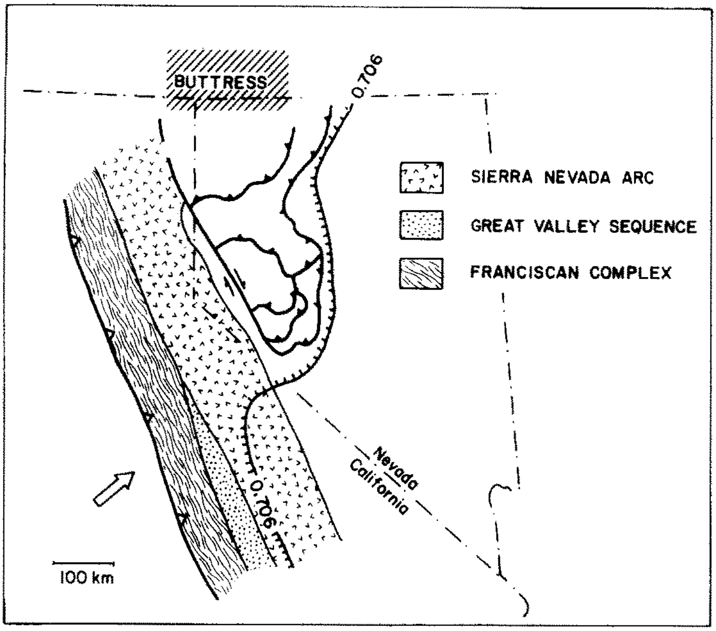


Fig. 6. Tear-fault model for decoupling of dismembered assemblages from the Sierran arc.

right-oblique subduction, but requires the mechanically unlikely configuration of a thin flap of allochthonous rocks migrating southerly with respect to a coupled back-arc autochthon and arc. This model is not compatible with the regional development of NE-trending folds nor the NW to SE transport of thrusts (D_2). It also is not easily reconciled with our present understanding of convergent-boundary configurations.

On the other hand, existence of a regional strike-slip fault (Fig. 1) with left-lateral displacement is much more reasonable and is compatible with the regional axis of shortening of the fold and thrust belt. It requires, however, north and south extensions of the fault system beyond the marginal basin. This carries the implication that the Sierran arc was, for some along-strike segment, decoupled from North America and acted as a relatively rigid block. The position of the northern continuation of the fault is unclear due to extensive Cenozoic deformation but probably lay behind the Mesozoic successions exposed in the Shasta region of northern California and extended northerly and east of the Seven Devils and Huntington arc terranes of eastern Oregon. Though now curvalinear, the trace was presumably much straighter and owes its current shape to Cenozoic extension and rotation. The southern projection of the fault may correspond with an ancestral White Mountains–Owens Valley fault system or Laurel-Convict fault system lying east of the main body of the Sierra Nevada batholith. Left-lateral slip has been proposed for such a system in the

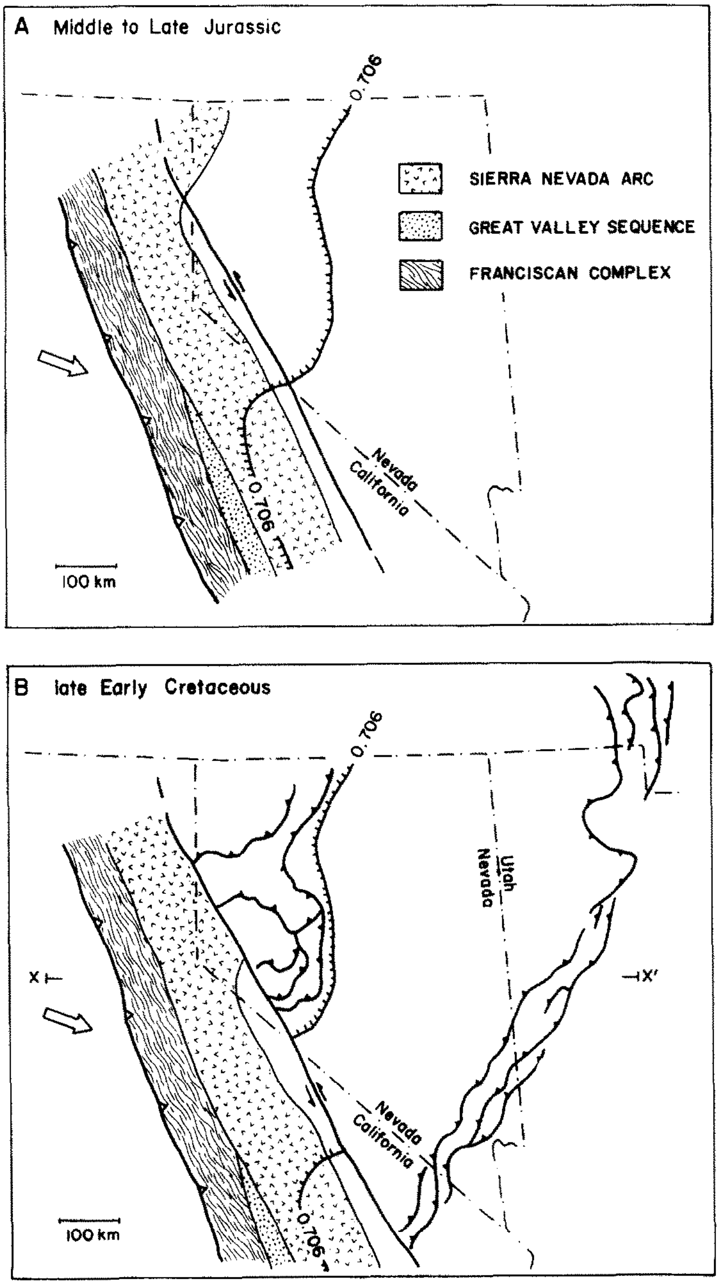
Late Jurassic and Early Cretaceous (Dunne et al., 1978; Schweickert, 1981). The timing of inception, sense of motion, and southern projection of the Pine Nut fault is also compatible with the major Jurassic mega-shear proposed by Silver and Anderson (1974). The duration of the Sonora–Mojave mega-shear, however, is much shorter than that of the Pine Nut fault system and may have lasted only 10–15 m.y. (T. Anderson, pers. commun., 1982). If the two systems acted in concert for part of their histories, the suggestion is that the southern locus of displacement may have migrated through time.

The magnitude of the left-lateral displacement of the Pine Nut fault approximately equals or exceeds the amount of stratal shortening in the back-arc fold and thrust belt. The dismembered assemblages overlying the Luning-Fencemaker thrust are composed of numerous thrusts which, where detailed work is complete, have individual displacements of kilometers to tens of kilometers (Oldow, 1978, 1981a, 1981b). Displacements between lithotectonic assemblages are larger than displacements among thrust sheets within assemblages, which is apparent from the relative facies continuity within individual assemblages as opposed to the abrupt juxtaposition of age equivalent facies across lithotectonic boundaries. Aggregate contraction in several lithotectonic assemblages indicates that several hundred kilometers of shortening is accommodated by the back-arc fold and thrust belt as a whole.

The structural relations are set in a plate tectonic framework for the late Mesozoic and are illustrated in Fig. 7. Understanding of the origin of pre-Nevadan deformation(s) and the development of Nevadan structures in the Sierra Nevada is evolving rapidly, and specific models for the tectonic configuration are beyond the scope of this paper. Of importance here are the implications of the proposed fault system between the Sierran arc and the back-arc region and the localized development of the back-arc fold and thrust belt.

Estimation of the sense of relative plate-motion along active plate-margins by using the orientations of regional structures is common but controversial. Early interpretations of structures in the Sierra Nevada (Davis et al., 1978; Burchfiel and Davis, 1981; Saleeby, 1981) led investigators to propose right-oblique convergence since the Early Triassic, possibly with intermittent periods of collision and accretion of exotic fragments from the Pacific Ocean basin. Whether structures in the upper plate are developed during periods of collision or as a consequence of continuous convergence is not well understood. The significance of the orientation of structures, particularly folds, in the magmatic arc and fore-arc regions is also questionable. A growing body of evidence (Walcott, 1978; Moore and Karig, 1980) suggests that the axis of shortening in the leading edge of the upper plate is not directly controlled by the vector of relative plate-convergence, but rather by the local orientation of the plate boundary. In oblique subduction, the shortening axis probably reflects the normal component of convergence, whereas the strike-slip component apparently is accommodated by strike-slip faults in or behind the arc region. Thus, the structures within the Sierra Nevada may not yield information concerning the sense of obliquity of plate convergence but rather the orientation of the plate boundary.

On the other hand, development of the left-lateral strike-slip fault and associated fold and thrust belt in the back-arc region east of the Sierra Nevada may carry



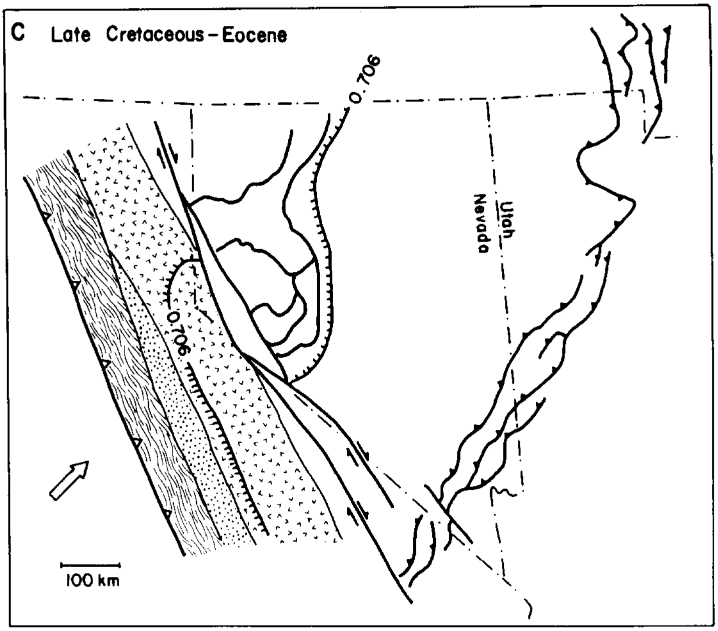


Fig. 7. Preferred tectonic model for the development of back-arc thrusting in the late Mesozoic.
A. Initial condition in the Middle to Late Jurassic immediately after inferred change in plate convergence direction from right-oblique to left-oblique and before left-lateral displacement on the Pine Nut fault system. Initial strontium 0.706 contour is restored to presumed original geometry (after Oldow, 1984a). No compensation for Cenozoic extension is made.
B. Final stage of development of the Luning-Fencemaker thrust belt and the Sevier thrust belt in the late Early Cretaceous. The Sierran arc has translated south as a relatively rigid block along the western margin of North America and offsets the initial strontium contour.
C. Right-lateral migration of the Sierran-arc after reinstatement of right-oblique convergence in the latest Early Cretaceous (pre-100 m.y. B.P.). The Luning-Fencemaker thrust belt is no longer active, whereas the Sevier belt undergoes continued shortening. The Sierran arc has undergone northward translation on faults succeeding the Pine Nut system.

important implications concerning relative plate motions. In all likelihood, the structures are largely the product of the strike-slip component in a system of rapid plate convergence. The inferred left-lateral slip on the Pine Nut fault and the NW–SE shortening axis for the Luning-Fencemaker thrust belt developed over a period of 40–50 m.y. and is incompatible with the commonly held belief of right-oblique subduction for the Sierran arc. Rather, they suggest a component of left-oblique convergence for much of the late Mesozoic. It must be stressed, however, that the axis of shortening is not necessarily parallel to the direction of relative plate-motion and is useful only as a general, quadrant estimate of the motion. The inferred structural history of part of the western Cordillera is given for three time frames in Fig. 7.

South of the simatic enclave, in the back-arc region underlain by continental crust, thrusts of late Mesozoic age are developed but have moderate displacements (Dunne et al., 1978) and do not constitute a regionally significant belt of contraction. This suggests that decoupling between the arc and back-arc was more complete and occurred in a much narrower zone in the region underlain by continental crust than in the region underlain by simatic crust. This generalization appears valid until the southwestern end of the Sevier fold and thrust belt is encountered in southern Nevada (Fig. 1).

The trend of the Sevier fold and thrust belt through the western Cordillera corresponds with the depositional hingeline inferred for the Paleozoic miogeocline. The congruity of trends suggests that a change in crustal thickness corresponding to the site of flexure may have been a critical ingredient in locating the thrust belt. The age of inception of the Sevier belt (Allmendinger and Jordan, 1981) may be contemporaneous with the Luning-Fencemaker system, but that notwithstanding, they were active concurrently for parts of their histories. Thus, until the mid-Cretaceous, the "Andean-type" arc of North America had paired, back-arc fold and thrust belts separated by a zone of tectonic quiessence that underwent relatively minor internal shortening. In all likelihood, the "quiet zone" was detached at depth, as proposed by Speed (1983), suggesting that the Sevier and Luning-Fencemaker systems may have shared a common décollement (Fig. 8).

During translation of the Sierran arc south, motion on the Pine Nut fault necessarily was transpressional allowing rocks of the marine-province to be partially coupled with the arc, to become detached from their substratum, and to undergo imbrication. An attractive model involving underthrusting of rocks overlying thick sialic basement beneath those deposited on the simatic basement cannot be substantiated. Regional timing relations for structures within the allochthonous terranes indicate that the thrusting and folding were initiated near the center of the

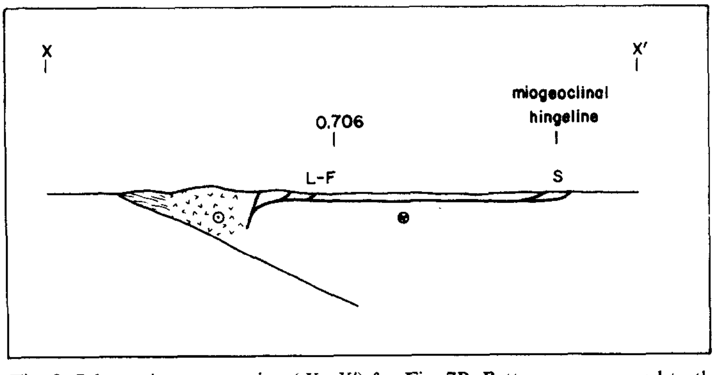


Fig. 8. Schematic cross section ($X-X'$) for Fig. 7B. Patterns correspond to those in Fig. 7, bold lines are faults: *L-F* denotes the Luning-Fencemaker thrust belt; *S* denotes the Sevier thrust belt. Circle-dot and circle-cross indicate motion out of and into plane of page, respectively.

marine-province and not at the southern margin. Strict compliance with an overthrust scenario also fails because the youngest rocks deformed are located both in the uppermost plate (Black Rock assemblage) and the lower plate (Gold Range assemblage). Various arguments calling upon reactivation of older structures can be proposed to support either model, but sufficient data are not now available to discriminate between them.

Following the period of left-lateral convergence on the plate margin (Fig. 7), right-lateral convergence was apparently reinstated as suggested by the regional development of NW-trending folds. The NW-trending folds are not spatially restricted to areas involved in the back-arc thrusting and are found throughout the region, including the Sierra Nevada. The NW-trending structures post-date NE-trending folds in Lower Cretaceous rocks in the Gold Range assemblage and pre-date the major plutonic pulse of the Cretaceous (100–90 m.y. B.P.) suggesting a very rapid change in compressional direction. Apparently, the sense of displacement on the retro-arc strike-slip fault system reversed and became right-lateral.

An important implication is that strike-slip displacements of opposite sense can be superimposed on the fault system. Thus, the observed displacement of the fault bounding the eastern side of the Pine Nut assemblage may be the product of a complex history of motion. It seems unlikely, however, that a significant proportion of the accumulated left-slip was lost during later right-lateral displacement. Post-Oligocene right-lateral displacement in the Walker Lane, a descendent of the Pine Nut fault system, is estimated at about 48 km (Hardyman et al., 1975), and no evidence exists to support earlier displacements of significant magnitude. A more likely candidate for the locus of right-lateral displacement lies to the west of the Pine Nut assemblage and parallels the eastern margin of the Sierra Nevada batholith. Our current understanding of regional stratigraphic relations between the Sierra Nevada and the western Great Basin does not allow the necessary control for deducing the ages of inception of individual strands of the fault system. It is not possible to ascertain whether or not the proposed fault west of the Pine Nut assemblage existed during left-lateral displacement.

CONCLUSIONS

Structural relations in the western marine province of the Great Basin and their association with events in the Sierra Nevada bear several important implications:

(1) A regionally extensive back-arc fold and thrust belt developed in a marginal basin associated with the Mesozoic Sierran arc and resulted in major NW–SE shortening and imbrication of lithotectonic assemblages during the Middle or Late Jurassic and Early Cretaceous.

(2) Simatic crust underlying the northwestern Great Basin profoundly affected Mesozoic deposition and localized development of the Luning-Fencemaker fold and thrust belt.

(3) For much of their histories during the late Mesozoic, the Sierran arc and the back-arc region were at least partially decoupled and separated by a regionally extensive strike-slip fault with left-lateral displacement.

(4) The partial synchroneity in development of the Luning-Fencemaker and Sevier fold and thrust belts suggests that they are related and may have shared a décollement during part of the late Mesozoic.

(5) The shortening axis of the Luning-Fencemaker fold and thrust belt is apparently incompatible with right-oblique plate convergence for the western margin of North America and suggests an interval of left-oblique subduction from the Middle or Late Jurassic through the Early Cretaceous.

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