Paleogeographic and tectonic implications of the Golconda allochthon, southern Toiyabe Range, Nevada

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ABSTRACT

The Golconda allochthon in the southern Toiyabe Range, Nevada, is composed of several thrust slices of rocks that belong to three age categories: early Paleozoic, Early Triassic, and late Paleozoic. These slices were stacked mainly during the Sonoma orogeny that occurred between Permian and late Early Triassic time. The late Paleozoic rocks, composing the bulk of the allochthon, range in age between Mississippian and Late Permian and contain four lithotypes that occur in the following structural order above the Golconda thrust: terrigenous, hemipelagic, volcanogenic, and pelagic.

The upper Paleozoic rocks were deformed several times during their eastward emplacement onto the North American continental margin. The deformation involved isoclinal folding and thrusting of these rocks into slices. The slices were openly and gently folded during emplacement, and sporadic mesoscopic folds were formed along their boundaries. Post-thrust movements along slice boundaries formed kink bands and related folds in slate and chert interbeds. The kinks were later deformed when the Golconda thrust was reactivated. Further contraction, during the Sonoma orogeny or possibly later, was accommodated by imbrication of slices of Early Triassic and their unconformably underlying lower Paleozoic rocks.

INTRODUCTION

Middle and late Paleozoic tectonic history of the western edge of the United States in Nevada is dominated by the emplacement of (1) the Roberts Mountain allochthon during the late Early Mississippian Antler orogeny (Roberts, 1951; Roberts and others, 1958; Stewart and Poole, 1974; Smith and Ketner, 1977; Nilsen and Stewart, 1980; Johnson and Pendergast, 1981; Speed and Sleep, 1982) and (2) the Golconda allochthon during the Early Triassic Sonoma orogeny above coeval shelf deposits and their underlying Roberts Mountains allochthon and autochthon (Silberling and Roberts, 1962; Speed, 1971, 1977b, 1979; Silberling, 1973). Both the Antler and Sonoma events have been interpreted as due to overthrusting of the continental margin, but the origin of the allochthons is controversial. There are two models: forearc and backarc. The forearc model assumes that the rocks of both alochthons were deposited on oceanic lithosphere that fronted the passive margin of western United States and were accreted to a forearc or accretionary prism (Seely and others, 1974) of a migrating island arc above a west-dipping subduction zone (Moores, 1970; Speed, 1971, 1978, 1977b; Dickinson, 1977; Schweickert and Snyder, 1981; Speed and Sleep, 1982; Dickinson and others, 1983; Snyder and Brueckner, 1983; Brueckner and Snyder, 1985). The backarc model maintains that the rocks of the two allochthons were deposited in local backarc basins of subduction zones that dipped east below the arc, toward Nevada, and were thrust onto the North American continental margin when the basins were closed (Burchfiel and Davis, 1972, 1975; Silberling, 1973; Whiteford and others, 1983; Churkin, 1974; Miller and others, 1984).

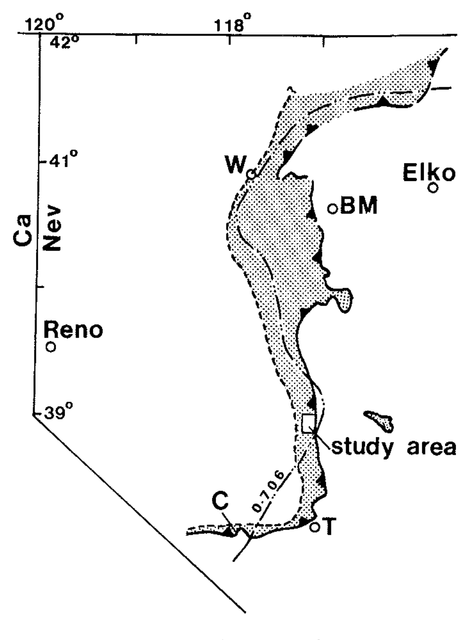


Figure 1. Regional extent of the Golconda allochthon in Nevada showing the trace of the $^{87}Sr/^{86}Sr$ = 0.706 line and location of the study area (after R. C. Speed, unpub. map). W = Winnemucca, BM = Battle Mountain, C = Candelaria, T = Tonapah. The solid line with teeth is the Golconda thrust.

During the Mississippian, after the Antler event, enormous volumes of clastic debris were shed eastward from the Roberts Mountains allochthon into a foreland basin where thick clastic wedges were deposited (Poole, 1974; Poole and Sandberg, 1977; Harbaugh and Dickinson, 1981; Speed and Sleep, 1982). Clastic debris was also transported west of the Antler orogenic belt to be deposited in the upper Paleozoic overlap sequence (Roberts and others, 1958; Silberling and Roberts, 1962; Speed and others, 1977; Miller and others, 1981; Saller and Dickinson, 1982) and in the oceanic sediments of the Golconda allochthon, the subject of this paper. Major problems related to the Golconda allochthon are (1) the location, age, extent, and tectonic history of the basin(s) in which the rocks of the allochthon were deposited and (2) the general architecture, mechanism, and relative age of the tectonic processes responsible for folding and thrusting of the internal slices, and direction and magnitude of the tectonic transport.

The Golconda thrust was first defined in its type locality in northern Nevada near Winnemucca (Muller and others, 1951) and was later extended 250 km south by Speed (1971) to Candelaria (Fig. 1). The emplacement of the allochthon has been assigned an early to late Early Triassic age (Speed, 1971; Nichols, 1971; MacMillan, 1972; Silberling, 1973), although Jurassic and Cretaceous ages have also been proposed (Ketner, 1984).

The tectonostratigraphic units of the allochthon in different parts of Nevada include the Havallah and Pumpernickle Formations (Haval-

Geological Society of America Bulletin, v. 99, p. 231–243, 13 figs., 2 tables, August 1987.

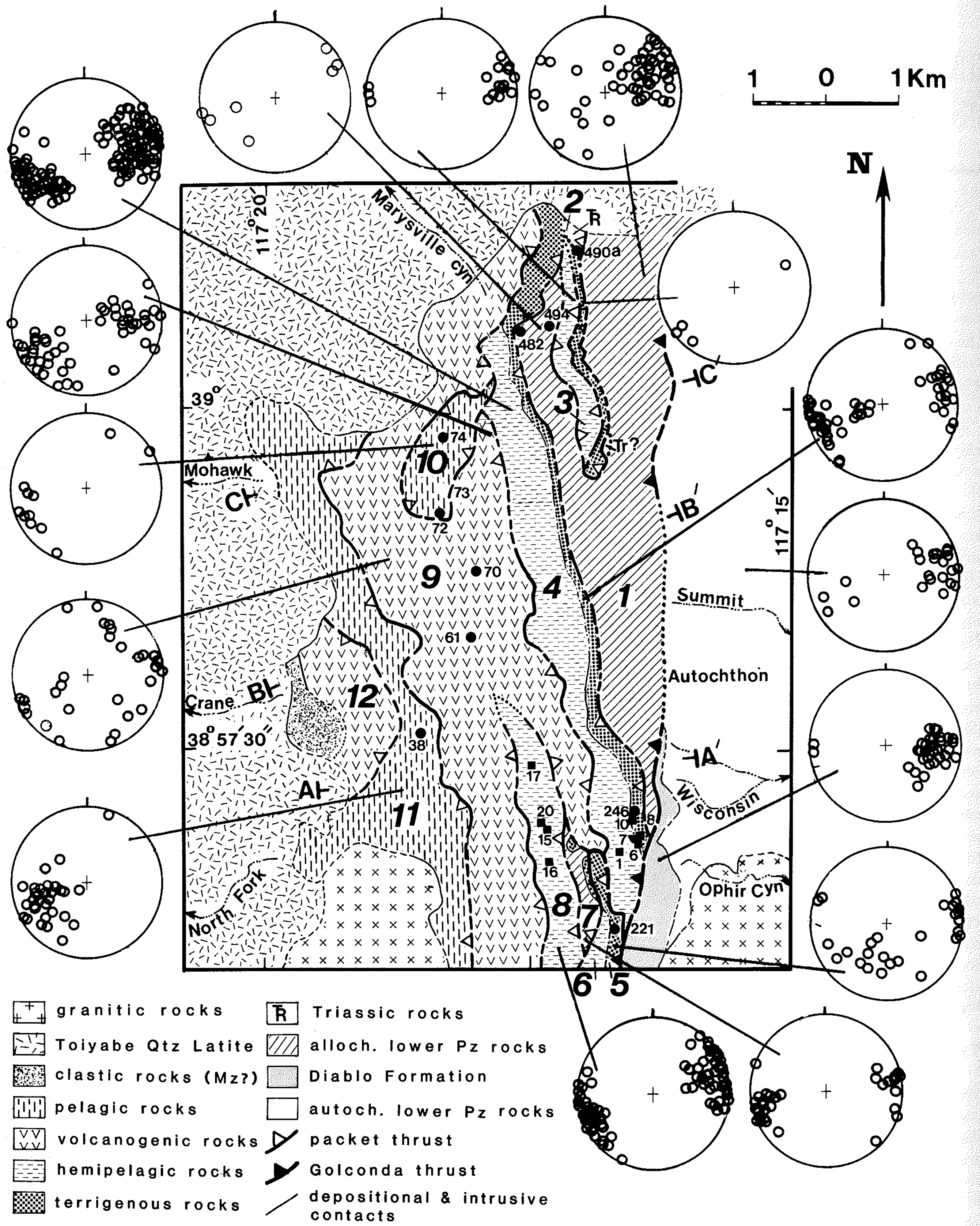


Figure 2. Geologic map showing the Golconda allochthon, autochthon, cover strata, and intrusions in the study area. Stereograms show the orientation of bedding and its coplanar cleavage in different domains. Filled circles and squares with numbers (in the map) show location and number of the specimens dated or used for strain measurement, respectively. A–A′, B–B′, and C–C′ are lines of cross section (see Fig. 3). Large numbers identify slices. See Table 1 for lithology and age of the slices.

lah sequence) (Muller and others, 1951; Silberling and Roberts, 1962; Silberling, 1973; Stewart and others, 1977; Snyder and Brueckner, 1983; Brueckner and Snyder, 1985), the Pablo Formation in the study area (Ferguson and Cathcart, 1954; Speed, 1977a), the Schoonover sequence (Fagan, 1962; Churkin and Kay, 1967; Miller and others, 1981, 1982, 1984), and the Willow Canyon Formation (Laule and others, 1981). The objective of this paper is to describe the tectonostratigraphy and structure of the allochthonous, and to a lesser extent, autochthonous units relative to the Golconda thrust in the southern Toiyabe Range.

CONTINENTAL MARGIN IN THE SOUTHERN TOIYABE RANGE

Major features of the Golconda allochthon in the southern Toiyabe Range (Fig. 2) are (1) thrust slices of Permian and Triassic rocks below a succession of slices of Mississippian-Permian rocks, (2) sedimentary and tectonic fragments of serpentinite along slice contacts, (3) slices of Ordovician rocks at lowest structural levels, and (4) penetrative cleavage in both upper and lower Paleozoic rocks.

The Toiyabe Range lies along a significant stratigraphically recognized, north-trending, shelf-edge, outer ridge that existed before the emplacement of the Roberts Mountains allochthon (Kay and Crawford, 1964; Matti and McKee, 1977). During the late Paleozoic, this portion of the Roberts Mountains allochthon occupied the seaward slope of the continental orogenic belt, and it formed the basement for the deposition of the Diablo Formation (Speed and others, 1977). The near coincidence of the two early and late Paleozoic, north-trending shelf edges in the southern Toiyabe Range, indicates either: (1) the continental margin was not extended westward after the emplacement of the Roberts Mountains allochthon, or (2) the part of the allochthon that lay west of the pre-Roberts shelf edge, if it existed, was either eroded, tectonically or otherwise, or subsided before the late Paleozoic.

The southern Toiyabe Range lies near the buried edge of sialic Precambrian North America which is suggested by separation between high and low isotopic strontium ratio ($^{87}Sr/^{86}Sr = 0.706$) for Mesozoic and Cenozoic magmatic rocks (Fig. 1) (Kistler and Peterman, 1978). In this part of Nevada, the contour follows a belt across which an eastward thickening of the crust is suggested by seismic and gravity data (Prohdehl, 1978; Cogbill, 1979). The edge may represent a preserved segment of the western passive margin (Speed, 1983) of the sialic North American continent that formed by rifting during the late Precambrian or possibly later (Stewart, 1972).

TECTONOSTRATIGRAPHY AND DEPOSITIONAL ENVIRONMENT

The southern Toiyabe Range includes extensive exposures of pre-Tertiary rocks that are interpreted to contain the following tectonostratigraphic units: (1) autochthonous North America, (2) Roberts Mountains allochthon, (3) cover strata (Diablo Formation) to the Roberts Mountains allochthon, (4) Golconda allochthon, and (5) Tertiary intrusive and cover volcanic strata. The Golconda thrust, best exposed in Ophir and Wisconsin Canyons (Fig. 2), defines a sharp facies, age, and deformation boundary between the multiply deformed rocks of the Golconda allochthon and underlying rocks of the Diablo Formation and Roberts Mountains allochthon (Roberts and others, 1958; Kleinhampl and Ziony, 1967; Speed, 1977a; Speed and others, 1977; Babaie, 1984). The thrust dips moderately to the west as indicated from its three-dimensional exposure in Wisconsin Canyon. The Golconda allochthon is composed of several thrust slices that can be grouped into three age categories in the following ascending structural order: (1) early Paleozoic, (2) Early Triassic, and (3) late Paleozoic. The faults that separate the slices are identified by abrupt lithologic, structural, and, in places, age discordance; by the occurrence of fault slices of serpentinite in low structural levels; and by the presence of tectonized rocks. Although three-dimensional views of the slice-bounding thrusts are not seen, they are interpreted to be openly folded based on their internal fabrics (Fig. 3). The exposed width of the slices varies between 0.1 and 2.2 km (Fig. 2), and the general trend is north-northwest. The following paragraphs concentrate on the tectonostratigraphy of the Golconda allochthon.

Allochthonous Rocks

Upper Paleozoic rocks, belonging to the Pablo Formation (Speed, 1977a), make up the bulk of the Golconda allochthon. These include four lithotypes that occur in the following ascending structural order above the Golconda thrust: terrigenous, hemipelagic, volcaniclastic-basaltic, and pelagic (Table 1). Radiolaria in the chert beds of the pelagic and volcanogenic rocks yield depositional ages between very latest Mississippian and Late Permian (Guadalupian) (Table 1). Clast ages within terrigenous rocks include Ordovician and Early to Middle Pennsylvanian. These clastic rocks are extremely poorly sorted and in places contain large (as much as 50 m in diam.) blocks of gray, medium- to coarse-grained quartzite. These blocks are texturally similar to autochthonous lower Paleozoic quartzites in Summit and Marysville Canyons,

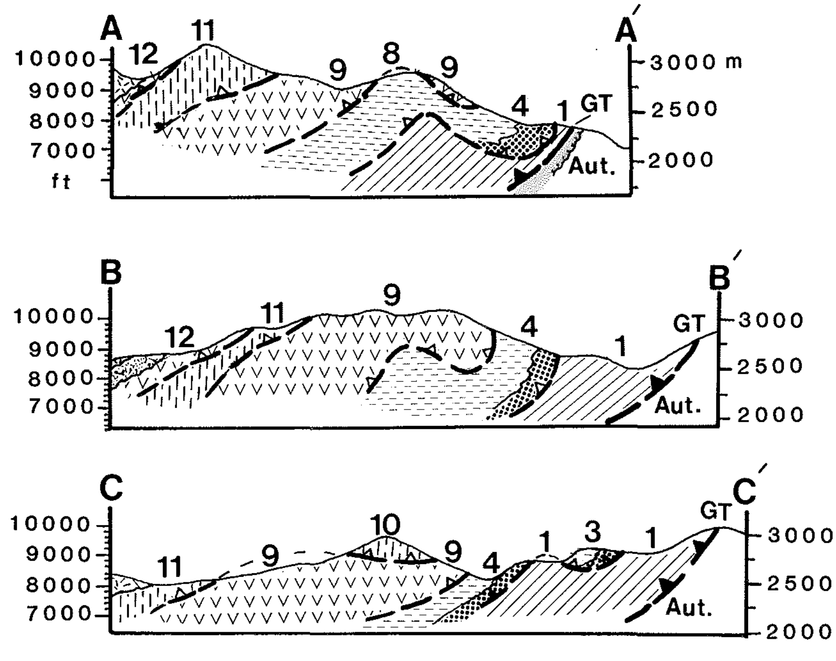


Figure 3. Cross sections of the Golconda allochthon. Numbers identify slices. See Figure 2 and Table 1 for lithology and age of the rock units in different slices. GT is the Golconda thrust. "Aut." represents the autochthon with respect to the Golconda thrust. The thick dashed lines with open teeth depict slice-bounding thrusts.

TABLE 1. ROCK TYPE AND AGE OF THE ALLOCHTHONOUS AND AUTOCHTHONOUS UNITS

Slices and lithotype	Rock type	Fossils/age
Allochthonous upper Paleozoic rocks		
10, 11 pelagic	Radiolarian cherts and thin-bedded pelite with lenses of volcanogenic sandstone.	Sp. 38: rads/ v. Miss. or v. earliest Penn. Sp. 72: rads/ probably Miss. Sp. 73: rads/ v. latest Miss. or v. earliest Penn. Sp. 74: rads/ probably Miss. or Penn.
9, 12 volcanogenic and volcanic	Poorly sorted volcanogenic sandstone and sedimentary breccia, pelite, radiolarian chert, and pillowed and massive basalts, clasts: chert, pelite, carbonates, and igneous rocks. Igneous clasts: (1) porphy, volc. rocks w/ plag. and chloritized and serpentinized cpx and horn. phenos. (2) granitoids w/ qtz, K-spar, and chloritized muscovite.	Sp. 61: rads/ Guadalupian (Permian) Sp. 70: rads/ Pennsylvanian or Early Permian
3, 4, 6, 8 hemipelagic	Interbedded pelite, chert, and sporadic lenses of graded chert- and quartz-bearing sandstone.	Sp. 246: conodonts in terrigenous rocks base of 4 *Declinognsthodus Noduliferus* *Idiognathoidus* sp. aff., I. *Convexus* of Dunn (1970). Age: E. to Mid. Penn. (Narrowan to Atokan).
5 and bases of 3 and 4 terrigenous	Poorly sorted sandstone and sedimentary breccia with limestone, chert, quartzite, quartz, serpentinite, and pelite clasts.	Sp. 221: conodonts (in 1st clasts): *Oneotodus* cf. *O. Simplex, Paroistodus*? sp. *Paltodus Bassleri. Scalpellodus* ? sp. *Teridontus* sp. oistodontiform element indet. drepanodontiform and paltodontiform elements. Age: Early Ordovician Sp. 246: see above.
Allochthonous lower Paleozoic rocks		
1, 7	Dark chert, pelite, limestone, chert, and quartzite-bearing sandstone.	Sp. 482: conodonts (1st): *Cordylodus Proavus* Muller. indet, phosphatic problematicum. Age: latest L. Cambrian or ear. E. Ordovician Sp. 494: conodonts: oistodontiform element. gen. and sp. indet. Age: probably Ordovician.
Allochthonous ? Triassic rocks		
2	Limy quartz-, chert-, volcanic rock-bearing sedimentary breccia, sandstone, and conglomerate, and limy mudstone, chert, and quartzite.	Sp. 490a: conodonts: *Neogondolella* sp. elements of prob. Triassic morphotype (all incomplete). Age: probably Triassic.
Autochthonous Diablo Formation	Interbedded quartz- and chert-bearing, graded sandstone and conglomerate, and clayey and silty sandstone, pebbly calcarenite, and marble	Age: unknown in this area. Same rocks, 20 km south, have yielded fairly L. Permian age (Wolfcampian-Guadalupian) (Speed and others, 1977)

Note: for slice numbers and specimen locations, see Figure 2.

suggesting local derivation. The Ordovician age of the limestone clasts supports this correlation. The occurrence of limestone and pelitic clasts with blocks of harder bedded chert and quartzite, suggest deposition by mass flows near a steep slope. Such a slope must have been developed west of the Roberts Mountains allochthon and its Pennsylvanian shelf cover strata.

The hemipelagic rocks are defined by thin-bedded chert and pelite (Table 1). Hemipelagic sedimentation was dominated by the deposition of suspended or eolian mud and biogenic ooze (Speed, 1977a). The contact with terrigenous clastic rocks is gradational and defined by an increase in the occurrence of lenses and beds of graded sandstone and fine-grained breccia. The hemipelagic rocks were probably deposited in those parts of the basin too remote for the reach of the continent-derived detritus except when sporadic distal turbidity currents carried in terrigenous sand- or granule-sized particles.

The pelagic sediments, dominated by radiolarian cherts (thicker than the hemipelagic cherts) and thin pelite (Table 1), were probably deposited in more distal areas of the oceanic basin. These areas received little continent-derived debris but copious volcanogenic detritus. The chert and pelite are interbedded and occur as clasts in the volcanogenic breccia and sandstone. The large size of the chert and pelite clasts suggests that they were transported a short distance and that they were intraclasts derived from the walls of local channels. Moreover, some granitic clasts (Table 1) indent the adjacent chert clasts, indicating that the cherts were relatively soft during deposition of the breccia. The volcanogenic debris was probably transported to the oceanic floor by channelized gravity flows that were, at least partly, dominated by turbidity currents. These flows formed the observed massive and graded sandstone beds with load structures and finer-grained sandstones with convolute and cross stratification. The presence of channels is suggested by the localized massive, thick, and poorly sorted volcanogenic breccia that contains abundant resedimented intraformational coeval pelagic clasts that were probably plucked from channel walls. The influx of the volcanogenic sediments was intermittent and non-uniform in terms of the load transported, made evident by the sporadic occurrence of volcanogenic beds and their variable thickness in the pelagic rocks.

Pillowed basalts apparently occur as thrust slices intercalated with slices of volcanogenic and pelagic sediments and as basement on which these sediments were deposited. About 20 km south of the study area, Speed (1977a), using major-element chemical analysis, suggested that these were oceanic basalts. Deposition of the volcanogenic sediments on an ocean crust is further suggested by the radiolarian chert and red pelitic interbeds.

Allochthonous lower Paleozoic rocks occur under slices of upper Paleozoic rocks along low-

TABLE 2. DEFORMATION EVENTS IN THE UPPER PALEOZOIC ROCKS OF THE ALLOCHTON

Event	Bedding	Deformation	Fold	Axial plane	Axis	Foliation	Fault	Comments
Pre-thrust	S0							Deposition of the upper Paleozoic terrigenous, hemipelagic, volcanogenic, and pelagic rocks.
		D1	F1	apl	apl × S0	S1		Tight and isoclinal folding (F1) in bedding with axial-plane slaty cleavage (S1). Boudinage of the chert beds.
Syn-thrust		D2	F2	ap2	ap2 × S0 ap2 × S1		Slice-bounding thrusts	Thrusting of the isoclinally folded rocks in slices and formation of minor mesoscopic folds (F2) along the slice boundaries.
		D3	F3	ap3	ap3 × S0 ap3 × S1			Large-scale open and gentle folds (F3) of bedding, cleavage, and thrusts, affecting one or more slices.
Post-thrust		D4	F4	ap4	ap4 × S0 ap4 × S1			Formation of two orthogonal sets of kink bands in cleavage and sinusoidal folds in bedding (F4). Formation of axial plane fractures.
		D5	F5	ap5	ap5 × S0 ap5 × S1		Movement along the Golconda thrust. Local faulting	Translation of the upper plate of the Golconda thrust. Deformation of the kink bands and formation of mesoscopic folds (F5) and associated faults at the thrust zone.

and high-angle faults at the base of the allochthon. Although their lithology is similar in part to those of the upper Paleozoic hemipelagic rocks, the thin-bedded limestone and organic-rich pelite allows them to be distinguished and allied with the lower Paleozoic rocks of the Roberts Mountains allochthon below the Golconda thrust. Ordovician conodonts in the limestone interbeds support this interpretation (Table 1). Moreover, as will be discussed below, the tectonic fabric of these rocks is very different from that of the upper Paleozoic slices.

One thrust slice of rocks with probable Early Triassic age and unknown extent and thickness is recognized between overlying upper Paleozoic rocks and underlying lower Paleozoic rocks (Fig. 2; Table 1). Pectens and ammonites collected by G. A. Poole, B. R. Wardlaw, and F. G. Poole from the same locality were provisionally dated as Early Triassic by N. J. Silberling (Norm J. Silberling, 1986, written commun.). The lithology of these Lower Triassic rocks (Table 1) is very similar to other undated Diablo-like allochthonous clastic rocks that occur in several slices near the Golconda thrust in the study area. It is possible that at least some of these coarse clastic rocks are also of Early Triassic age.

Autochthonous Rocks, Intrusions, and Cover Strata

The Diablo Formation (Table 1) lies immediately under the Golconda thrust in Ophir and Wisconsin canyons and is equivalent to unit 1 of Speed and others (1977), which can be correlated with their unit 3 in Jett Canyon where macrofossils yield fairly late Permian ages. The Diablo Formation in Jett and Pablo Canyons rests unconformably on lower Paleozoic rocks and has been interpreted as shallow-marine, outer-shelf sediments (Speed and others, 1977).

Autochthonous rocks are mostly lower Paleozoic. These have been mapped as Cambrian Gold Hill Formation and Cambro-Ordovician Palmetto Formation by Kleinhampl and Ziony (1967) and are probably part of the Roberts Mountains allochthon (autochthon relative to the Golconda thrust) as indicated by their lithology, age, and tectonic fabric.

Granitic rocks of the Ophir pluton intrude and locally metamorphose both autochthonous and allochthonous rocks in lower and upper Ophir Canyon (Fig. 2). The Ophir pluton has previously provided a 54 Ma K-Ar date from biotite (Speed and McKee, 1976). A 42 Ma K-Ar date from biotite has been obtained in the present study from the eastern part of the pluton in Ophir Canyon.

Toiyabe quartz latite (Ferguson and Cathcart, 1954; Speed and McKee, 1976) unconformably overlies most of the Golconda allochthon on the western part of the Toiyabe Range and an undated granitic pluton in upper Ophir Canyon. It includes partly welded crystal ash-flow and crystal-pumice ash-flow tuffs that give a 21.5-m.y. K-Ar age (Kleinhampl and Ziony, 1967; Speed and McKee, 1976). It is believed that they were deposited after a period of early Miocene caldera formation and volcanism (Speed and McKee, 1976).

STRUCTURAL GEOLOGY

Slices of Upper Paleozoic Rocks

Contacts between terrigenous clastic, hemipelagic, pelagic, volcanogenic, and basaltic rocks are dominantly faults but locally depositional (Figs. 2 and 3). The slice-bounding thrusts deform all of these upper Paleozoic rocks and are used as a reference for dividing the deformation into a pre-, syn-, and post-thrust sequence as described below (Table 2). All stereo nets are equal-area lower-hemisphere projections.

Pre-Thrust Deformation (D1)

Bedding within all upper Paleozoic slices is tight to isoclinally folded. Such folds are more common in thin-bedded chert and pelite and are apparently not well developed in the massive terrigenous and volcanogenic rocks. In chert, the axial planes of these folds (Fig. 4E) are parallel to (1) a penetrative slaty cleavage in interbedded pelite and a spaced foliation in sparse sandstones (Fig. 4B) and to (2) bedding in the limbs (Fig. 4A). These folds and cleavages are the earliest structures (F1) and are cut by slice boundaries. Other F1 structures include boudins and bedding-cleavage intersections (bxc lineations) in chert (Fig. 4C). The hingelines (directly measured) and axes (bedding intersections around fold closures) of the F1 folds lie in the plane defined by the bxc lineations and trend between north and N45°W and plunge between 24° and 75° (Fig. 4D). Where strain has been

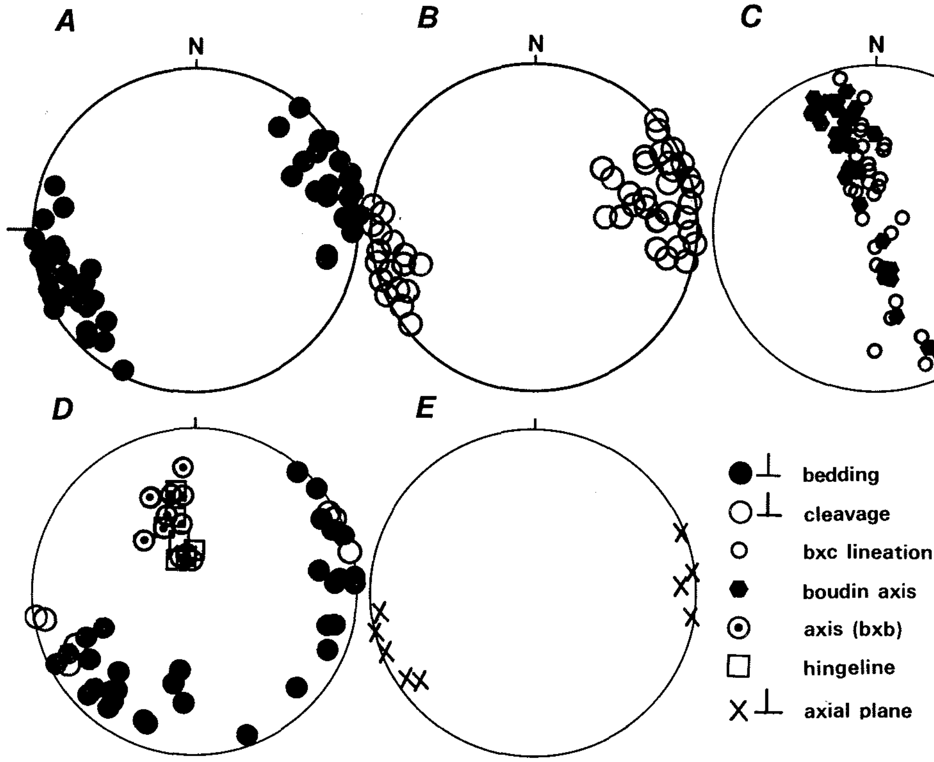
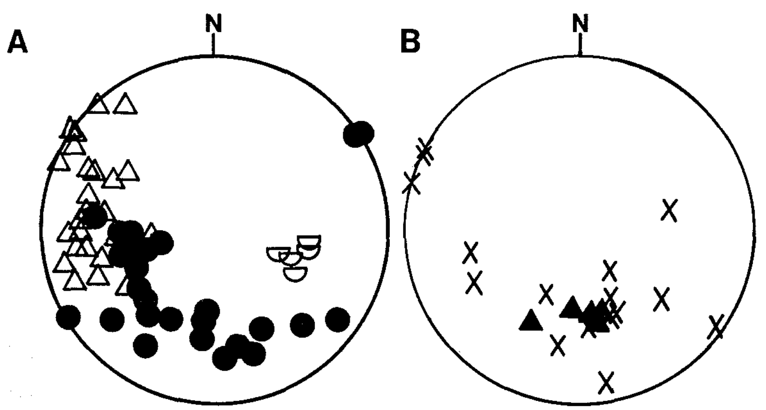


Figure 4. Orientation data of F1 structures in slice 8: (A) homoclinal bedding; (B) homoclinal cleavage; (C) boudin axes and bxc lineations; (D) fold axes, hingelines, and bedding around fold closures; and (E) axial planes.

measured (see below), most lineations are subparallel to the Y axis of the strain ellipsoid (X>Y>Z) in each domain.

The great circle distribution of the D1 lineations (Fig. 4C) could have been formed by several mechanisms. (1) Inhomogeneous coaxial strain: cleavage of different initial orientations formed, for example due to anisotropy, resulting in variably oriented intersections with a bedding of constant or variable initial orientation (possible pre-D1 structures in bedding). Progressive or later flattening caused the cleavages to become subparallel and the lineations to become coplanar. (2) Inhomogeneous noncoaxial strain: orientation of the Z principal strain was constant, but X and Y rotated, leading to the formation of differently oriented boudins on a constantly oriented cleavage. (3) Superposed folding: unidentified pre-D1 deformation caused folding in bedding and led to varied orientation of intersection with D1 cleavage and possibly affected the orientation of boudins during D1 phase of deformation. (4) Rotation of D1 lineations on the cleavage plane: progressive pure or simple shear caused migration of D1 lineations toward the X deformation direction within the flattening (cleavage) plane (Ramsay, 1967; Sanderson, 1973; Escher and Watterson, 1974; Bell, 1978; Ramsay, 1979; Rhodes and Gayer, 1979; Cobbold and Quinquis, 1980; Ramsay, 1980; Skjernaa, 1980; Coward and Potts, 1983; Ridley, 1986). Clustering of the fold axes and hingelines along a constantly oriented cleavage is spatially controlled and occurs in subdomains where boudins are also clustered, supporting hypotheses 1 and 3 (Babaie, 1984). No lineation attributed to shear, to satisfy the fourth hypothesis, is shown in these rocks. Hypothesis 2 cannot be ruled out.



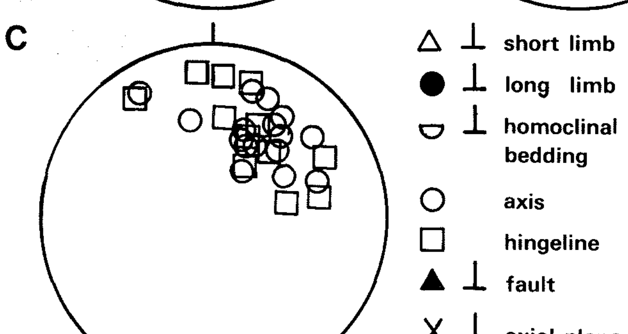


Figure 5. Orientation data of F2 folds along the boundary between slices 4, 5, and 9. (A) Folded bedding on the short and long limbs. (B) Axial planes of folds with different vergence and the associated faults. (C) Hingelines and axes of the folds.

Syn-Thrust Deformation (D2 and D3)

After the formation of F1 folds and cleavage, rocks were thrust in slices during D2 (Fig. 3). The age of rocks within each slice is not uniform, indicating intraslice folding probably during F1 phase and (or) faulting during D2. Local mesoscopic folds (F2), cut in places by faults, occur at the fault contacts of slices 4, 5, and 9 (Fig. 2). Poles to coplanar cleavage and bedding, in F2 fold closures (Fig. 5A), form approximately a great circle with large spread, indicating a folding event younger than isoclinal F1 folding. The poles to homoclinal bedding and cleavage, away from the thrust where F2 folds are absent, do not lie on the F2 phase bedding girdle (Fig. 5A) indicating that bedding was rotated, probably during D2 faulting, before becoming folded at the thrust boundary. The distribution of the poles to F2 axial planes (Fig. 5B) is approximately a great circle subparallel to that of the poles to cleavage and bedding,

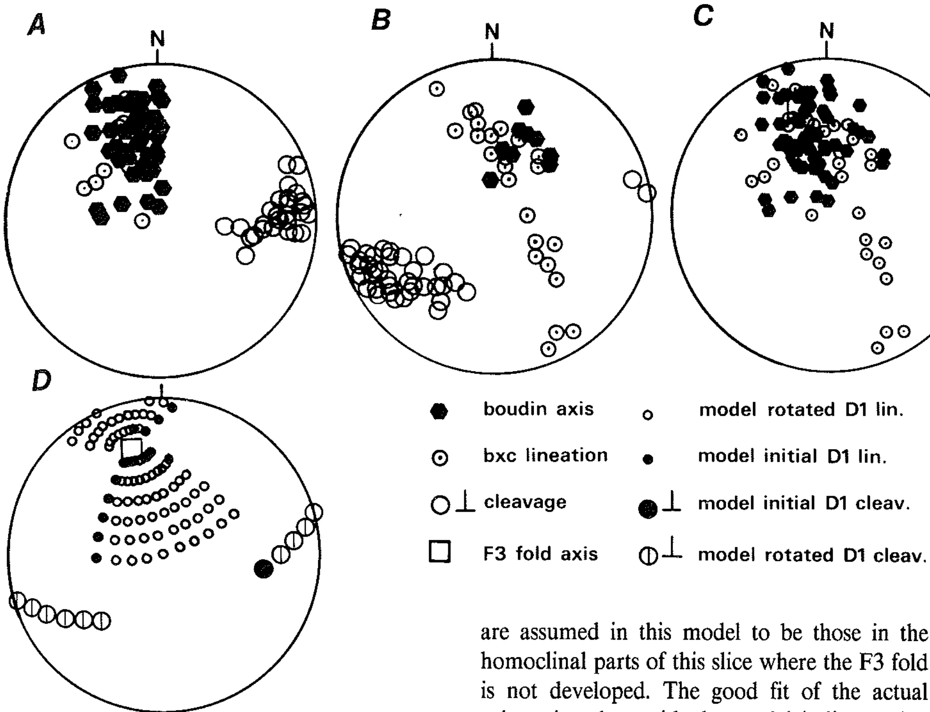


Figure 6. Deformation of F1 structures by F3 folds. (A and B). Boudin and bxc lineations on the westerly and northeasterly dipping limbs of the macroscopic F3 fold in Ophir Canyon. The large open circles represent poles to cleavage on the two limbs (coplanar bedding is not shown). (C) Summary diagrams of the F1 structures on both limbs. (D) A flexural-slip model to explain the distribution of F1 lineations and cleavage (of Figs. 6A–6C) by F3 folds (see text for explanation).

indicating rotation of the folds. Figure 5C depicts the variable orientation of fold axes and hingelines of the F2 folds. The presence of local faults cutting across these folds (Fig. 5B) and the sole occurrence of these folds at the slice boundary suggest that F2 folds are syn-thrust structures.

Slices were gently and openly folded (F3) during the D3 phase, leading to the variability in cleavage and bedding orientation in all of the slices (Figs. 2, 3, 6A, and 6B). The statistical axes of F3 folds plunge shallowly to moderately northwest. The best example of the F3 folds is a fold in Ophir and Wisconsin canyons in slice 4 (Fig. 6). The fold is an isolated single halfwave with halfwavelength of at least 200 m. The F1 lineations and poles to bedding and cleavage have rotated about the axis of the F3 fold in small and great circles, respectively (Figs. 6A–6C). The distribution can best be described by a flexural slip model as depicted on Figure 6D. The initial F1 linear and planar structures are assumed in this model to be those in the homoclinal parts of this slice where the F3 fold is not developed. The good fit of the actual orientation data with the model indicates that there was flexural-slip kinematics during F3 folding. This idea is further supported by the strain data to be discussed later. F3 folds, although homoaxial with average undeformed F1 lineations, are later structures, because the large range of the F1 cleavage around F3 folds (Figs. 6A, 6B) cannot be accounted for by fanning, and because these folds deform the slice-bounding D2 thrusts.

Post-Thrust Deformation (D4 and D5)

Slices dominated by slate and thin-bedded chert, such as those containing hemipelagic rocks, are kink folded (F4). Kink bands in slate (Fig. 7A) are continuous along their axial planes with folds of more rounded profiles (Fig. 7B) in

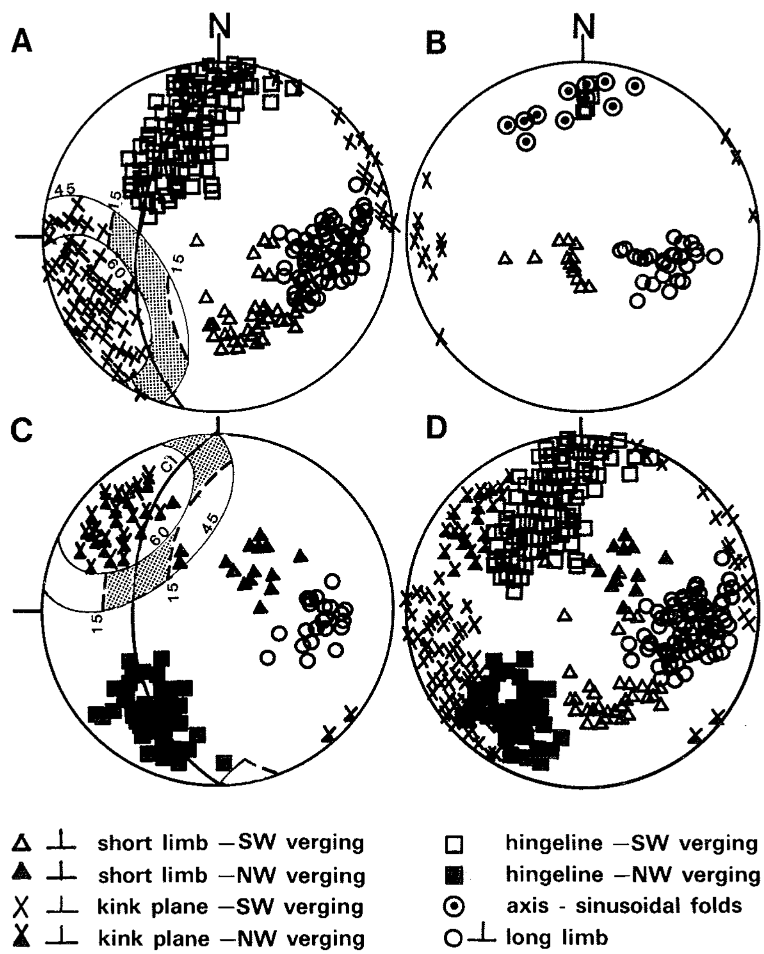


Figure 7. Orientation data of F4 (A) southwest-verging kinks, (B) dominantly west-verging sinusoidal folds, (C) northwest-verging kinks, and (D) all northwest- and southwest-verging kinks in slice 4. Note: "SW-verging" and "NW-verging" refer only to the vergence of the axial planes of the folds. Shaded areas show the directions of shortening during kink folding measured assuming that they lie (1) within 15° of the modal cleavage (cl) and (2) within 45° to 60° of the mean kinks planes. The solid small circles are drawn at 30° and 45° to the pole to the mean kink planes (Babaie and Speed, unpub. data). Solid great circles are modal cleavage (cl). The dashed small circles are at 75° to pole to the modal cleavage.

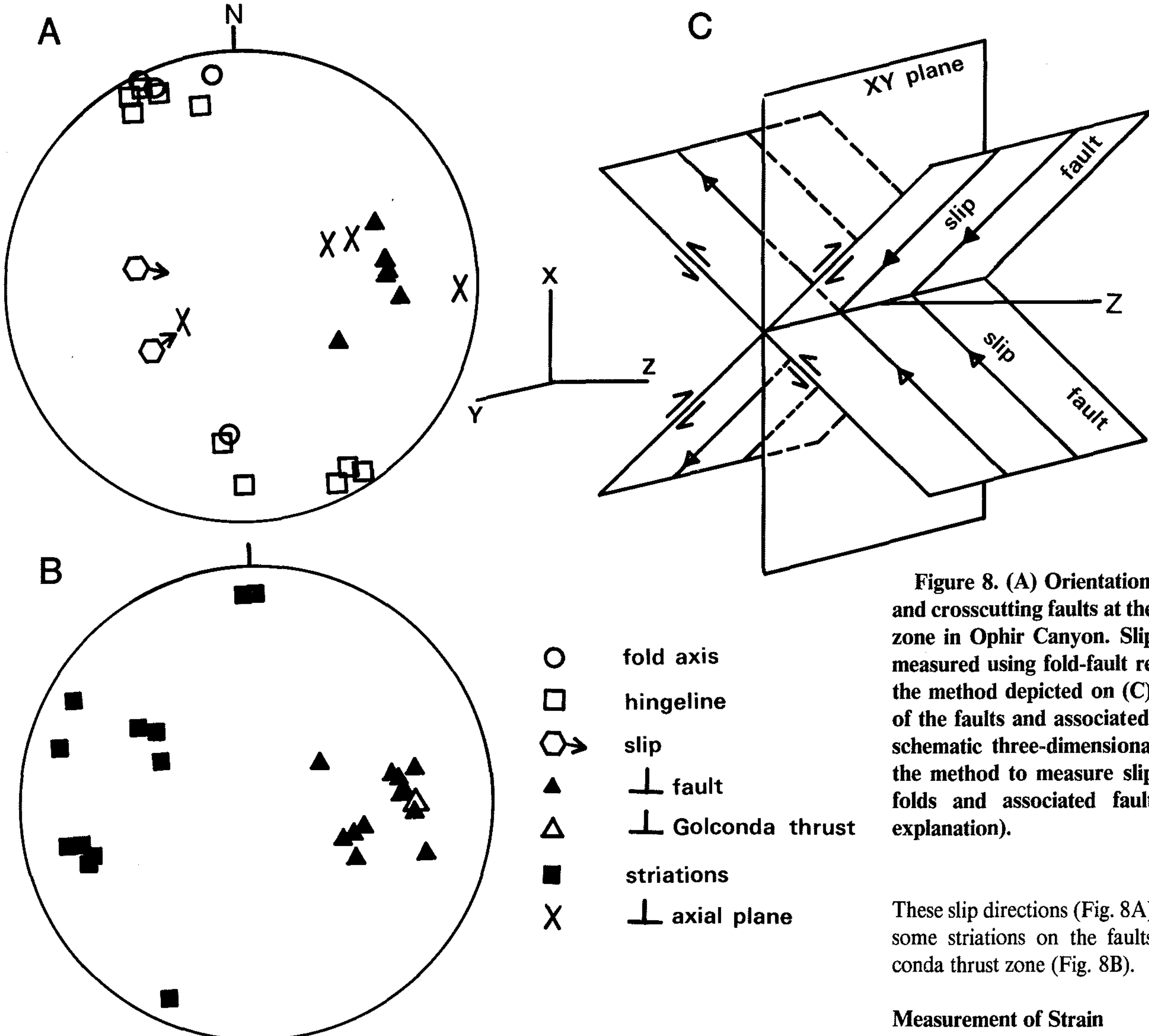


Figure 8. (A) Orientation data of F5 folds and crosscutting faults at the Golconda thrust zone in Ophir Canyon. Slip directions were measured using fold-fault relationships using the method depicted on (C). (B) Orientation of the faults and associated striations. (C) A schematic three-dimensional presentation of the method to measure slip directions from folds and associated faults (see text for explanation).

chert interbeds (sinusoidal folds). Orientation data of kink bands and sinusoidal folds in the southern part of slice 4 are given in Figure 7. Kinks are younger than the mascrosopic F3 folds that deform the slices and older than the latest movement of the Golconda thrust (D5) where they are deformed. Orientation of the kink bands and possible maximum principal shortening resolved from them indicate two kink folding events in the allochthon with orthogonal directions of contraction (Babaie and Speed, unpub. data). The earlier kink bands indicate an east-west shortening about parallel to the direction of tectonic transport of the allochthon probably during a final stage of emplacement (Figs. 7A, 7C). The later kink bands were formed due to northwest-southeast–oriented shortening of the allochthon either because of a nonrectilinear transport of the allochthon concurrent with earlier kinks or a later contraction not necessarily related to its emplacement (Babaie and Speed, unpub. data).

The Golconda thrust zone in Ophir and Wisconsin Canyons is characterized by the occurrence of mesoscopic folds (F5) that are cut by faults. Although one set is conjugate, F5 folds along the thrust have mainly east and east-northeast vergence, and their axial planes are at low angles to the associated faults (Fig. 8A, 8B). The faults at the thrust zone dip mostly west (Fig. 8B). The intersection of the faults and axial planes is subparallel to the hingelines at each location. Slip directions are approximated from the intersection of the XZ plane and fault assuming that (1) the faults and folds are contemporaneous, (2) the axial plane is parallel to the XY plane, and (3) the Y axis is defined by the intersection of the fault and axial plane (Fig. 8C). These slip directions (Fig. 8A) are subparallel to some striations on the faults within the Golconda thrust zone (Fig. 8B).

Measurement of Strain

Strain was measured in deformed quartz and chert grains from slates and sandstones in upper Paleozoic hemipelagic rocks (Fig. 2). For each specimen, the axial ratio of the strain ellipse was measured on three thin sections, perpendicular to foliation, using methods such as Rf/ϕ (Ramsay, 1967; Dunnet, 1969; Dunnet and Siddans, 1971; Lisle, 1977), polar graph (Elliott, 1970), Shimamoto-Ikeda (Shimamoto and Ikeda, 1976), and Robin (Robin, 1977), among others (Babaie, 1986). Assuming that cleavage is parallel to the XY plane of the strain ellipsoid, dimensions and orientations of the three-dimensional strain ellipsoid were calculated for the case where the quadratic elongations of three lines (trace of cleavage on the three thin sections) and the angles between them are known from two-dimensional data (Ramsay, 1967). The orientation of the principal strain axes and deformation

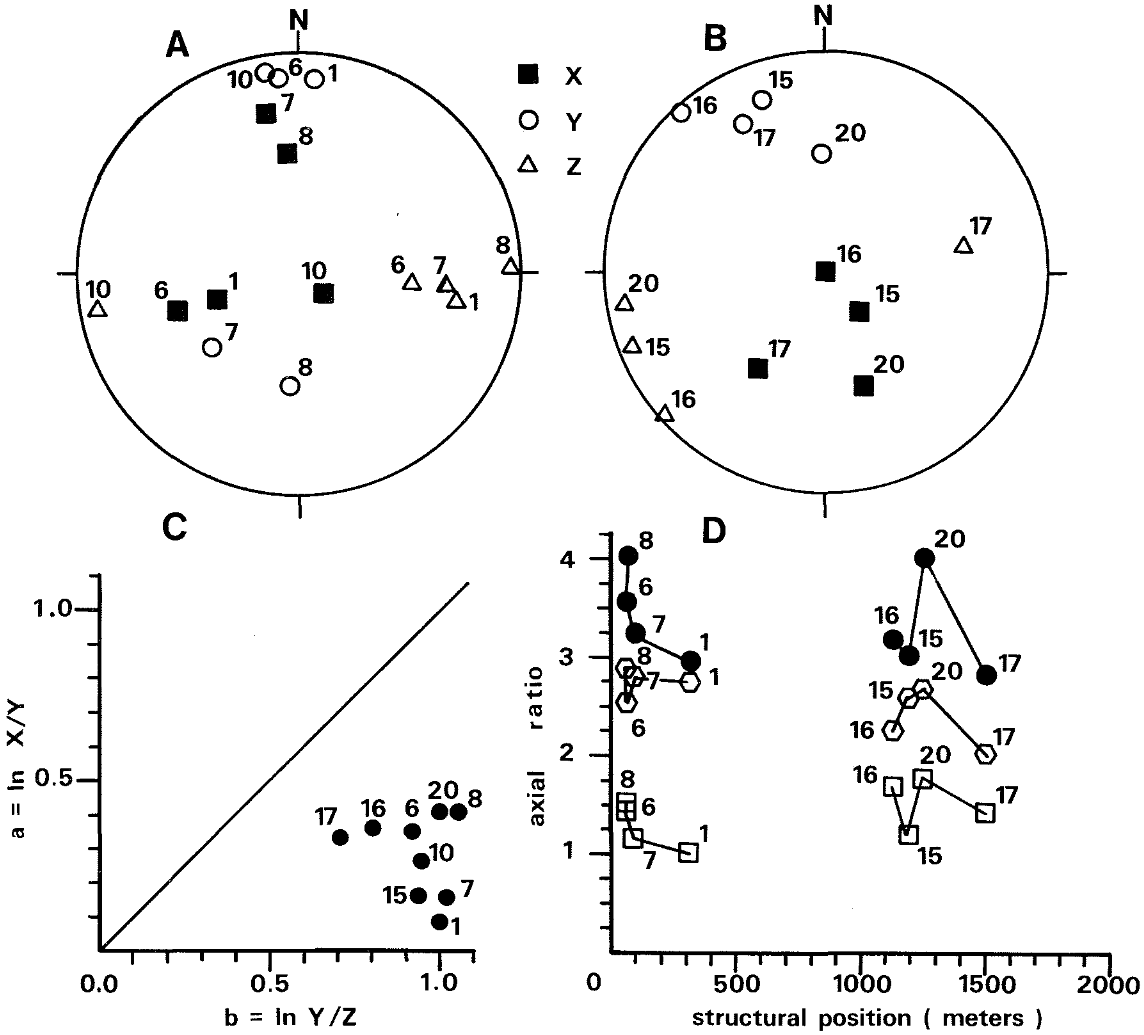


Figure 9. The principal directions for the specimens taken in (A) slice 4 and (B) slice 8. (C) Deformation plot for all the specimens in slices 4 and 8. (D) Spatial variation of the principal strain axial ratios as a function of distance of the specimens from the Golconda thrust. See Figure 2 for the location of the specimens.

plot of specimens in slices 4 and 8 are shown in Figure 9. All of the deformation ellipsoids plot in the apparent flattening field (Fig. 9C) (Flinn, 1962; Ramsay and Wood, 1973). The subparallelism of the XY planes (cleavage) and the axial planes of the F1 folds, as well as that between the Y axes of strain and F1 fold and boudin axes in each domain, indicate that the strain measured was developed mainly during D1 deformation.

Orientations of the X and Z axes in specimens 10 and 17 are rotated about the axis of the large-scale F3 folds which is homoaxial with the Y axis and F1 hingelines. This and the fact that magnitude of strain in specimens 10 and 17 is not significantly different from others in their corresponding thrust slices, support the conclusion that the strain is older than the F3 folding and that the F3 folds have deformed cleavage by flexural slip such that grain shapes remained constant and the X and Z axes rotated about the Y and F3 fold axes. No pattern is evident in the principal strain ratios as a function of structural position when thrust slices 4 and 8 are compared (Fig. 9D). The principal directions, however, vary within each slice (Figs. 9A, 9B). This can be explained by (1) superposition of a later, local incremental strain with its X and Y axes at high angles to those of the finite strain after the D1 phase, (2) variable displacement in the XY plane, and (3) rotation. Specimen 8, a sandstone, was sampled about 50 m away from specimen 7 near the Golconda thrust. The variation in the orientation of the X and Y in these two specimens can best be explained by cases 1 and 2 above, suggesting a local effect of the Golconda thrust on the finite strain. Rotation due to a hidden fault is also possible.

Significance of the Fabric and Strain Data

A summary of the deformation events in the Golconda allochthon is given in Table 2. In brief, the axes of F1 and F3 folds are subparallel in many slices, and plunge shallowly and moderately north-northwest (Figs. 2, 4, 6). F2 folds occur at slice boundaries and are syn-thrust structures. Axial planes of F1 folds strike between north and north-northwest and dip moderately to steeply southwest and northeast, subparallel to the axial planes of F4 southwest-verging kink bands and sinusoidal folds (Figs. 4E, 7A, 7B). The axial planes of the F5 folds at the Golconda thrust zone strike mainly between north-northeast and northwest, dipping steeply and moderately to west and southwest (Fig. 8). These fold orientations and the strain data are compatible with a generally east-west contrac-

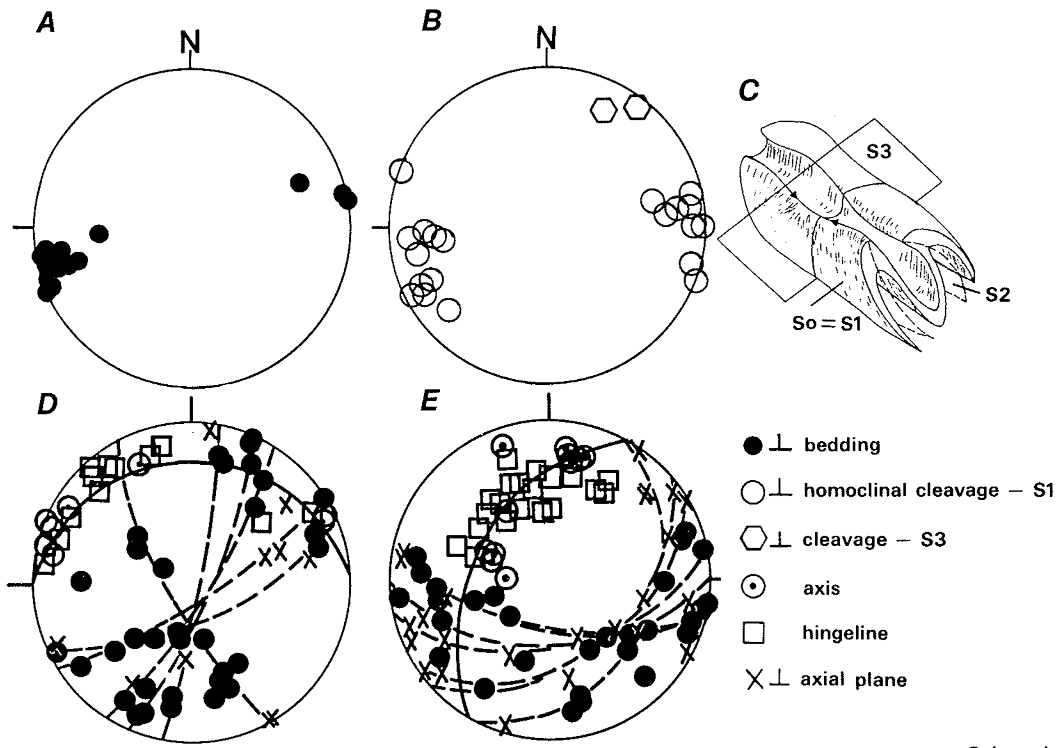


Figure 10. Orientation data of (A) homoclinal bedding, (B) homoclinal cleavage (S1) and younger S3 cleavage, and (D and E) folds in two domains of lower Paleozoic rocks of slice 7. (C) A schematic three-dimensional view of S1, S2, and S3 fabrics. Dashed great circles are the best fit planes containing poles to folded bedding and axial planes for each individual fold. Solid great circles are the best fit girdles of axes and hingelines for each domain.

tion during a sequence of events that persisted during the time interval represented by D1 through D5.

The results of this study are in many ways similar to those of Speed (1977a and unpub. data), MacMillan (1972), Miller and others (1981, 1982, 1984), and Brueckner and Snyder (1985), who studied structural fabrics of the Golconda allochthon. These works share the views of the present study by indicating one or all of the following: (1) a generally eastward movement of the allochthon above the Golconda thrust on coeval shelf sediments, (2) the complex internal architecture of the allochthon represented by several folded and faulted slices of deformed rocks of different type and age, and (3) presence of a similar sequence of deformational structures in each of the thrust slices. Brueckner and Snyder (1985) report the presence of several features which indicate high pore-fluid pressures along slice-bounding thrusts that they attribute to tectonic and diagenetic processes in accretionary prisms. These features include crack-seal fractures, dilation breccias, and clastic dikes and sills. Although chert slices elsewhere in northern Nevada contain a bedding-parallel solution cleavage (Brueckner and Snyder, 1985), they apparently lack the well-developed pre-thrust slaty cleavage which is typical of the rocks of the allochthon in the southern Toiyabe Range. This may be a reflection of a higher strain and(or) the larger volume of pelitic rocks in the Toiyabe Range.

Slices of Lower Paleozoic Rocks

Among the identified and inferred lower Paleozoic rocks, slice 7 (Fig. 2) contains tectonic fabrics that differentiate it from those in upper Paleozoic rocks. Two generations of cleavage exist in slice 7 (Figs. 10A, 10B). S1 is parallel to bedding and is penetrative, indicating either F1 isoclinal folds with axial plane cleavage or burial of horizontal layers (compaction). S3 is sporadic, weakly developed, and oblique to bedding and S1. The S1 cleavage is folded by chevron, conjugate, and other types of folds and shows large variations in attitude (Figs. 10D, 10E). Orientation of the axes and axial planes (S2) indicates multiphase generation. For each subdomain, hingelines and axes of folds with different vergence display a girdle distribution. Local bedding and axial plane (S2) girdles have nearly colinear intersections that parallel the normal to the girdle of the hingelines and axes in each subdomain (Figs. 10D, 10E). This suggests either refolding, fanning of the axial planes, conjugate geometry, or a combination of these. The continuous girdle of axial planes makes fanning unlikely. A perfect box fold was recorded at one station, indicating that at least some folds are conjugate. Refolding is the most likely origin.

The relationship of S1, S2, and S3 is schematically shown in Figure 10C. S3 is about normal

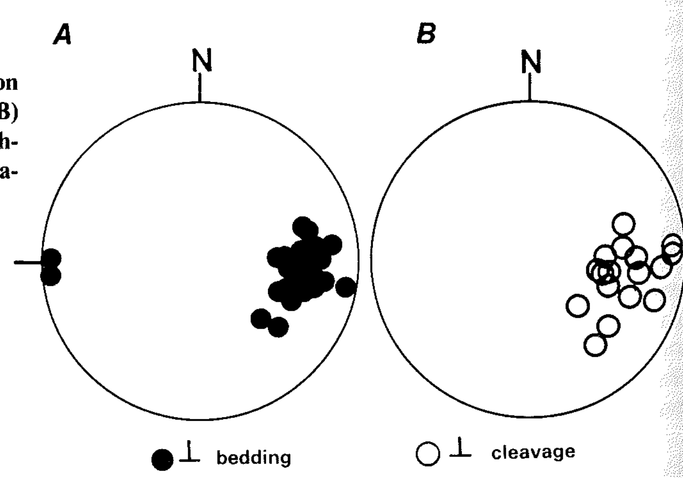


Figure 11. Orientation of (A) bedding and (B) cleavage in the autochthonous Diablo Formation.

to both the hingeline girdles at the two subdomains. The S3 cleavage and refolded folds are absent in the upper Paleozoic section of the allochthon and may in part be pre-Golconda structures. The refolding may at least partly be due to the deformation that occurred when these rocks were emplaced into the allochthon. Bedding and cleavage in slices of lower Paleozoic rocks (slice 1) are similarly oriented as those in the upper Paleozoic slices (Fig. 2).

Autochthonous Diablo Formation

Unlike deformed beds above the Golconda thrust, sandstones, conglomerates, and marbles of the Diablo Formation are homoclinal and are not affected by folds in the allochthon. In addition to tectonic reasons, this could be due to the thicker beds or lithology of the Diablo Formation. Foliation is the prevalent structure of the Diablo and is defined by the preferred orientation of tabular clasts of chert that lie parallel to subparallel to bedding and to an anastomosing foliation in the matrix defined by the alignment of platy minerals and finer planar grains (Fig. 11). The strong foliation in the chert conglomerate of the Diablo Formation can be attributed to either the emplacement of the Golconda allocohthon, burial compaction normal to bedding, or to intrusion of the Ophir pluton. The rocks of the Diablo Formation are thermally metamorphosed as indicated by the presence of marbles. An age younger than the Golconda thrust for the foliation is suggested by its homoclinal attitude, in contrast to the intensely deformed rocks in the Golconda allochthon and underlying lower Paleozoic rocks. Although there is no apparent increase in the penetrative nature of the foliation toward the pluton, intrusion of the Ophir pluton as the cause of formation of the foliation cannot be ruled out.

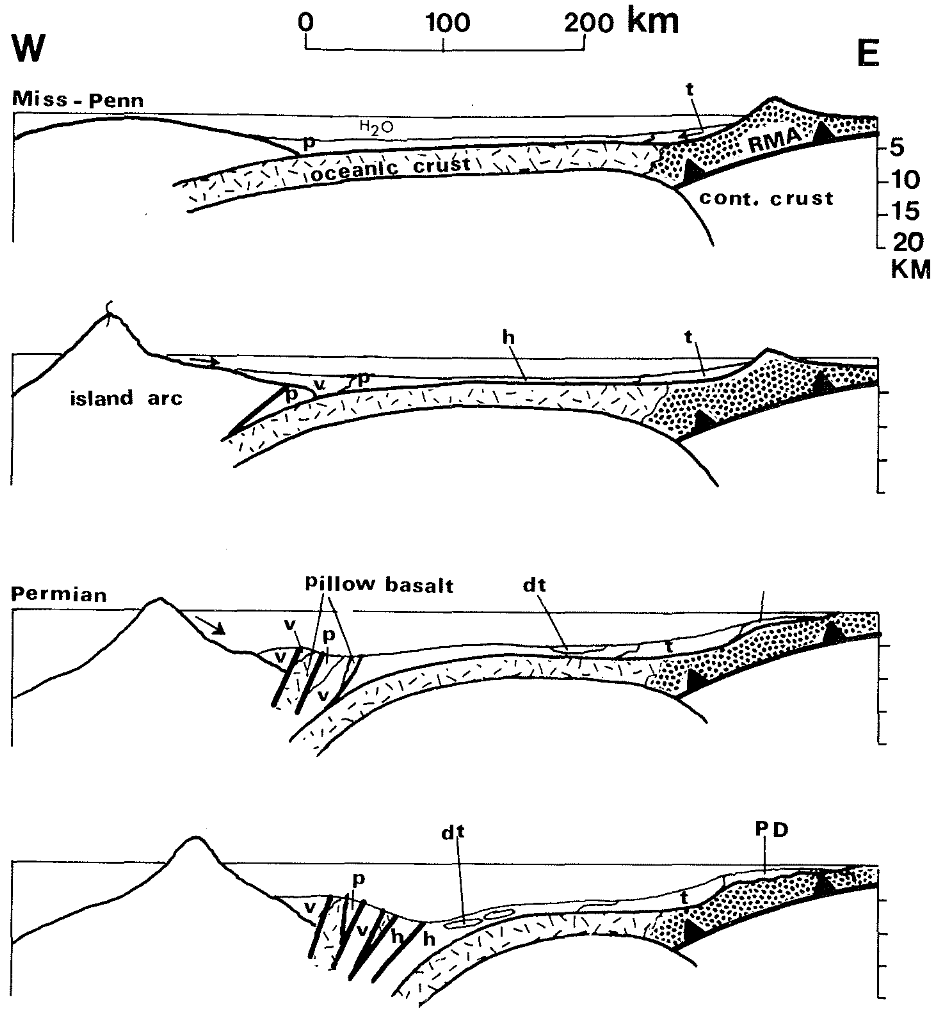


Figure 12. A forearc paleogeographic model of the western margin of continental North America in the latitude of Nevada. RMA = Roberts Mountains allochthon; PD = Permian Diablo Formation; t, h, v, and p refer to the upper Paleozoic terrigenous, hemipelagic, volcanogenic, and pelagic sediments, respectively. dt = distal turbidites. Arrows indicate directions of sediment transportation by gravity flows. The line with filled triangular teeth is the Roberts Mountains thrust. Heavy solid lines are slice-bounding thrusts in the Golconda accretionary prism.

DISCUSSION: A MODEL

The Golconda allochthon in the southern Toiyabe Range includes thrust slices of late Paleozoic clastic rocks which either are terrigenous and probably locally derived from the part of the North American margin upon which the allochthon was emplaced, or are volcanogenic and exotic to North America.

These diverse Mississippian-Permian deposits were subcoaxially folded and thrusted in several stages before being translated over the homoclinal Wolfcampian to Guadalupian outer-shelf Diablo Formation. This assigns a post-Guadalupian age for the emplacement of the slices and for the Golconda thrust. The stack of thrust slices itself was assembled over a long period of time between the Mississippian and Late Permian. The relatively undeformed Lower Triassic rocks that apparently occur structurally beneath the upper Paleozoic slices suggest that the contraction continued into the Early Triassic or possibly later. The upper bound for the age of the thrust is the 42- or 54-m.y. K-Ar age of the Ophir pluton that intrudes it. The lower Paleozoic thrust slices, occurring intercalated with and under the Lower Triassic rocks in the Golconda allochthon, probably indicate that further contraction included imbrication of the lower Paleozoic rocks along the continental margin. This may have occurred during the Sonoma or a later Mesozoic contraction.

Stacking of slices of terrigenous and volcanogenic rocks in the inner and outer parts of the allochthon, respectively, and of slices of oceanic hemipelagic and tholeiitic basalts in the middle of the allochthon suggest long transport distance of the allochthon; this conclusion was also reached by Brueckner and Snyder (1985) about northern Nevada. If all of the major upper Paleozoic facies were deposited in one basin, the data would not support the idea of a narrow backarc basin, where terrigenous sedimentation is more likely to extend across the width of the basin and mix with arc-derived sediments. A wide but narrowing backarc basin remains a possibility, however. The occurrence of the same

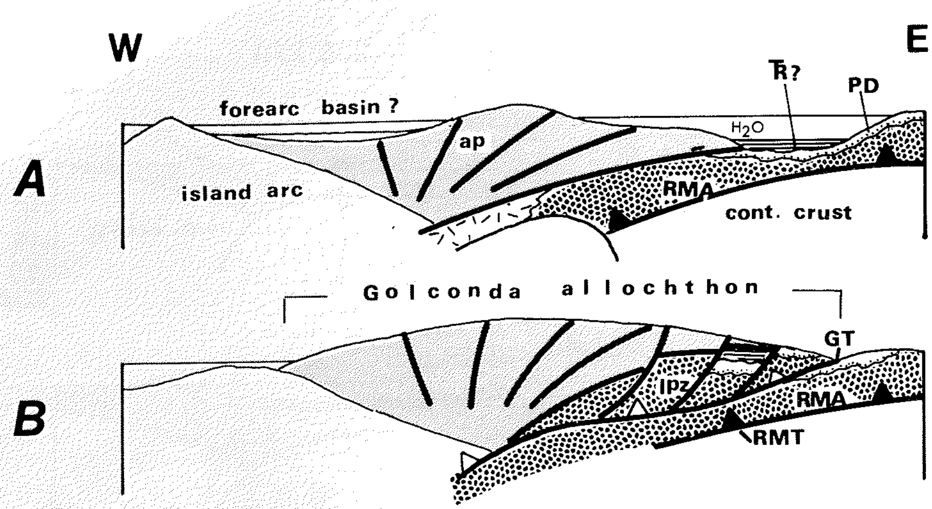


Figure 13. A model for the Early Triassic paleogeographic, tectonic, and depositional processes during earlier (A) and later (B) stages of overriding of the continental margin. PD = Permian Diablo Formation, TR = Triassic rocks, lPz = slices of lower Paleozoic rocks of the Roberts Mountains allochthon (RMA) in the Golconda allochthon, ap = accretionary prism of upper Paleozoic rocks. The Golconda thrust and Roberts Mountains thrust are shown by unfilled and filled triangular teeth, respectively. Heavy solid lines represent slice-bounding thrusts within the Golconda allochthon.

sequence of pre-, syn-, and post-thrust deformation in each thrust slice can best be accounted for by formation in an accretionary prism of a forearc basin as described in the following model based on that of Speed (1979). It should be noted, however, that this accretionary model is not the only possible one that can be constructed from the data.

From the latest Mississippian to Guadalupian, an oceanic basin fronted the western passive continental margin of the North American continent, probably at the Toiyabe latitude (Fig. 12). The basin received sediments shed both from the west-facing continent, dominated by the then accreted Roberts Mountains allochthon, and from an island arc above a west-dipping subduction zone. The continental shelf was dominated by the deposition of carbonate and siliciclastic rocks of the Diablo Formation between the Pennsylvanian and Guadalupian and possibly during the Mississippian. Terrigenous debris was deposited at the shelf, slope, and rise of the continental margin. The volcanogenic debris, carried east from the arc, supplied sediments to the distal parts of the ocean floor probably at the trench where pelagic sedimentation was dominant. Presence of an active spreading center in the basin is uncertain because the age of pillow basalts, upon which sediments were laid down, is unknown. The magmatic arc was under active erosion at least since the Pennsylvanian, as indicated by the Pennsylvanian volcanogenic sediments. East-west convergence of the arc and North American continent resulted in sequential tectonic stacking of the upper Paleozoic rocks probably into an accretionary prism that was translated across the continental shelf sometime between the Guadalupian and Early Triassic (Fig. 12).

Elastic loading of the lithosphere by the prism may have resulted in downwarping of the margin that included the Diablo Formation and lower Paleozoic rocks. This may have led to the formation of an Early Triassic foreland basin at the toe of the beached prism similar to a model proposed for the Roberts Mountains allochthon by Speed and Sleep (1982) (Fig. 13). This Early Triassic basin may have received sediments shed east from the Golconda accretionary prism and west from the continent as suggested by the lithology of the Triassic rocks. The Lower Triassic rocks were later overthrust by the upper Paleozoic section due to continued Triassic or possibly later contractions. These compressions also led to thrust imbrication of the lower Paleozoic rocks of the continental margin and their incorporation into the base of the upper Paleozoic–Lower Triassic stack to form the Golconda allochthon (Fig. 13B). The allochthon and its relative autochthon were intruded by the Tertiary Ophir pluton and covered by Tertiary volcanic rocks. Finally, Tertiary Basin and Range tectonism, mainly by rotation, affected all rocks of the area.

ACKNOWLEDGMENTS

I thank Bob Speed for continuous support, valuable discussions and criticism and review of the manuscript. This paper was also improved through reviews by Norm Silberling, Walter Snyder, Gregory Davis, and Tim LaTour, whose comments are greatly appreciated. Identification of conodonts by A. G. Harris and radiolaria by B. M. Murchey and D. L. Jones is gratefully acknowledged. Field assistance by A. A. Babaie and S. Babaie is appreciated. This work has been supported by Gulf Oil Co., and National Science Foundation Grant no. EAR-7911150 awarded to R. C. Speed.

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MANUSCRIPT RECEIVED BY THE SOCIETY JUNE 2, 1986
REVISED MANUSCRIPT RECEIVED FEBRUARY 2, 1987
MANUSCRIPT ACCEPTED FEBRUARY 10, 1987