LOWER PALEOZOIC AND UPPERMOST PRECAMBRIAN CORDILLERAN MIOGEOCLINE, GREAT BASIN, WESTERN UNITED STATES[1]

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ABSTRACT

Shallow-marine intertidal and supratidal detrital and carbonate strata of latest Precambrian (<850 my) and early Paleozoic (>345 my) age thicken from a few hundred meters in cratonic areas east of the Great Basin to nearly 10,000 m (c. 30,000 ft) in the central Great Basin 350 to 450 km (c. 250 mi) to the west. Coeval rocks in the western Great Basin are deep-water strata characterized by shale and radiolarian chert associated with mafic pillow lavas. Strata deposited at moderate depths between the shallow- and deep-water facies have a limited distribution that suggests a relatively abrupt transition from shelf to deep water. The thick accumulation of shallow-water deposits in the Great Basin is similar to deposits along present-day stable continental margins. Such accumulations have been termed miogeoclines, rather than miogeosynclines, because they are bordered outward by open ocean, or a marginal sea, and are not synclinal in form.

The continental margin along which the early Paleozoic and latest Precambrian miogeocline was constructed apparently developed by rifting in late Precambrian time (<850 my). Extensional faulting and flowage related to this rifting extended well into the continent and may have caused major crustal thinning as far east as the Wasatch line, across which the rate of westward thickening of uppermost Precambrian and Paleozoic strata increases markedly. A persistent positive belt, perhaps analogous to the buried ridge beneath the outer edge of the present-day Atlantic continental shelf, may account for regional thinning and local erosional truncation of lower Paleozoic strata along the western margin of the Cordilleran miogeocline.

INTRODUCTION

The extensively exposed lower Paleozoic and uppermost Precambrian strata in the Great Basin are an excellent record of ancient sedimentation along a continental margin. The region described extends from the Wasatch Mountains in central Utah to the Sierra Nevada in westernmost Nevada and eastern California and from the Snake River Plain in southern Idaho to the Mojave Desert in California. The total area is about 5×10^5 km.[2] The completeness of the record within the region contrasts with that in adjoining areas where upper Precambrian and lower Paleozoic rocks either are less extensively exposed, are covered by younger rocks, or occur in structurally complex settings where stratigraphic details are difficult to interpret. The purpose of this paper is to summarize the upper Precambrian and lower Paleozoic stratigraphy in the Great Basin, with particular attention to characteristics that indicate environment of deposition and tectonic setting.

The strata described range in age from latest Precambrian[2] (<850 my) to Late Devonian (>345 my) and were deposited during a time of relative tectonic stability following a marked change in the tectonic framework of the region in latest Precambrian time. Deposition preceded the Antler orogeny, a major deformational event of Late Devonian and Early Mississippian time (Poole, this volume). The deposits lie mostly within the Cordilleran geosyncline (fig. 1), the dominating structural feature of western North America during late Precambrian, Paleozoic, and early Mesozoic time.

Various classifications have been proposed for major belts of Paleozoic sediments in the Cordilleran geosyncline in the Great Basin. Geologists have long recognized that the eastern part of the geosyncline contains carbonate and quartzite rocks, and the western part chert, shale, and volcanic rocks. The terms "eastern" and "western" facies or assemblages have been applied to these belts of rocks (Merriam and Anderson, 1942, p. 1704; Nolan and others, 1956, p. 6, 23, 34; Roberts and others, 1958, p. 2816-2817). These same facies were also noted by Kay (1951), who recognized an eastern miogeosynclinal (Millard) belt and a western eugeosynclinal (Fraser) belt of the Cordilleran geosyncline. Roberts and others

[1] Publication authorized by the Director, U.S. Geological Survey.
[2] Strata referred to here as latest or uppermost Precambrian correspond in large part or entirely to the Precambrian Z age (800 to 570 my) of James (1972).

(1958) described rocks of an intermediate facies that crop out in areas between the two major assemblages and referred to these rocks as the "transitional" assemblage. More recently the terms "carbonate" and "siliceous and volcanic" (or simply "siliceous" have replaced the terms "eastern" and "western" in the nomenclature (Roberts, 1964, p. A8, 1968a, 1972; Gilluly and Gates, 1965; Smith and Ketner, 1968). Roberts (1968b), in describing Silurian and Devonian strata, further divided each of the major assemblages into two subfacies.

This report uses the threefold facies nomenclature of common usage—carbonate, transitional, and siliceous assemblages; we further divide the siliceous assemblage in the Great Basin into two subassemblages, the inner and outer belts (table 1). Assignment of a particular rock unit to one assemblage or another is not always clear; some rocks called transitional here are part of the carbonate assemblage of other geologists.

The eugeosynclinal (Fraser) belt, as defined by Kay (1951), includes lower Paleozoic rocks that outcrop in the northern and western Sierra Nevada and in the Klamath Mountains in California as well as in the Great Basin. In this report, only those rocks of this belt that outcrop in the Great Basin are described.

No entirely satisfactory nomenclature for geosynclines is available. Dietz and Holden (1967) and Dietz (1972) have pointed out that a modern analogy of a geosyncline is a deposit along a stable continental margin where a seaward-thickening wedge of sediment underlies the continental shelf and is limited seaward by the continental slope. Farther oceanward, sediment in this system consists of continental-rise and ocean-basin deposits. This system does not have the classic form of a geosynclinal trough bounded on both sides by landmasses, or bounded on one side by a landmass and on the other by an island arc system. Dietz and Holden (1966) and Dietz (1972) proposed the term miogeocline to describe the wedge of sediment underlying the continental shelf and slope, and the term eugeocline to describe continental rise deposits. They used these terms instead of miogeosyncline or eugeosyncline in order to indicate the nonsynformal shape of the deposit.

In this report we use the term miogeocline to describe the wedge-shaped deposit formed by lower Paleozoic and uppermost Precambrian carbonate and transitional assemblage rocks in the Great Basin. We do not intend, however, that this practice should prohibit use of the term miogeosyncline in a more general sense to describe thick sequences of nonvolcanic sediments in geosynclines of different character. The term eugeocline is awkward etymologically, meaning "true earth slope," but is tentatively used in the Great Basin to describe lower Paleozoic siliceous assemblage rocks that may be continental rise deposits. The term eugeosyncline is used in a general sense to describe the volcanic part of a geosyncline.

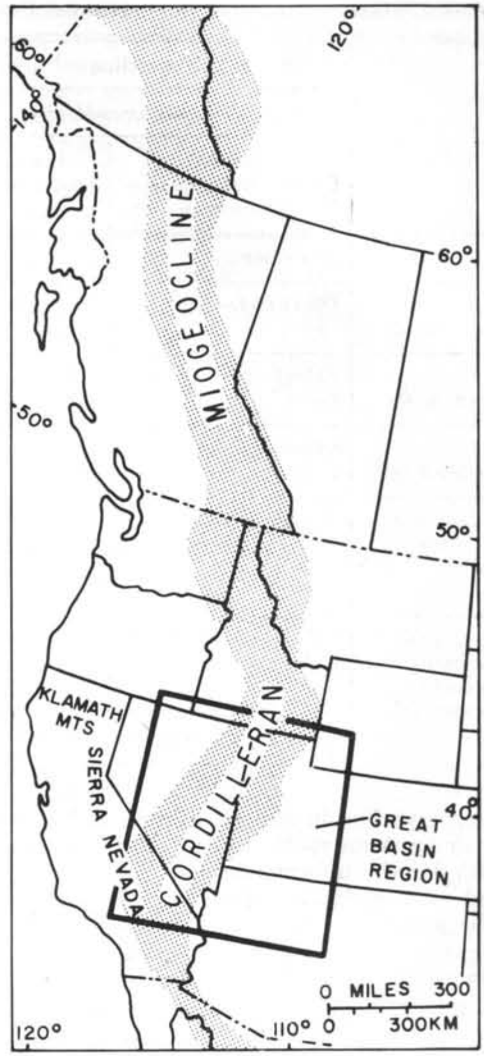


FIG. 1.—Map of western North America showing location of Cordilleran miogeocline and Great Basin region.

This paper is based on our own studies as well as on published reports by hundreds of geologists. Where possible we have acknowledged the work of others, but we have not indicated the sources of all the data on the isopach maps. The excellent summary of the Cordilleran

TABLE 1.—LOWER PALEOZOIC AND UPPERMOST PRECAMBRIAN ASSEMBLAGES AND FACIES IN THE GREAT BASIN.

AGE	Eugeocline		Miogeocline	
	Siliceous assemblage		*Transitional assemblage* Limestone and shale, including quartzite in upper Precambrian and Lower Cambrian	*Carbonate assemblage* Includes largely detrital rocks in upper Precambrian and Lower Cambrian
	Outer belt Chert, shale, quartzite, greenstone, arkose, and feldspathic sandstone	*Inner belt* Shale and chert		
DEVONIAN	Chert facies	Shale and chert facies	Limestone and shale shale	Carbonate and quartzite facies
SILURIAN	Feldspathic sandstone facies	Chert and shale facies	Laminated limestone facies	Dolomite facies
ORDOVICIAN	Siliceous and volcanic facies	Shale and chert facies	Shale and limestone facies	Carbonate and quartzite facies
LATE AND MIDDLE CAMBRIAN	Arkosic facies	—	Limestone and shale facies, includes abundant chert in Emigrant Formation	Carbonate facies
LATEST PRECAMBRIAN AND EARLY CAMBRIAN	Siliceous and volcanic facies	—	Siltstone, carbonate, and quartzite facies	Quartzite and siltstone facies

miogeosyncline in eastern Nevada and western Utah by Armstrong (1968a) is of great value both for the information and ideas it contains and as a guide to methods of presenting data. Our palinspastic base maps are similar to those prepared by Armstrong. We are indebted for discussions with many colleagues, some of whom have supplied unpublished data. The report has benefited from thoughtful comments of Michael Churkin, Jr., M. D. Crittenden, Jr., W. R. Dickinson, H. Gabrielse, R. K. Hose, K. B. Ketner, F. J. Kleinhampl, and T. E. Mullens.

REGIONAL STRUCTURE AND PALINSPASTIC BASE MAP

Original geographic positions of upper Precambrian and lower Paleozoic strata in the Great Basin have been severely disrupted by major displacements on thrust and strike-slip faults (fig. 2) and distorted by oroflexural bending. To compensate for this disruption and distortion, thickness and facies data in this report are plotted on a palinspastic base map (fig. 3).

The oldest structure shown on the palinspastic base map is the Roberts Mountains thrust along which siliceous and some transitional assemblage rocks have been transported about 145 km (90 mi) to the east primarily over carbonate and transitional assemblage rocks during the Late Devonian and Early Mississippian Antler orogeny (Roberts and others, 1958; Roberts, 1968a, p. 106; 1972, p. 1995; Smith and Ketner, 1968). On the palinspastic base, the reconstructed map position for upper plate (siliceous assemblage) rocks is an area where standard county line symbols are used and is west of an area, shown with dotted county lines, where lower plate rocks occur. In westernmost Nevada, only rocks in northern Esmeralda County and southern Mineral County are considered here to be allochthonous and part of the upper plate of the Roberts Mountains thrust; McKee (1968), on the other hand, has suggested that Ordovician rocks in southern Esmeralda County may also be allochthonous and part of the upper plate, an interpretation not followed here.

The western Great Basin is an area of 130 to 195 km (80 to 120 mi) of cumulative right-lateral distortion resulting from fault slip and more pervasive large-scale drag from oroflexural bending (Albers, 1967; Stewart, 1967; Poole and others, 1967; Stewart and others, 1968), although other geologists (Wright and

Troxel, 1966, 1967, 1970) have questioned this concept. Our palinspastic base was constructed on the assumption of about 160 km (100 mi) of right-lateral displacement in a general northwest direction. More than half of the displacement is considered to be by oroflexural bending, which can be detected on the palinspastic base by curving of originally straight county and state lines; the remainder, by fault displacement on two major right-lateral strike-slip fault zones, the Las Vegas Valley shear zone and the Death Valley-Furnace Creek fault zone.

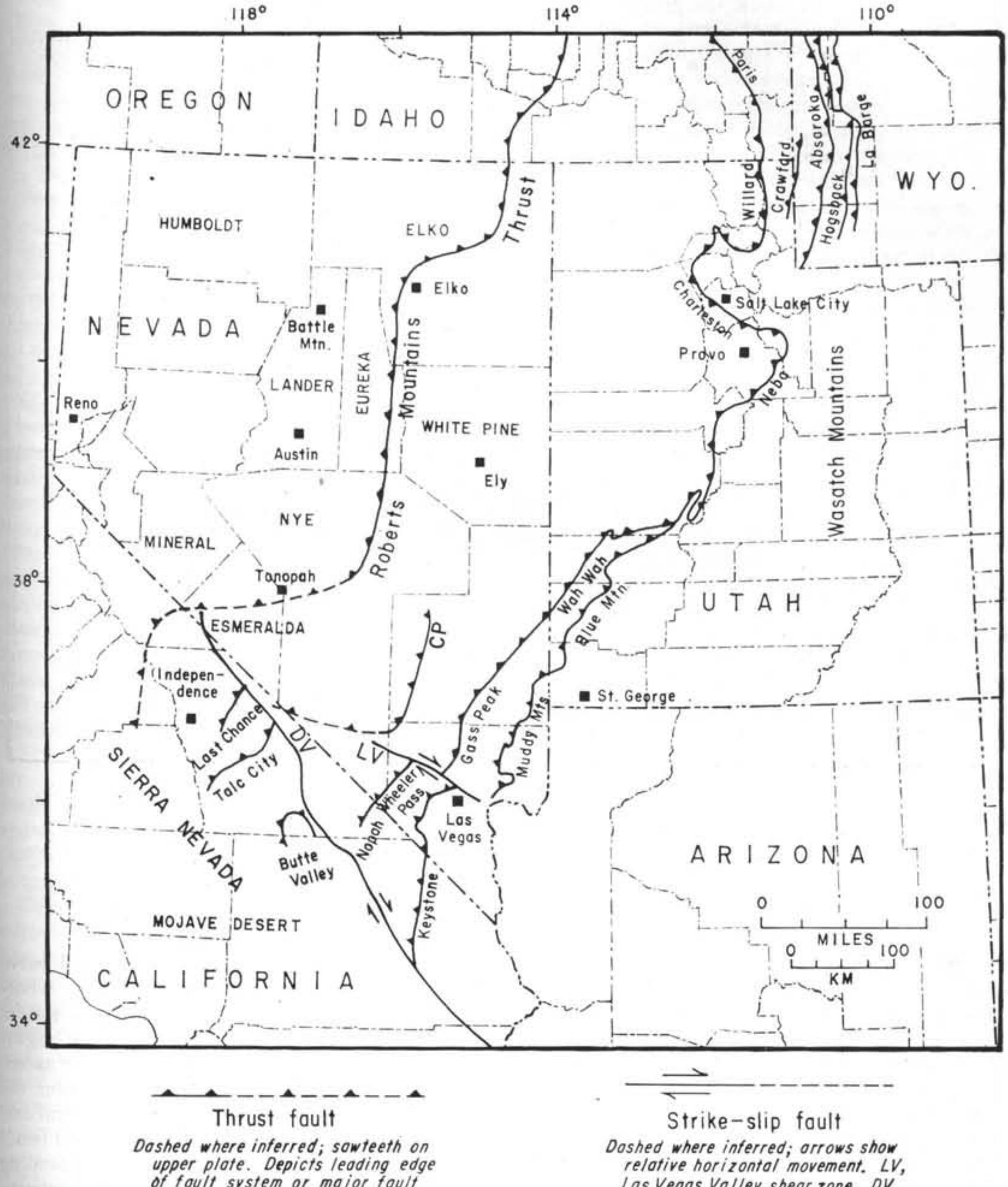
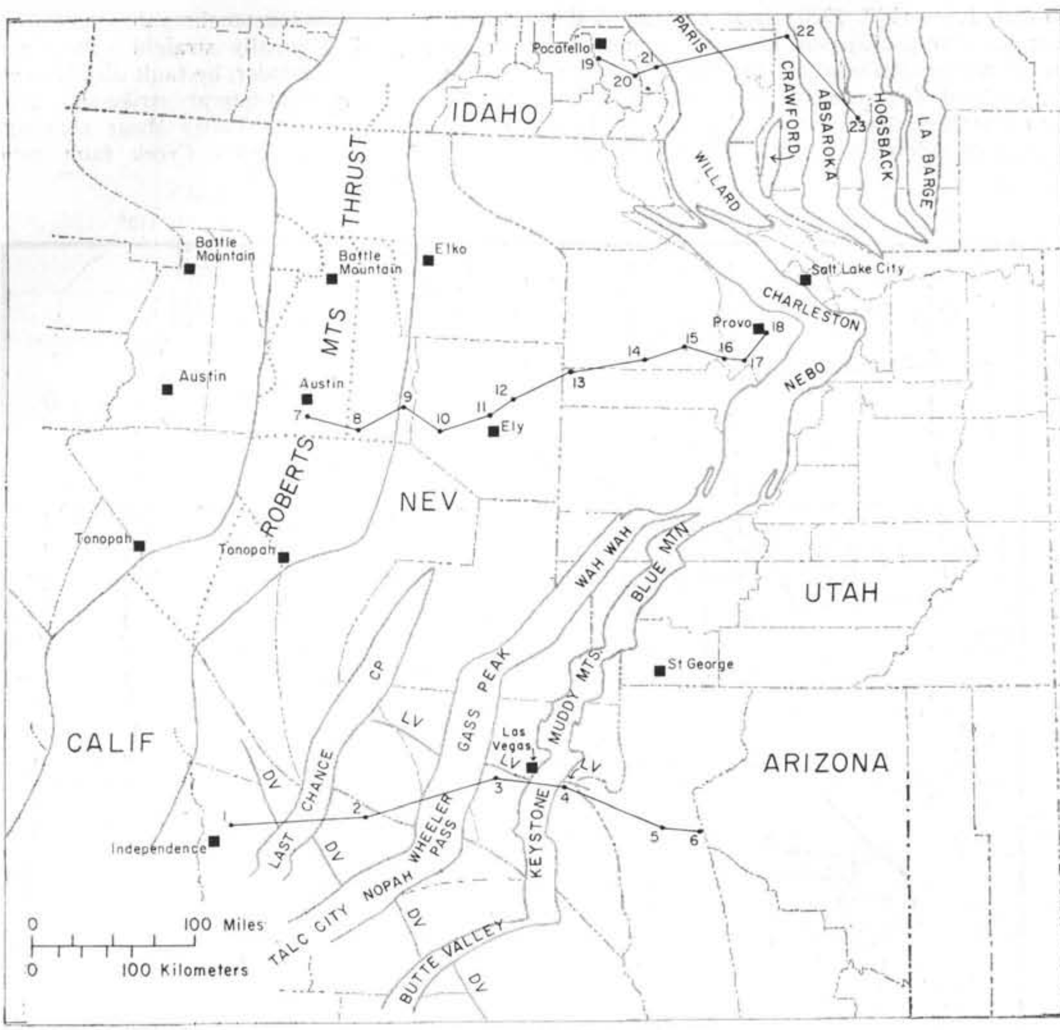


FIG. 2.—Index map of Great Basin region showing major thrust and strike-slip faults. Only counties referred to in text are labeled.



EXPLANATION

LV, Las Vegas Valley shear zone
DV, Death Valley – Furnace Creek fault zone

FIG. 3.—Palinspastic base map showing location of cross sections shown in figures 5, 6, 7 and 8. Dotted county lines indicate location of lower plate of Roberts Mountains thrust.

The displacement on the Las Vegas Valley shear zone occurred in late Tertiary time (Ekren and others, 1968; Fleck, 1970a; Anderson and others, 1972, p. 284), and perhaps much of the other right-lateral displacement in the western Great Basin is of this age.

Another oroflexural feature may occur in northeastern Nevada, where lower and upper Paleozoic facies and thickness trends and the leading edge of the Roberts Mountains thrust curve northeastward at latitude 41° for 120 km (75 mi) before regaining a more northerly direction. The curving of these trends has generally been assumed to mark an original curve in the Cordilleran geosyncline, although Thorman (1970, p. 2441) has suggested that displacement along major northwest-trending right-lateral strike-slip faults may account for anomalous facies trends in the area. We prefer the interpretation that the curving of structural and stratigraphic features is due to oroflexural bending. Several regionally anomalous east-trending fold structures (Hope, 1970; oral commun., 1973) near latitude 41° N. are compatible with the presumed oroflexural feature.

A system of thrust faults extends along the eastern part of the Great Basin and the adjoining part of the Rocky Mountains to the east. In Wyoming, this belt consists of several major thrusts along each of which tectonic transport may have been about 15–25 km (10–15 mi) subject to a factor of uncertainty of about two, either way, according to Rubey and Hubbert (1959, p. 187). S. S. Oriel (oral commun., 1973) considers Rubey and Hubbert's estimate of displacement to be conservative. Our reconstruction is based on an assumed total tectonic transport of 120 km (75 mi) across the thrust belt in Wyoming. In Idaho and Utah, we used 65 km (40 mi) as the amount of transport on the Paris, Willard, Charleston, and Nebo thrust system; this figure appears to be a minimum based on the estimates of Crittenden (1961), Rigo (1968, p. 64), and Armstrong (1968b, p. 440–441). Estimates of displacement on faults farther south in the thrust belt are less certain. A minimum of 30–35 km (c. 20 mi) is required on the combined Muddy Mountains and Glendale thrusts in southern Nevada (Longwell, 1961, as quoted in Armstrong, 1968b, p. 440). Fleck (1970b) has suggested a total displacement of 30 to 50 km (18 to 32 mi) on the Keystone, Wheeler Pass, and associated minor thrusts. A minimum of 30-35 km (c. 20 mi) has been suggested on the Last Chance thrust (Stewart and others, 1966), although Stevens and Olson (1972) have a somewhat different interpretation of the geometry of this thrust. Tectonic transport on the CP thrust has been inferred to be about 55 km (35 mi; Barnes and Poole, 1968). We show a total of about 120 km (75 mi) telescoping on thrust faults in southern Nevada and adjacent California.

The correlation of thrust faults in the southern Great Basin follows the interpretation of Poole and others (1967, p. 891–892). Burchfiel and Davis (1972) have presented a somewhat different picture and interpretation of thrusting in this region.

Crustal shortening due to folding and to telescoping on seemingly minor thrust faults has not been incorporated on the palinspastic map, although we realize that some of these structures could be of major significance. Nor have we included late Tertiary crustal extension, which resulted in the development of basin and range faulting and is commonly considered to be 50–100 km (30–60 mi) across the entire Great Basin (see summary by Stewart, 1971, p. 1035–1036) or even as much as 160 km (100 mi; Proffett, 1971). The factors of crustal shortening and extension that were not evaluated on the palinspastic base are compensating and perhaps could be nearly balancing.

PREGEOSYNCLINAL ROCKS

Three major divisions of Precambrian are recognized in the Great Basin and adjoining regions: metasedimentary (schist and paragneiss) and plutonic rocks older than about 1400 my; unmetamorphosed sedimentary rocks approximately comparable in age to the Belt Supergroup, 850 to 1250 my; and unmetamorphosed uppermost Precambrian sedimentary rocks probably less than about 850 my old.

A distinctive diamictite[3] unit (Crittenden and others, 1972; Stewart, 1972) at or near the base of the third (uppermost Precambrian) group of these rocks occurs at scattered localities throughout westernmost North America from Alaska to California and is the oldest unit that clearly has a depositional pattern related to the Cordilleran geosyncline. A major change in the tectonic pattern of North America appears to have taken place shortly before the deposition of the diamictite, and this change is inferred to mark the beginning of the geosyncline (Stewart, 1972). Pregeosynclinal rocks in the Great Basin and adjoining regions (fig. 4), therefore, consist of the two older groups of Precambrian rocks.

Metasedimentary and plutonic rocks older than 1400 my crop out or are known in the subsurface over a large area of Wyoming, Utah, Arizona, and California east of the Wasatch line[4] (fig. 4) but are sparsely exposed west of the Wasatch line. They are as old as 2300–2400 my (Hansen, 1965, p. 31; Armstrong and Hills, 1967, p. 1300), although most are 1500–1800 my old (King, 1969, p. 41). West of the Wasatch line, rocks clearly older than 1400 my occur in the Death Valley-Mojave Desert region of southeastern California (Wasserburg and others, 1959; Lanphere and others, 1963), in the Albion and Raft River Ranges in southern Idaho and northern Utah (locs. 1 and 2, fig. 4; Armstrong and Hills, 1967; Compton, 1972), and in the upper plate of the Willard thrust in the Wasatch Mountains (loc. 5; Crittenden, McKee, and Peterman, 1971). Rocks in the Ruby (loc. 3), East Humboldt (loc. 4), and Snake Ranges (loc. 8) in eastern Nevada commonly have been considered

[3] Diamictite: A nonsorted sedimentary rock consisting of sand and(or) larger particles in a muddy matrix (Crittenden, Schaeffer, Trimble, and Woodward, 1971, *modified from* Flint and others, 1960).

[4] The Wasatch line (Kay, 1951, p. 14) is the hinge line along the eastern border of the Cordilleran geosyncline across which the rate of westward thickening of upper Precambrian, Paleozoic, and Mesozoic strata increases greatly.

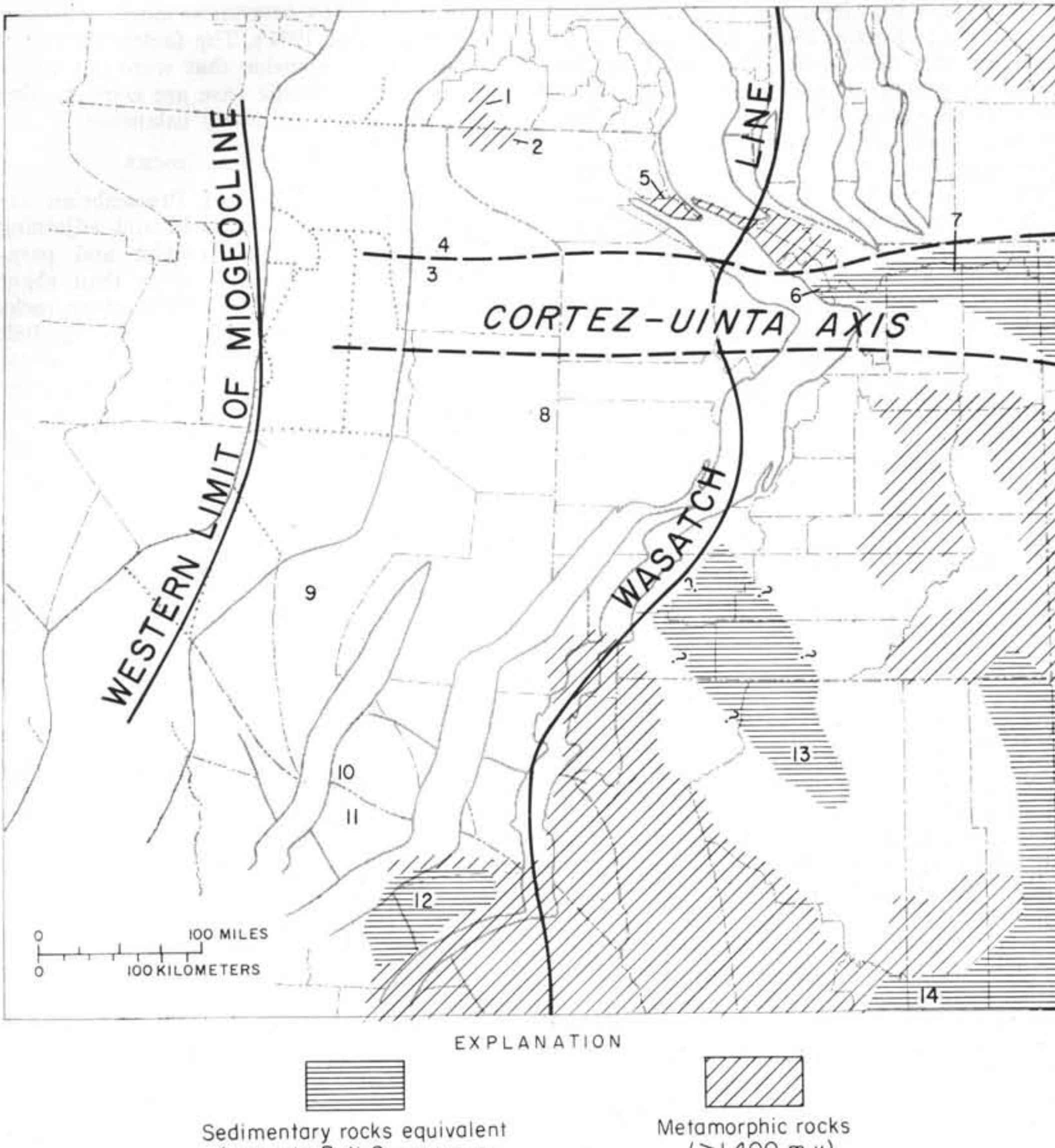


FIG. 4.—Distribution of pregeosynclinal rocks. Numbers refer to localities discussed in text.

to be Precambrian in age (Bayley and Muehlberger, 1968), but recent studies (Howard, 1971; Hose and Blake, 1970) indicate that they are probably metamorphosed Paleozoic strata and that the only Precambrian present is very young Precambrian (Howard, 1971, p. 260; Hose and Blake, 1970) that is a part of the geosynclinal sequence of this paper. Supposed older Precambrian rocks that crop out in southern Nevada (locs. 9 and 10; Ekren and others, 1971; Cornwall and Kleinhampl, 1964), may also be metamorphosed strata of latest Precambrian or Paleozoic age.

Unmetamorphosed to slightly metamorphosed pregeosynclinal sedimentary rocks equivalent in age to the Belt Supergroup are inferred to include (1) the Uinta Mountain Group (loc. 7) and the Big Cottonwood Formation (loc. 6) in northern Utah (Wallace and Crittenden, 1969; Wallace, 1972; Crittenden and others, 1952), (2) the Crystal Spring Formation and Beck Spring Dolomite in California (loc. 12, and possibly at loc. 11; Hewett, 1956; Wright and Troxel, 1967; Troxel and Wright, 1968), (3) the Grand Canyon Supergroup in northern Arizona, loc. 13; Walcott, 1894; Maxson, 1961; Ford and Breed, 1973), and (4) the Apache Group and Troy Quartzite in southern Arizona (loc. 14; Shride, 1967). More data on the correlation of these strata, including possible alternative correlations, were given recently by Crittenden and others (1972).

The Uinta Mountain Group appears to be at least 7350 m (24,000 ft) thick (Hansen, 1965, p. 33); the Big Cottonwood Formation is 4900 m (16,000 ft) thick (Crittenden and others, 1952). These rocks crop out in an east-west belt that seems to represent at least in part an original depositional trough (Wallace and Crittenden, 1969, p. 140). Deep linear troughs extending into a platform or cratonic area at high angles to the trend of a bordering geosyncline have been termed aulacogens in the USSR (Salop and Scheinmann, 1969, p. 586; Hoffman, 1971) and Paul Hoffman (oral commun. to M. D. Crittenden, Jr., 1971) has suggested that the Uinta Mountain trough may be such a feature. This aulacogen could extend west of the Wasatch and Uinta Mountains and underlie the Cordilleran geosyncline in the Great Basin. Such a postulated buried aulacogen may coincide with the Cortez-Uinta axis (fig. 4) of Roberts and others (1965, p. 1928), along which thickness trends of lower Paleozoic rocks are disrupted. Whether or not the aulacogen was connected to a geosyncline to the west is not known. C. A. Wallace (Crittenden and others, 1972, p. 337) has suggested that the Uinta Mountain Group may be correlative with post-diamictite strata of the Cordilleran geosyncline, but this correlation is not accepted here. If his correlation is correct, the Uinta Mountains aulacogen would be an offshoot of the Cordilleran geosyncline, which seems unlikely to us.

The Grand Canyon Supergroup is about 3650 m (12,000 ft) thick (Walcott, 1894) and may have been deposited in another aulacogen. The Apache Group and Troy Quartzite, on the other hand, have a combined thickness of less than about 765 m (2500 ft; Shride, 1967) and the Crystal Spring Formation and Beck Spring Dolomite have a combined thickness of 900 to 1500 m (3000 to 5000 ft; Hewett, 1956; Wright and Troxel, 1967), suggesting that these formations may have been deposited in a platform area rather than in an aulacogen.

GEOSYNCLINAL ROCKS

Geosynclinal rocks described here range in age from latest Precambrian to Late Devonian and consist of miogeoclinal carbonate and transitional assemblage rocks and of eugeoclinal siliceous assemblage rocks in the Great Basin. In the eastern and central Great Basin, miogeoclinal rocks consist of lithologically distinctive and persistent units exposed in relatively simple structural blocks where accurate measurements of thickness can be made, whereas in the western Great Basin, eugeoclinal rocks consist of lithologically monotonous units exposed in highly complex structural blocks where only crude estimates of thickness are possible. On the isopach maps, therefore, thicknesses are shown only in the eastern and central Great Basin.

A summary of the stratigraphy of the geosynclinal rocks for each of five major groups of strata is presented below. The assemblages and constituent facies described here have been summarized in table 1.

Uppermost Precambrian and Lower Cambrian Rocks

At least half of the thickness of the lower Paleozoic and uppermost Precambrian Cordilleran miogeocline is represented by uppermost Precambrian and Lower Cambrian rocks consisting of quartzite and siltstone, and, to a lesser extent, of carbonate rock and conglomerate (figs. 5–8). These strata are less than 150 m (500 ft) thick along the eastern edge of the Great Basin and thicken to over 6000 m (20,000 ft) in areas 240–320 km to the west (fig. 9). Regional descriptions of these rocks have been given by Misch and Hazzard (1962), Woodward (1963, 1965, 1967, 1968), Stewart (1970), Crittenden, Schaeffer, Trimble and Woodward (1971), and Oriel and Armstrong (1971).

Three facies of uppermost Precambrian and Lower Cambrian rocks are recognized: (1) a quartzite and siltstone facies in the eastern Great Basin; (2) a siltstone, carbonate, and quartzite facies in the central Great Basin; and (3) a siliceous and volcanic facies (one locality) in the western Great Basin (fig. 9).

The quartzite and siltstone facies typically consists of cliff-forming fine- to medium-grained quartzite in units that are 25 to 1225 m (100 to 4000 ft) thick and are separated by units of siltstone and fine- to very fine-grained quartzite from 15 to 300 m (50 to 1000 ft) thick. The quartzite ranges in composition from arkose to highly pure (97 percent SiO_2) orthoquartzite (Stewart, 1970). Cross strata of both trough and tabular planar types are common and indicate westerly transport (fig. 9; Seeland, 1968, 1969; Stewart, 1970, p. 11 and 12). Conglomerate that contains pebbles of quartz and

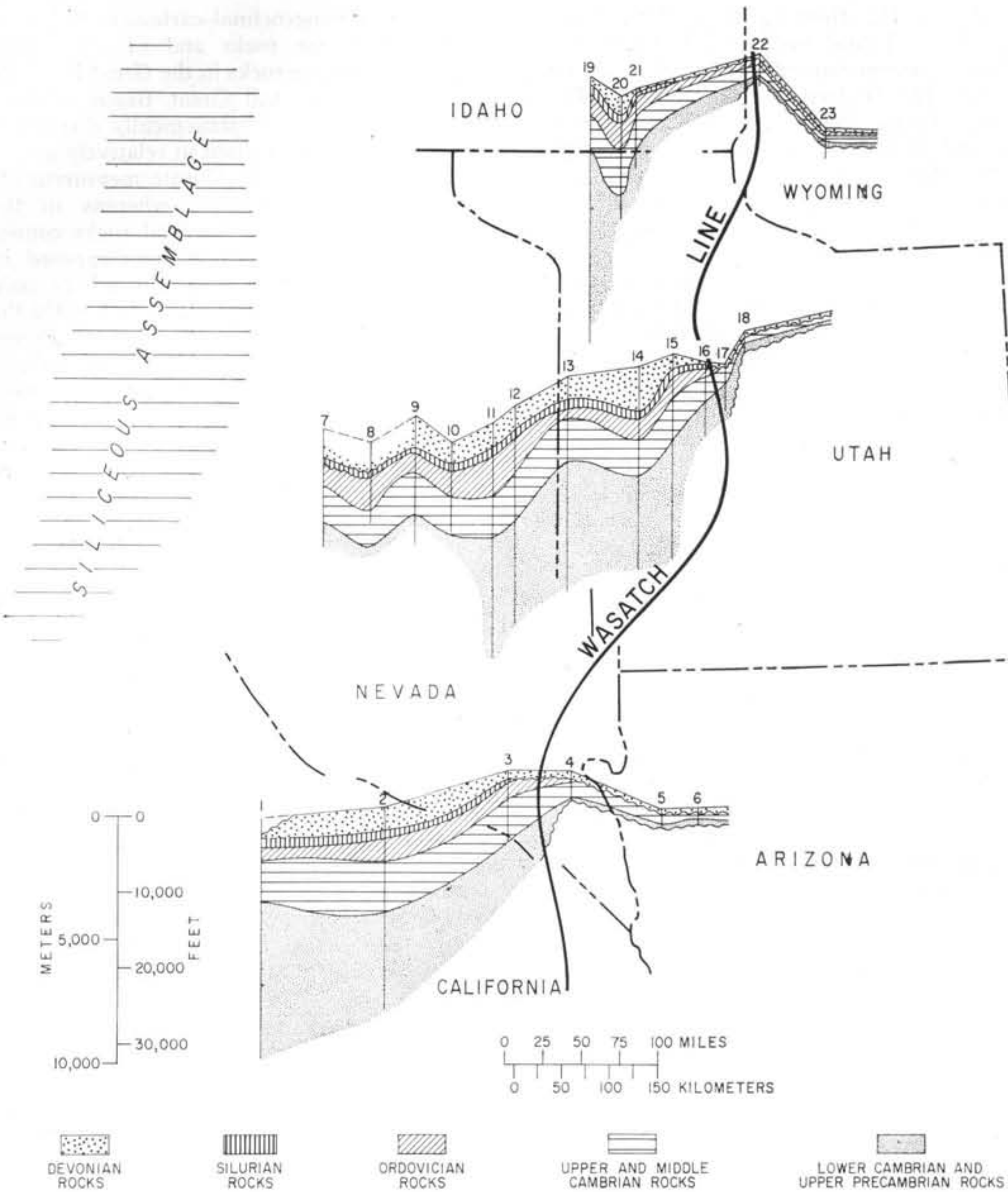


FIG. 5.—Stratigraphic diagram showing the miogeoclinal carbonate and transitional assemblages and the eugeoclinal siliceous assemblage in the Great Basin. Top of column is location of section. Column numbers are same as those on figures 6, 7, and 8.

quartzite is sparsely distributed in the quartzite. Limestone and dolomite are present in layers less than a foot to several hundred feet thick. Individual units of quartzite, siltstone, or carbonate persist for hundreds of miles along the trend of the miogeocline (Stewart, 1970, p. 64; Crittenden, Schaeffer, Trimble, and Woodward, 1971).

As noted above, a poorly sorted diamictite consisting of rounded to subangular pebbles to boulders of diverse rock types in a sandy or argillaceous matrix is recognized at or near the base of the geosynclinal sequence at several localities in Utah and California (Stewart, 1972, table 1), and appears to be of glacial origin (Troxel, 1967; Crittenden, Schaeffer, Trimble, and Woodward, 1971; Crittenden and others, 1972; Stewart, 1972).

Volcanic rocks occur in the quartzite and siltstone facies in the eastern Great Basin (Stewart, 1972, table 2) and are unknown in other Paleozoic rocks in this part of the miogeocline. The thickest unit is the Bannock Volcanic Member of the Pocatello Formation consisting of porphyritic, vesicular, or amygdaloidal volcanic flows and breccias about 300 m (1000 ft) thick (Crittenden, Schaeffer, Trimble, and Woodward, 1971) occurring low in the uppermost Precambrian and Lower Cambrian sequence in Idaho (fig. 8, column 19). Vesicular basalt flows from less than a hundred to a few hundred feet thick occur within quartzite units (Tintic Quartzite and Prospect Mountain Quartzite) high in the sequence in Utah and Nevada (Abbott, 1951; Morris and Lovering, 1961, p. 15; Kellogg, 1963, p. 687–688).

The siltstone, carbonate, and quartzite facies of the uppermost Precambrian and Lower Cambrian sequence in central Nevada and southeastern California (fig. 8) is thick and fossil-rich. It contains large amounts of siltstone (or phyllitic siltstone), thin to thick units of limestone and dolomite, and fine- to very fine-grained quartzite (Nelson, 1962, Stewart, 1970; Albers and Stewart, 1973). The fine- to medium-grained type of quartzite characteristic of uppermost Precambrian and Lower Cambrian strata in the eastern Great Basin is largely absent. Trilobites, archeocyathids, pelecypods, echinoderms, pelmatozoan debris, *Hyolithes, Salterella, Scolithus* and algae (Palmer, 1971; Stewart, 1970), are locally abundant.

The siliceous facies is represented by only one formation, the Lower or Middle Cambrian

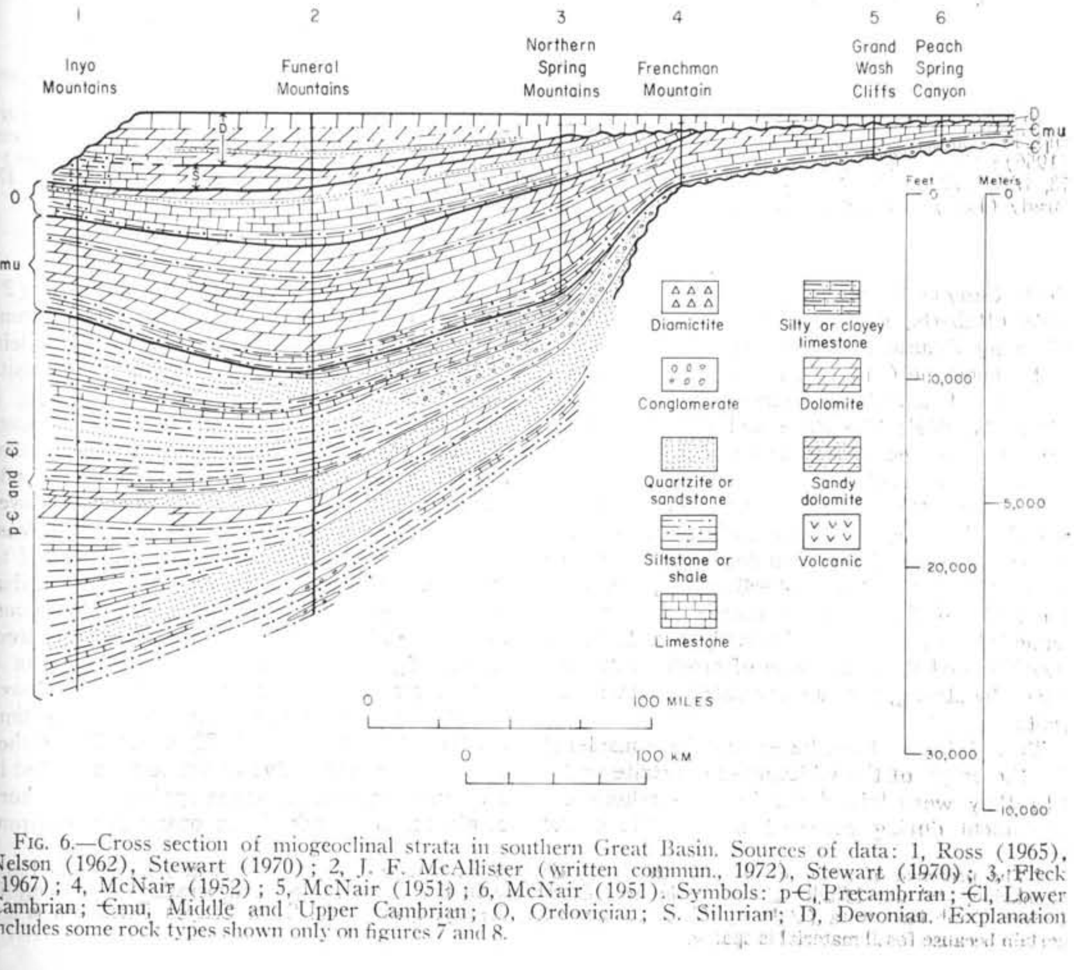


FIG. 6.—Cross section of miogeoclinal strata in southern Great Basin. Sources of data: 1, Ross (1965), Nelson (1962), Stewart (1970); 2, J. F. McAllister (written commun., 1972), Stewart (1970); 3, Fleck (1967); 4, McNair (1952); 5, McNair (1951); 6, McNair (1951). Symbols: p€, Precambrian; €l, Lower Cambrian; €mu, Middle and Upper Cambrian; O, Ordovician; S, Silurian; D, Devonian. Explanation includes some rock types shown only on figures 7 and 8.

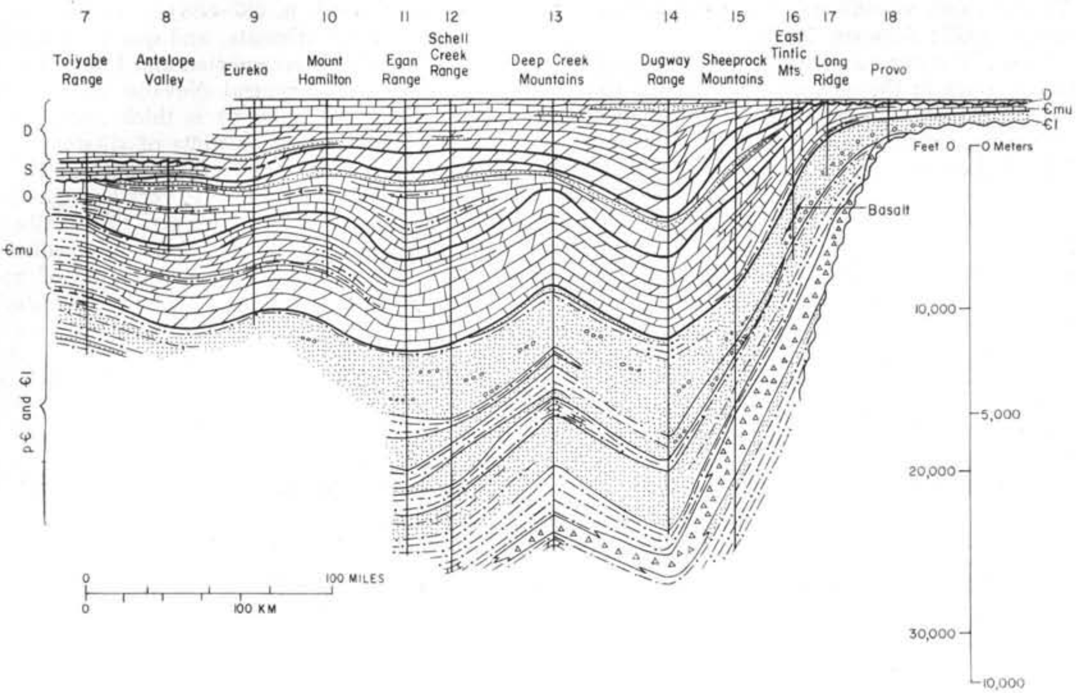


FIG. 7.—Cross section of miogeoclinal strata in central Great Basin. Sources of data: 7, J. H. Stewart and T. E. Mullens (unpub. data), Stewart and Palmer (1967); 8, Merriam (1963); 9, Nolan and others (1956); 10, Humphrey (1960); 11, Woodward (1962); 12, Young (1960) and R. K. Hose (unpub. data); 13, Bick (1966); 4, Staatz and Carr (1964); 15, Cohenour (1959); 16, Morris and Lovering (1961); 17, Brady (1965); 18, Baker (1947). Rock and letter symbols same as on figure 6.

Scott Canyon Formation[5] of north-central Nevada (Roberts, 1964; p. A14–A17). It consists of many thousands of feet of chert, argillite, and greenstone (in part pillow lavas) and minor amounts of sandstone, quartzite, and limestone (Roberts, 1964; Theodore and Roberts, 1971). Limestone lenses contain algae, sponges, archeocyathids, and trilobites.

Rocks of the quartzite and siltstone facies and the siltstone, carbonate, and quartzite facies are considered to have been deposited in shallow water (Stewart, 1970, p. 66–68; Klein, 1972), on the basis of the local abundance of such presumed shallow-water fossils as algae and archeocyathids and the abundance of cross strata produced by strong, presumably shallow-water currents.

Two different hypotheses can be considered for the origin of the widespread quartzite units: (1) they were laid down in a nearshore environment during repeated transgressions and regressions of the sea across the region; (2) they were laid down in an open-ocean environment far removed from a shoreline. Klein (1972) has suggested that they are deposits formed in a prograding tidal flat coastline, a concept that fits the first hypothesis. Stewart (1970, p. 66–68), however, has suggested that nearshore environment is difficult to reconcile with the lithologic uniformity of units over large areas, and with the absence, or at least scarcity, of deposits that can be interpreted as those of bars, beaches, inland bays, or distributary channels. The fact that current directions, as indicated by cross-strata studies, are unidirectional (fig. 9), and not bidirectional as in a tidal regime, and that the upper and lower boundaries of quartzite units are probably time conformable (Stewart, 1970, p. 64–66), rather than time transgressive as would be expected in a prograding system, argue against a nearshore origin for the sands. If an open-ocean environ-

[5] The presence of archeocyathids in the Scott Canyon Formation suggests that it may be entirely Early Cambrian, and not Middle Cambrian in age. Elsewhere in the Great Basin, archeocyathids occur only in Lower Cambrian strata (Palmer, 1971). In any case, the age span of the Scott Canyon Formation is uncertain because fossil material is sparse.

ment is a feasible alternative, strong ocean currents would be required to move the bedload sand. Perhaps storm-induced currents, such as the current velocity of 70 cm/sec registered in 80 m of water on the continental shelf of Washington (Smith and Hopkins, 1971), would be sufficient to move the sand.

The Scott Canyon Formation, by analogy with other siliceous assemblage rocks, is considered to be a relatively deep-water oceanic deposit. Fossil material such as algae and archeocyathids that suggest shallow-water conditions might occur in blocks that were transported into deep water by slumping, or they could indicate shallow-water conditions on the flanks of oceanic volcanoes.

Middle and Upper Cambrian Rocks

The maximum thickness of Middle and Upper Cambrian rocks in the Great Basin is about one-third that of uppermost Precambrian and Lower Cambrian strata. Middle and Upper Cambrian strata thicken (fig. 10) from 300–600 m (1000–2000 ft) in the eastern part of the

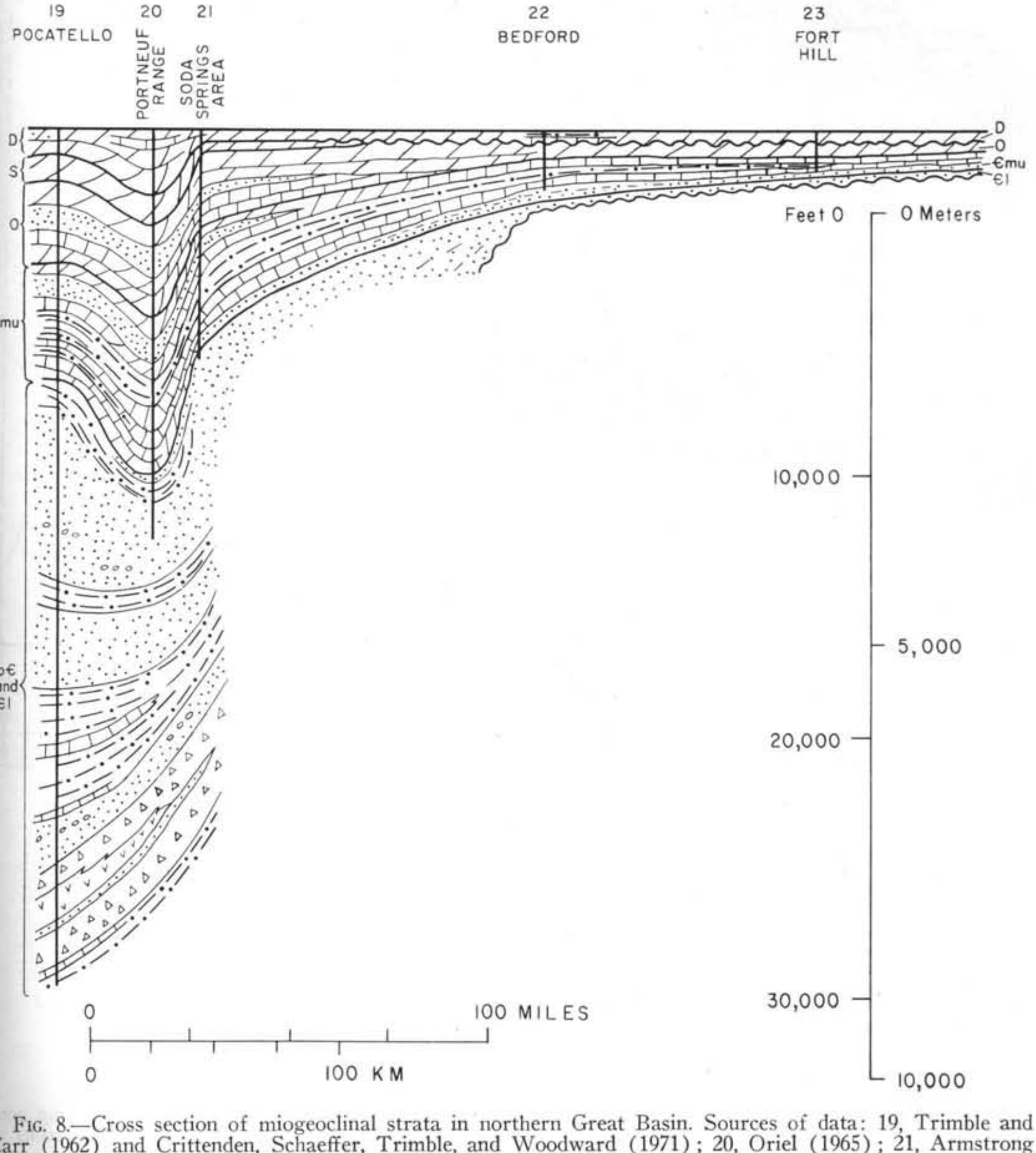


FIG. 8.—Cross section of miogeoclinal strata in northern Great Basin. Sources of data: 19, Trimble and Carr (1962) and Crittenden, Schaeffer, Trimble, and Woodward (1971); 20, Oriel (1965); 21, Armstrong (1969); 22, Rubey (1958); 23, Oriel (1969). Rock and letter symbols same as on figure 6.

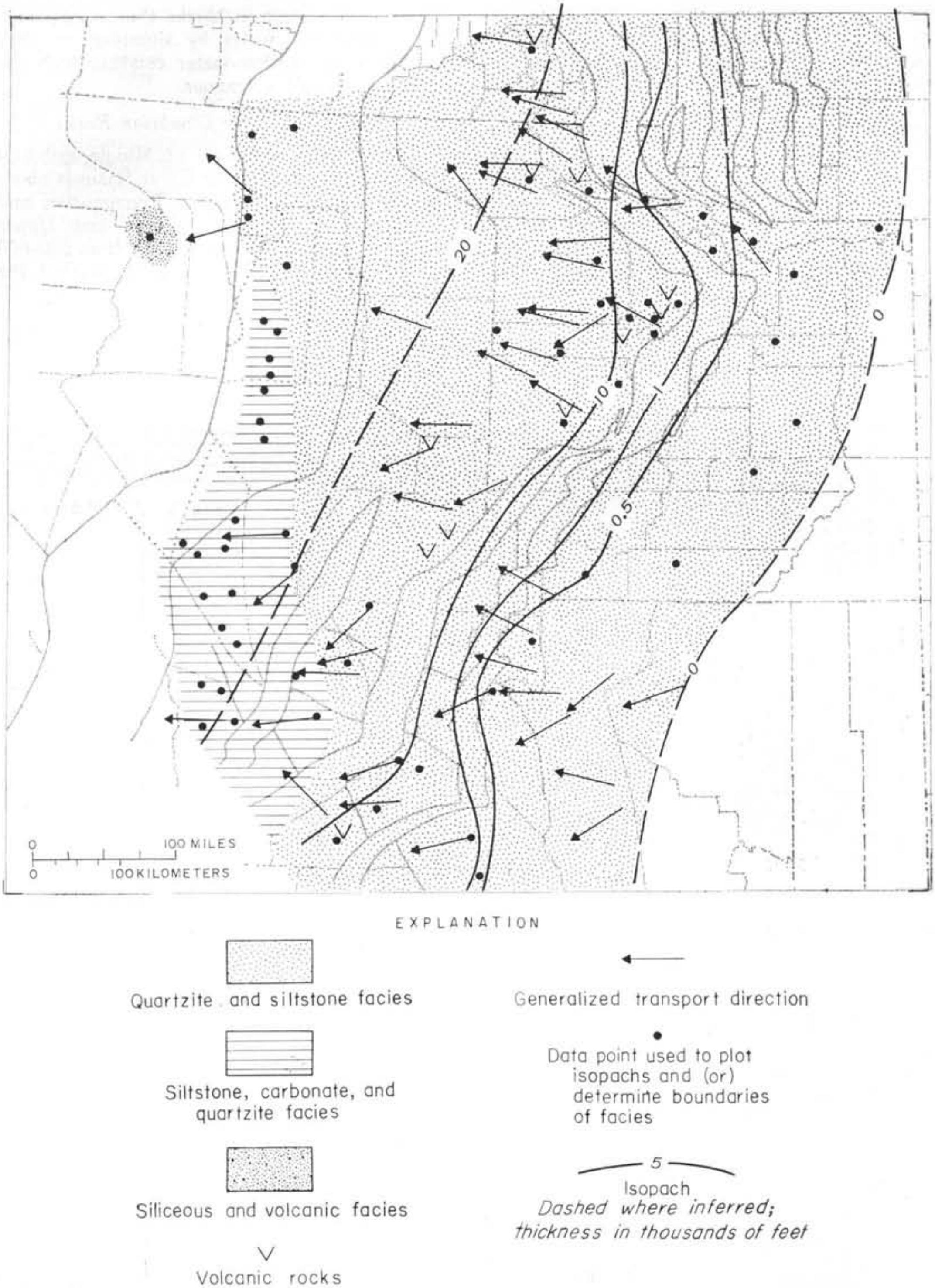


FIG. 9.—Isopach and facies map of uppermost Precambrian and Lower Cambrian strata showing generalized current directions (after Seeland, 1968, 1969; Stewart, 1970) and distribution of volcanic rocks in miogeoclinal sequence.

Great Basin to more than 2500 m (8000 ft) in the "Ibex basin"—a basin in western Utah and eastern Nevada (Webb, 1958; Hintze, 1959)—and to more than 1750 m (6000 ft) in the "northern Utah basin"—a basin in northwestern Utah (Webb, 1958). During much of Paleozoic time, these two basins were separated by the Tooele arch (Hintze, 1954, 1959; Webb, 1958),

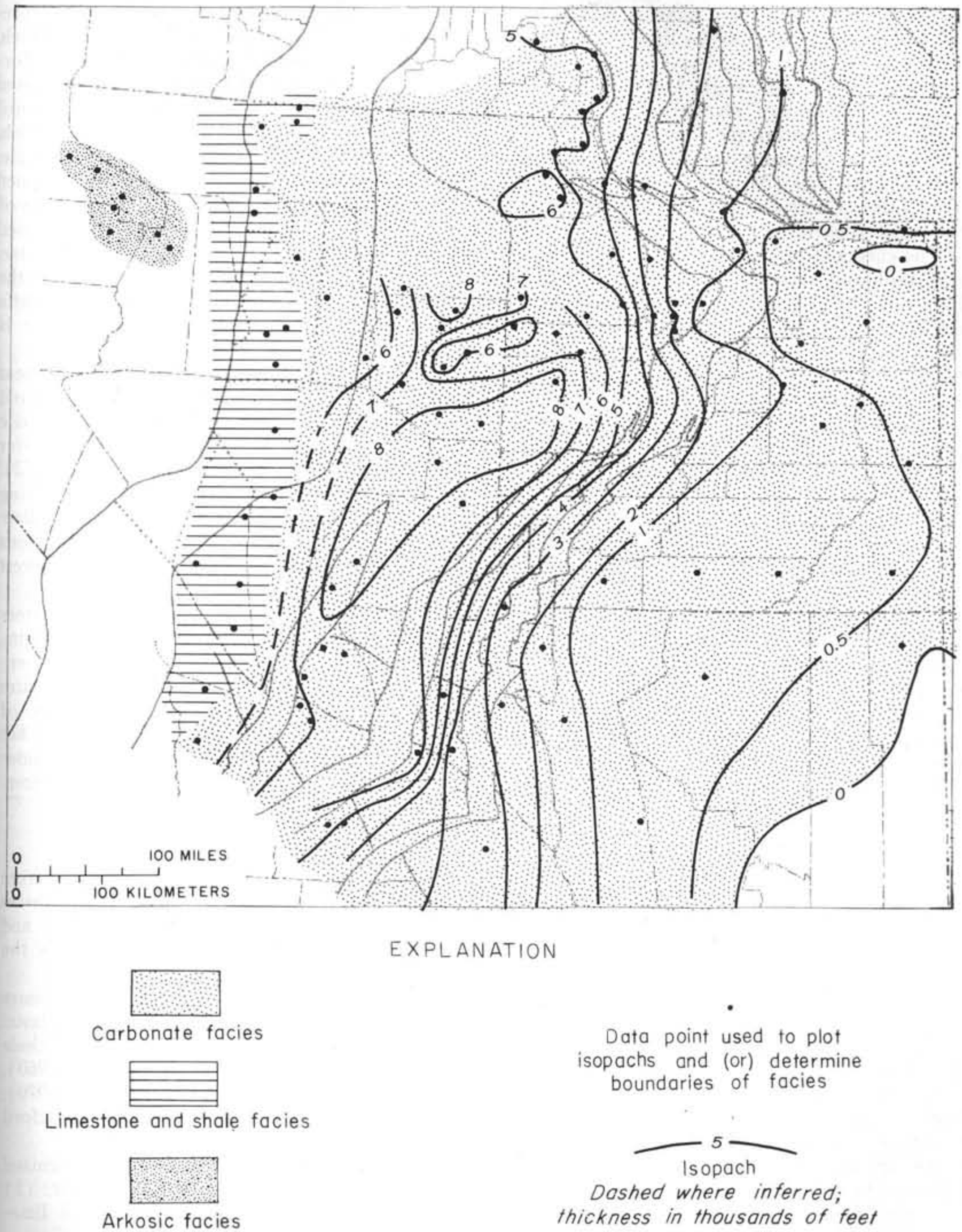


FIG. 10.—Isopach and facies map of Middle and Upper Cambrian strata (partly restored).

the Cortez-Uinta axis of Roberts and others (1965), but the Tooele arch was not well defined during Middle and Late Cambrian time.

An excellent summary of Cambrian stratigraphy in the Great Basin is given by Palmer (1971). Other important papers include those of Robison (1960, 1964), Rigo (1968), and Kepper (1972).

Three major facies of Middle and Upper Cambrian rocks are recognized in the Great Basin: (1) a carbonate facies in the eastern part, (2) a limestone and shale facies in the central part, and (3) an arkosic facies (Paradise Valley Chert and Harmony Formation) in the western part.

The carbonate facies consists typically of several thousand feet of massive- to well-bedded, light- to dark-gray dolomite and limestone. In detail, the stratigraphy is moderately complex; shale, laminated limestone, chert, and intraformational conglomerate are major units locally; some are regionally persistent. An unusual quartzite unit, the Worm Creek Quartzite Member of the St. Charles Limestone (Rigo, 1968; Oriel, 1965), occurs in the carbonate facies in southern Idaho. The lower few tens or hundreds of meters of the carbonate facies are shale-rich and gradational downward into the quartzite and siltstone of the Lower Cambrian.

Rocks of the limestone and shale facies consist of light- to dark-gray, evenly laminated to very thin-bedded limestone and dark-gray to olive-gray, extremely fine-textured shale. The proportion of limestone to shale in this facies is variable; some sections are predominantly limestone, others predominantly shale. The Emigrant Formation, a formation of this facies in western Nevada (Albers and Stewart, 1973), contains abundant beds of chert in addition to limestone and shale. Other strata in the limestone and shale facies include the Swarbrick, Tybo, and Hales formations in the Tybo area, Nye County, Nevada (Ferguson, 1933), the Crane Canyon sequence of Means (1962), part of the Broad Canyon sequence or Formation of Means (1962) and Washburn (1970), an unnamed sequence in the Mount Callaghan area, Lander County, Nevada (Stewart and Palmer, 1967), the Preble Formation (Hotz and Willden, 1964; Gilluly, 1967), and part of the Tennessee Mountain Formation of Bushnell (1967). Rocks lithologically similar to those of the limestone and shale facies also occur locally within the carbonate facies, and the boundary between the limestone and shale facies and the carbonate facies is irregular and difficult to define. The only volcanic rocks reported in the Middle and Upper Cambrian sequence anywhere in the Great Basin are in the Shwin Formation (Gilluly and Gates, 1965), a unit mainly of limestone and shale, which may represent that facies.

Strata of the arkosic facies consist of the Paradise Valley Chert (Hotz and Willden, 1964), an Upper Cambrian unit composed predominantly of chert and at least 100 m thick, and the depositionally overlying Harmony Formation, a late Late Cambrian unit composed predominantly of medium- to coarse-grained arkosic sandstone, perhaps 1000 to 1500 m thick. The Harmony contains minor amounts of shale, sparse limestone, and rare coarse-grained "gritty" beds containing granules and pebbles of quartz and feldspar. The sandstone is composed of quartz, orthoclase, microcline, plagioclase, and mica. Zircon from a sandstone bed in the Harmony has been dated as 958 my old (Jaffe and others, 1959, p. 130). Graded bedding is characteristic of the Harmony.

Carbonate facies rocks of the eastern Great Basin formed in shallow inner shelf lagoons and on shoals, whereas the limestone and shale facies formed in open deeper water of outer shelf environments to the west (Kepper, 1972). The shoals, which consisted of an interlacing pattern of tidal algal mudbanks and shallow subtidal basins, at times separated shelf lagoons on the east from the open water on the west (Kepper, 1972).

The graded bedding in the Harmony Formation suggests deposition by turbidity currents, probably in a deep-water continental rise environment. The presence of microline indicates a plutonic source rock, and the isotopic age of the zircons indicates a Precambrian terrane, but the source area's location (as discussed under "Sedimentary and tectonic models") is uncertain.

Ordovician System

Strata of the Ordovician System thicken from 0 near the eastern edge of the Great Basin to more than 1500 m (5000 ft) in the Ibex and northern Utah basins and are thin across the Tooele arch (fig. 11).

Ordovician strata have been studied in more detail than any other system in the Great Basin. Important regional studies or summaries include those of Webb (1958), Lowell (1958, 1960), Hintze (1959, 1963), Ross (1964a and b, 1970), Ross and Berry (1963), Kay and Crawford (1964) and Ketner (1966, 1968, 1969).

Four facies of Ordovician strata recognized in the Great Basin are, from east to west: (1) carbonate and quartzite; (2) shale and limestone; (3) shale and chert; and (4) siliceous and volcanic.

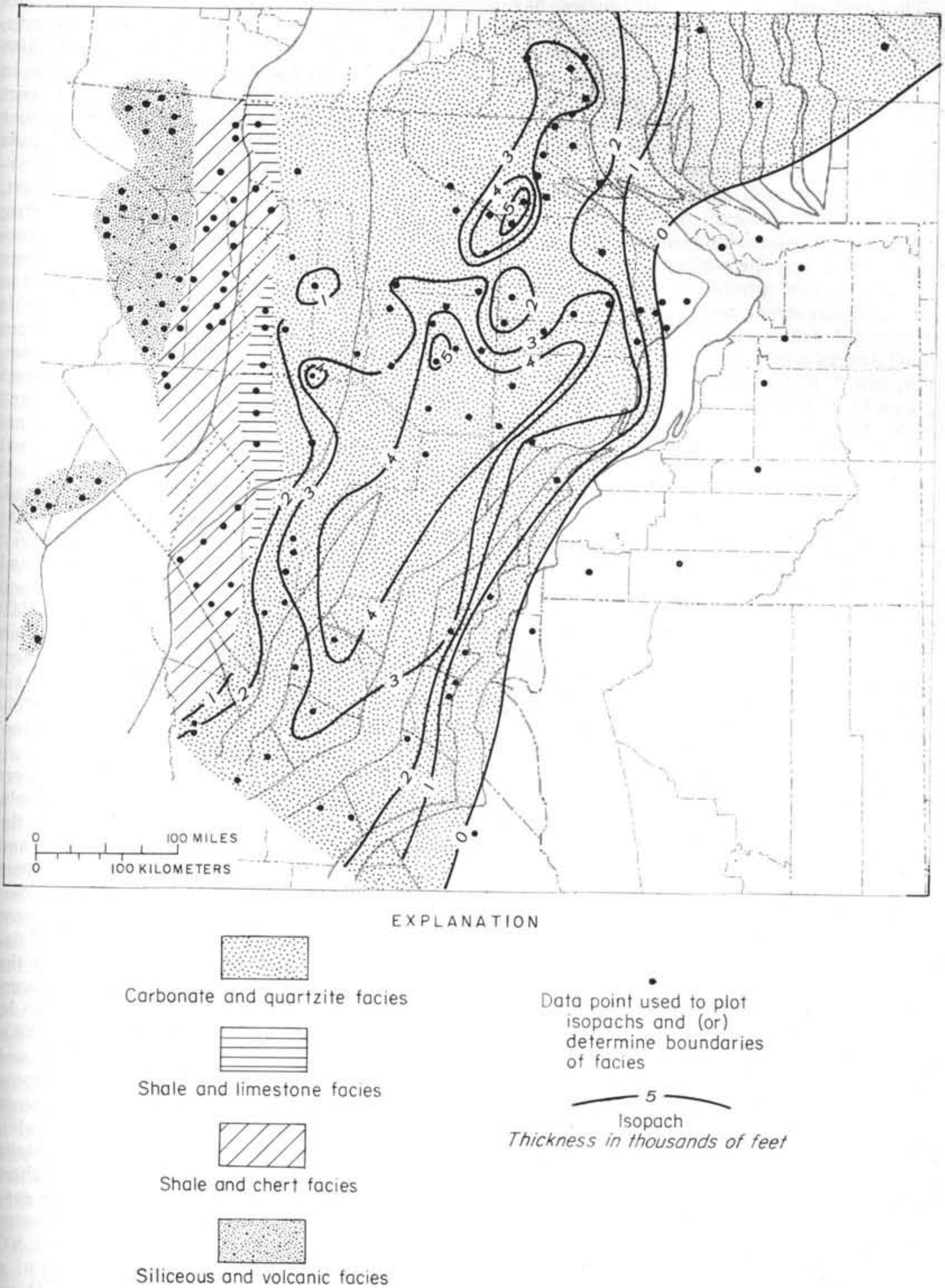


FIG. 11.—Isopach and facies map of Ordovician strata.

The carbonate and quartzite facies has a typical threefold division: the Pogonip Group or Garden City Formation below (limestone with minor shale), the Eureka Quartzite and(or) Swan Peak Quartzite in the middle, and the Fish Haven Dolomite and equivalent units above. The Eureka and Swan Peak quartzites are distinctive and widespread; they are absent over the Tooele arch (Webb, 1958) where Upper Ordovician dolomite overlies various parts of the Lower Ordovician sequence. The Eureka wedges out along the western or outer edge of the miogeocline where Ordovician, Silurian, and Devonian strata are characterized by local or regional facies changes as well as by internal disconformities or unconformities (Webb, 1958, figs. 7, 8, 11, and 13; Lowell, 1960, fig. 2; Kay and Crawford, 1964, fig. 4; Merriam, 1963, fig. 7; McKee and Ross, 1969, fig. 5; Ketner, 1968, fig. 3; Ross, 1966, p. 24 and pl. 3).

The shale and limestone facies is relatively limited in distribution and consists of dark shale and siltstone and well-bedded argillaceous limestone. It contains both graptolite and shelly faunas. This facies includes the Aura Formation of Decker (1962) in northwest Elko County, Nevada, the Perkins Canyon Formation of Kay (1960), the Zanzibar Limestone and Toquima Formation (Ferguson, 1924), all in northern Nye County, Nevada, and some related rocks (Lowell, 1958; 1960) in southern Lander County, Nevada.

The shale and chert facies is widespread and thick. It consists of dark-gray or commonly varicolored graptolite-rich shale, several thick units of thin-bedded chert, minor amounts of limestone, quartzite, and volcanic rock, and very minor amounts of bedded barite. Rocks of this facies occur in both the lower and upper plates of the Roberts Mountains thrust. Rocks in the lower plate include the Comus Formation (Hotz and Willden, 1964) and most of the Palmetto Formation (Albers and Stewart, 1973). Rocks in the upper plate include the Vinini Formation (Merriam and Anderson, 1942; Gilluly and Mazursky, 1965), the Clipper Canyon sequence of Kay and Crawford (1964), the Valder Formation and Agort Chert of Riva (1970), and the Basco Formation of Lovejoy (1959). Volcanic rocks, mostly mafic lava flows that include pillow lavas, are sparse in this facies (Merriam and Anderson, 1942; Kay and Crawford, 1964, fig. 2; Riva, 1970). Radiolaria have been reported in some beds (Merriam and Anderson, 1942; Kay and Crawford, 1964, p. 437, Riva, 1970, p. 2692 and 2697; Ketner, 1969).

Rocks of the siliceous and volcanic facies are mainly assigned to the Valmy Formation, a widespread formation in north-central Nevada (Gilluly and Gates, 1965, p. 23–34; Roberts, 1964, p. 17–22; Churkin and Kay, 1967). Also included in this facies are Ordovician strata in southern Mineral County and the northern part of Esmeralda County, Nevada and in the Mount Morrison roof pendant in the Sierra Nevada (Rinehart and Ross, 1964). The Valmy Formation consists of thousands of meters of chert, quartzite, shale, siltstone, greenstone (including pillow lava), sandstone, and very minor limestone and bedded barite. It contains much more greenstone and quartzite, and less shale, than the shale and chert facies. The quartzite is highly vitreous and quartz-rich (98 percent SiO_2; Roberts, 1964, p. A19; Gilluly and Gates, 1965, table 2; Ketner, 1966, table 1), and forms units that are commonly fault-bounded and roughly 100 m thick. The quartzite in the Valmy is distinctly different from that in miogeoclinal rocks (Eureka and Swan Peak Quartzites); it is darker gray and coarser, not so well sorted, and typically has a conspicuous seriate texture ranging from fine to coarse (Ketner, 1966). The Valmy Formation is highly faulted and thickness estimates are uncertain. Gilluly and Gates (1965, p. 23) suggest a thickness of 6000 to 7500 m (20,000–25,000 ft), although Churkin (1973) indicates that it is much thinner, generally only perhaps 1000 m thick. A few radiolaria have been reported in some chert beds (Gilluly and Masursky, 1965, p. 42; Ketner, 1969; Rinehart and Ross, 1964, pl. 2).

Ordovician carbonate rocks probably were deposited in shallow shelf lagoons and on shoals. Conspicuous bioherms occur locally near the western limit of carbonate facies (Ross and Cornwall, 1961). The dolomite in the upper part of the Ordovician System may be mainly dolomitized limestone related to the development of magnesium-rich water in shallow shelf lagoons and local supratidal areas. Although the Eureka and Swan Peak Quartzites are of uncertain origin, they are generally considered to be shallow-water marine deposits (Ketner, 1968).

Rocks of the shale and limestone, shale and chert, and siliceous and volcanic facies appear to have been deposited in progressively deeper water from east to west. The presence of radiolarian chert, volcanic rocks, and bedded barite (Poole and others, 1968) in the shale and chert facies and the siliceous and volcanic facies supports an interpretation of a relatively deepwater oceanic environment. Depths greater than 500 m (1500 ft) are suggested by the results of a preliminary study made by Churkin (1974) on the vesicularity of Ordovician pillow basalts as compared with modern pillow lavas. The silicarich quartzite, abundant in the siliceous and volcanic facies, is anomalous for such an environment, and the origin of these rocks is not clearly understood. As discussed under "Sedimentary and tectonic models," the source for the sand making up these quartzites is not clear.

Silurian System

The Silurian System is the most limited areally and the thinnest of any lower Paleozoic system in the Great Basin (fig. 12). Silurian strata thicken westward from an erosional edge in the eastern Great Basin to more than 300 m (1000 ft) in the Ibex and northern Utah basins and to at least 600 m (2000 ft) along the western edge of the miogeocline in central Nevada. Regional studies of Silurian rocks of the Great Basin include those of Winterer and Murphy (1960), Merriam (1963), and Berry and Boucot (1970). T. E. Mullens of the U.S. Geological Survey has recently completed a comprehensive study of the Silurian and Devonian Roberts Mountains Formation of central Nevada. Poole is completing a regional study of the dolomite facies in the eastern and southern Great Basin.

Four facies of Silurian rocks recognized in the Great Basin are, from east to west: (1) dolomite, (2) laminated limestone, (3) chert and shale, and (4) feldspathic sandstone.

The dolomite facies (Laketown Dolomite and equivalent strata) consists almost entirely of gray thin- to thick-bedded dolomite but contains some dark-gray, locally cherty units. Westward, in the central Great Basin, the dolomite changes rather abruptly into platy weathering laminated limestone (the laminated limestone facies) that is black to very dark gray on fresh surfaces and contains abundant detrital silt of quartz and feldspar, common carbonaceous material and pyrite, and locally abundant graptolites. The laminated limestone facies includes part of the Roberts Mountains Formation (Merriam, 1940; Winterer and Murphy, 1960) in north-central Nevada, the Chellis and Storff Formations of Decker (1962) in northwest Elko County, Nevada, the Noh Formation of Riva (1970) in the upper plate of the Roberts Mountains allochthon in northeast Elko County, the Roberts Mountains Formation of southern Nevada (Cornwall and Kleinhampl, 1964), and the basal part of the Sunday Canyon Formation in California (Ross, 1966).

Rocks of the chert and shale facies include the Fourmile Canyon Formation (Gilluly and Masursky, 1965) and unnamed Silurian rocks mapped by Lovejoy (1959), Kerr (1962), and Gardner and Peterson (1968), Evans (1972), and Evans and Cress (1972). These strata consist of chert, argillite, shale, and siltstone; the Fourmile Canyon Formation contains a few thin beds of fine-grained sandstone. Dark volcanic flow rocks with pillow structures have been identified by Kerr (1962, p. 449) in rocks that he assigned to the Silurian, but these rocks may be misidentified Ordovician units, as no other volcanic flow rocks have been identified in the Silurian of the Great Basin. Although thicknesses are uncertain, the Fourmile Canyon could be 1200–1800 m (4000–6000 ft), or even thicker (Gilluly and Masursky, 1965).

The feldspathic sandstone facies consists of the Elder Sandstone. The Elder is an unusual siliceous assemblage formation composed predominantly of light-colored, fine-grained, commonly silty sandstone containing quartz (70–80 percent), potassium feldspar (15–25 percent), muscovite (5 percent), and a little albite (Gilluly and Gates, 1965, p. 35). Siltstone, shale, and chert are present in minor amounts. Some siltstone beds contain ghosts of pumice shards and some sandstone beds contain grains that appear to be devitrified volcanic glass (Gilluly and Gates, 1965, p. 35–36; Gilluly and Masursky, 1965, p. 58). The thickness of the Elder is at least 600 m (2000 ft) and probably 1200 m (4000 ft) (Gilluly and Gates, 1965, p. 36).

Silurian dolomite may have originally been limestone formed on a broad shallow-water shelf containing many shoals and lagoons, and dolomitization may have been a secondary process related to the development of magnesium-rich waters in lagoons and on shoals. Possible reefs have been identified (Winterer and Murphy, 1960) along the western edge of the dolomite facies in north-central Nevada; the fossil-rich Vaughn Gulch Limestone (Ross, 1966, p. 30) of southeastern California could represent a similar, though undolomitized, reef complex. The laminated limestone probably represents moderately deep-water deposition on the outer shelf. Rocks of the chert and shale facies may have been deposited in relatively deep water, perhaps on the continental rise, although Gilluly and Masursky (1965, p. 55) have described abundant current bedding, which they believe indicates no unusual depth. The depositional environment of the Elder is uncertain, although the presence of algal fragments (Gilluly and Gates, 1965, p. 36), if not of detrital origin, would indicate at least local shallow-water deposition, whereas the presence of chert suggests deeper water deposition. The occurrence of tuffaceous beds in the Elder suggests that much of the feldspar in the formation was volcanically derived.

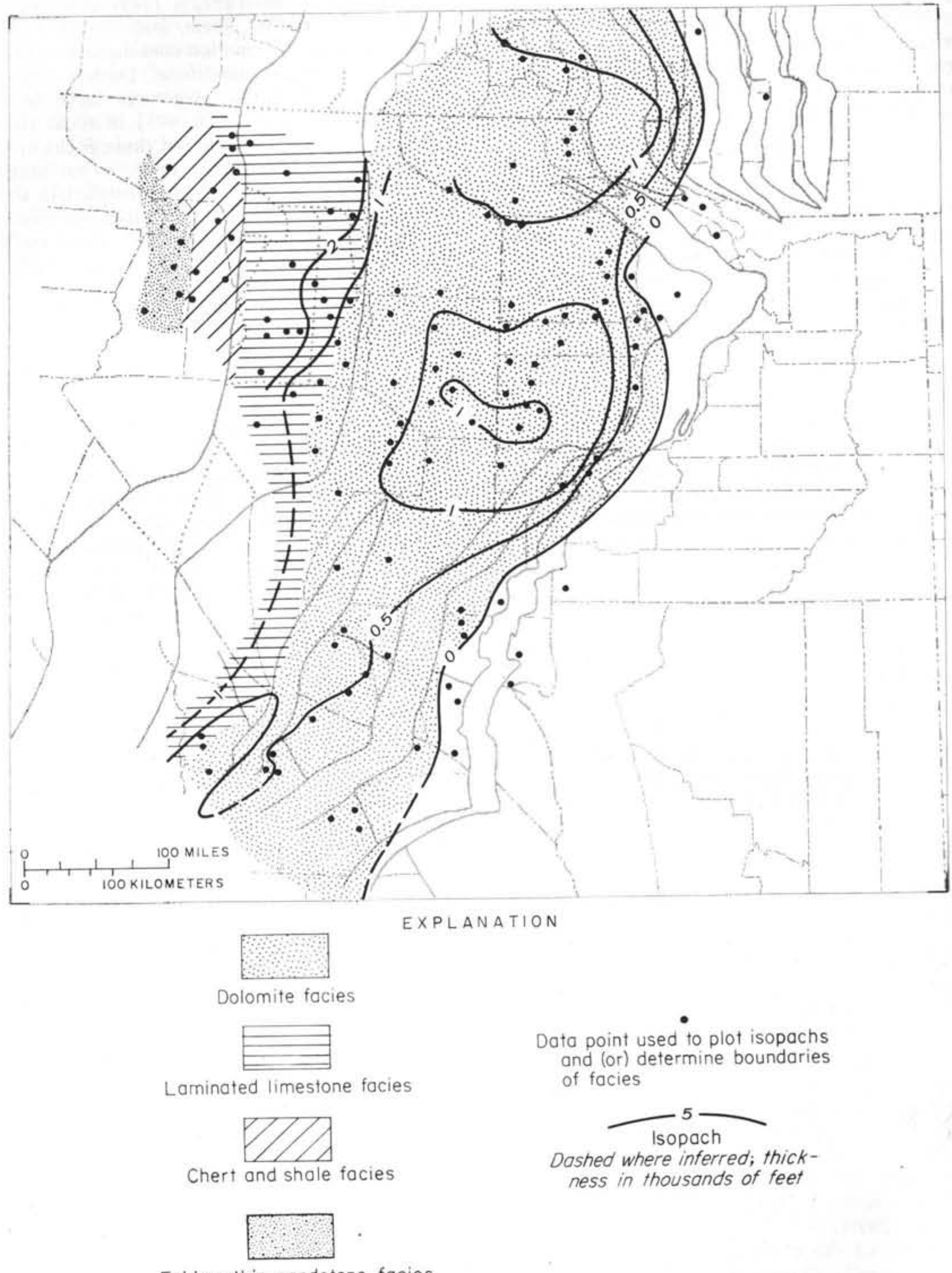


FIG. 12.—Isopach and facies map of Silurian strata.

Devonian System

Devonian rocks thicken from about 300 m (1000 ft) in the eastern Great Basin to more than 1800 m (6000 ft) in the Ibex basin, more than 1200 m (4000 ft) in the northern Utah basin, and locally more than 1500 m (5000 ft) in central Nevada (fig. 13). Uppermost Devonian strata in the eastern part of the Great Basin were deposited synchronously with the initial development of the Antler orogeny in the central part of the Great Basin (Poole, this volume). Some Devonian thickness trends may be slightly affected by the initial movement of the Antler orogeny during the Devonian, but the isopach map has been modified to restore any changes caused by post-Devonian erosion. The most comprehensive summary of Devonian stratigraphy in the Great Basin is that by Poole and others (1967). Other important regional studies include those of Osmond (1954, 1962), Merriam (1940, 1963), Carlisle and others (1957), Langenheim and others (1960), and Johnson (1965, 1971).

Four facies of Devonian strata recognized in the Great Basin are, from east to west: (1) carbonate and quartzite, (2) limestone and shale, (3) shale and chert, and (4) chert.

The carbonate and quartzite facies consists of cliff-forming thin-to thick-bedded limestone and dolomite, some of which contain sandy or silty units, and interbedded sandstone or quartzite. In north-central and northern Nevada, thin- to thick-bedded limestone and dolomite of the carbonate facies grade westward into units such as the Wenban Limestone (Gilluly and Masursky, 1965), Rabbit Hill Limestone (Merriam, 1963), and Van Duzer Limestone (Decker, 1962) of laminated limestone, silty limestone, calcareous shale, and clastic limestone. These formations and others are included in the limestone and shale facies. Farther east in Nevada (Elko and northern Eureka Counties), somewhat different Devonian strata are included with the limestone and shale facies. These include unnamed rocks composed of limestone, shale, and locally chert in (1) the Windermere Hills (Oversby, 1972), (2) the Snake Mountains (Gardner and Peterson, 1968), (3) the Pinon Range (Smith and Ketner, 1968, p. I1–I7), and (4) near Marys Mountain (Evans, 1972; Evans and Cress, 1972). Also included is the Coal Creek sequence of Lovejoy (1959) in western Elko County. This sequence consists of a lower plate of limestone, argillaceous limestone, and greenstone—the only Devonian volcanic rocks that have been reported in the Great Basin—and an upper plate of limestone, quartz-sandy limestone, siltstone, shale, and chert. In places the transitional Devonian strata in Elko and northern Eureka Counties occur as thrust slices within the upper plate of the Roberts Mountains thrust; in other places, they may be para-autochthonous. Their reconstructed position on the palinspastic base is not everywhere certain. Transitional Devonian rocks, composed of limestone and shale, also occur in California, where they constitute the upper part of the Sunday Canyon Formation (Ross, 1966).

The shale and chert facies includes the Woodruff Formation (Smith and Ketner, 1968), mostly dark-gray to black siliceous mudstone and radiolarian chert; the Cockalorum Wash Formation (Merriam, 1973), dark-gray mudstone and sparse coralline limestone; and siliceous mudstone, radiolarian chert, and sparse coralline limestone near Warm Springs in central Nye County, Nevada (Kleinhampl and Ziony, 1967).

The chert facies consists of the Slaven Chert (Gilluly and Gates, 1965; Gilluly and Masursky, 1965; Stewart and Palmer, 1967), composed predominantly of thin- to thick-bedded black chert, with very minor amounts of sandstone, shale, feldspathic siltstone, bedded barite, and limestone. Radiolaria occur in the chert (Gilluly and Masursky, 1965, p. 42). Thickness of the Slaven is uncertain but Gilluly and Gates (1965, p. 36) indicate that 1200 m (4000 ft) is probably not an excessive figure.

Rocks of the Devonian carbonate facies probably consist of shallow-water, subtidal, intertidal, and supratidal deposits formed on a broad inner shelf. Sand probably was derived from erosion of Ordovician sandstone on the shelf or from basal Cambrian and Precambrian sandstone on the craton (Osmond, 1962). Rocks of the limestone and shale facies are somewhat similar to those of the laminated limestone facies of the Silurian and probably were deposited in moderately deeper water, near the outer edge of the shelf. The radiolarian chert and the bedded barite of the shale and chert and the chert facies suggests deep water, probably oceanic conditions.

SEDIMENTARY AND TECTONIC MODELS

Two models of the Cordilleran geosyncline have been proposed (fig. 14). In one, deposition is visualized as occurring in a marginal sea bounded on the west by an island arc system (Eardley, 1947; Kay, 1951; Roberts, 1968a; Burchfiel and Davis, 1972. fig. 2; Moores, 1970; Churkin, 1974). In the other, deposition is visualized as taking place along a stable continental margin (Stewart, 1972; Burchfiel and Davis, 1972, fig. 3). Each model has its own particular

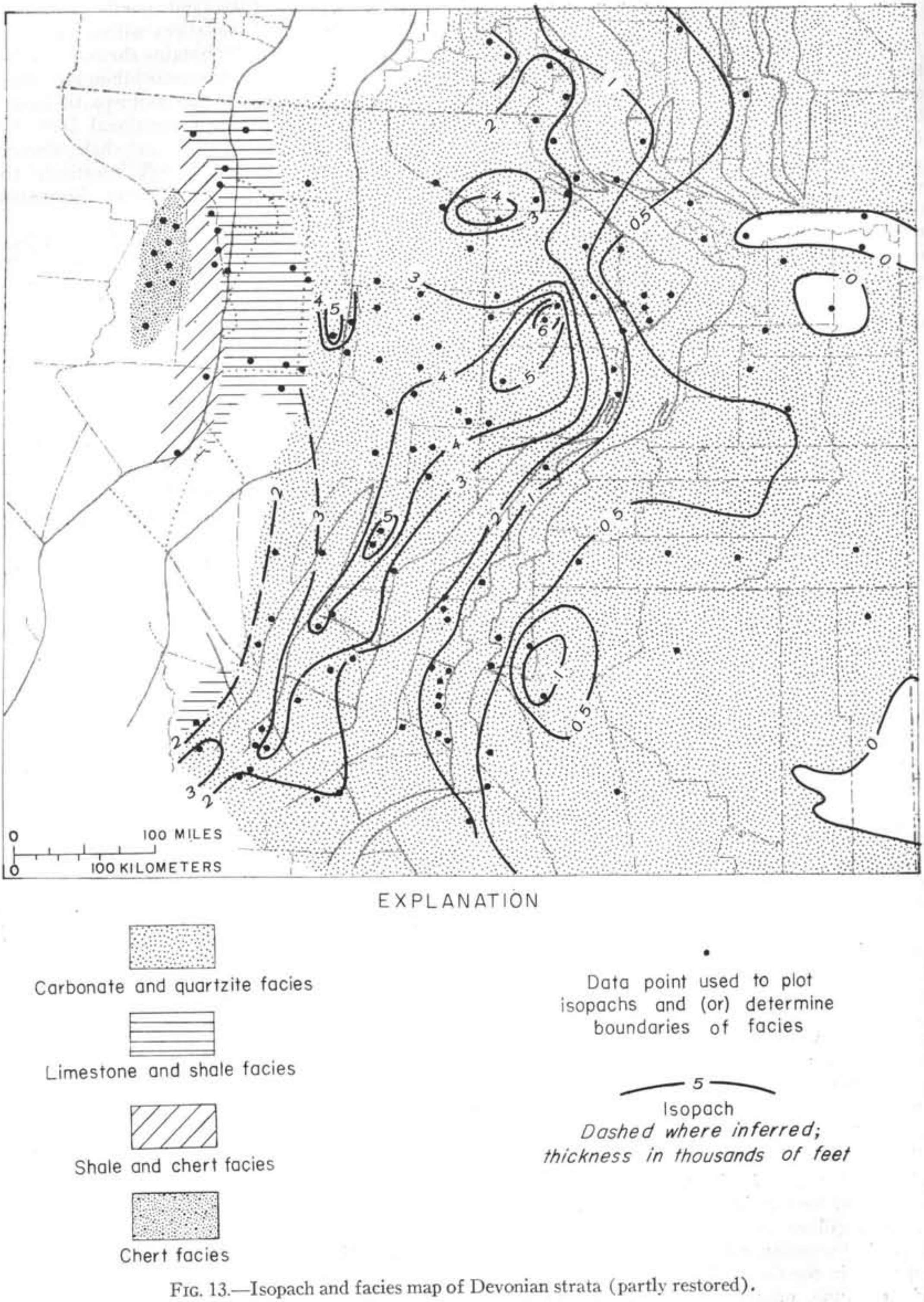


FIG. 13.—Isopach and facies map of Devonian strata (partly restored).

appealing attributes, as described below, and neither is necessarily unrelated to the other—theoretically a marginal sea-island arc system could change to a stable continental margin by the dying out of the subduction zone below the arc-trench system, or a stable continental margin could change to a marginal sea-island arc system by the development of a subduction zone.

In the marginal sea model, which has been described in detail by Churkin (1974), the miogeocline is along the inner side of the marginal basin at the edge of the main continental mass. Siliceous assemblage rocks are deep-water deposits within the basin; basalts formed by eruptions on the ocean floor. Small bodies of alpine-type serpentinite (Poole and Desborough, 1973) tectonically interleaved with strongly deformed allochthonous upper and lower Paleozoic oceanic rocks in western Nevada may in part represent lower Paleozoic mantle beneath the marginal sea. In this model, rocks farther west, including the lower Paleozoic rocks of the Klamath Mountains in California (Irwin, 1966), are inferred to be island-arc assemblages. Possible interpretations within the framework of this model are that a basement of continental crust in the island arc system could have supplied the plutonic rock debris in the Harmony Formation of north-central Nevada; that associated supracrustal sandstone could be a source of the mature sand-forming quartzite of the Ordovician Valmy Formation; and that island arc volcanic rocks could be a source for the probable abundant volcanic debris in the Silurian Elder Sandstone. A western source for the Valmy sand has been suggested by Ketner (1966) because quartzite of the Valmy is coarser and less well sorted than Ordovician sand on the shelf. A further indication of a western source is the greater abundance of quartzite in the outer belt of the siliceous assemblage rocks in the Great Basin than in the inner belt. Hopson (1973) has indicated, however, that Ordovician to Permian rocks in the Klamath Mountains rest on Ordovician oceanic crust, and Churkin (1974) also indicates that the island arc assemblages did not develop on continental crust. Conceivably, continental crust could have been present elsewhere in the island arc system and now be buried under younger rocks, or such continental crust could have been subsequently rifted away.

An alternative interpretation within the framework of the marginal sea model is that the Harmony and Valmy detritus had a source on the North American continent. Churkin (1974) suggests that the sand of the Valmy and sand of comparable age on the shelf in central Idaho may have had a common source—on the craton—and that the Valmy sand was carried into deeper parts of the basin and transported southward into the Great Basin, although he does not explain why basin sands are coarser than shelf sands. Another possibility is that the sand was derived from slumping or erosion of older (upper Precambrian and Lower Cambrian) mature sand deposits along the steep westward front (continental slope) of the mio-

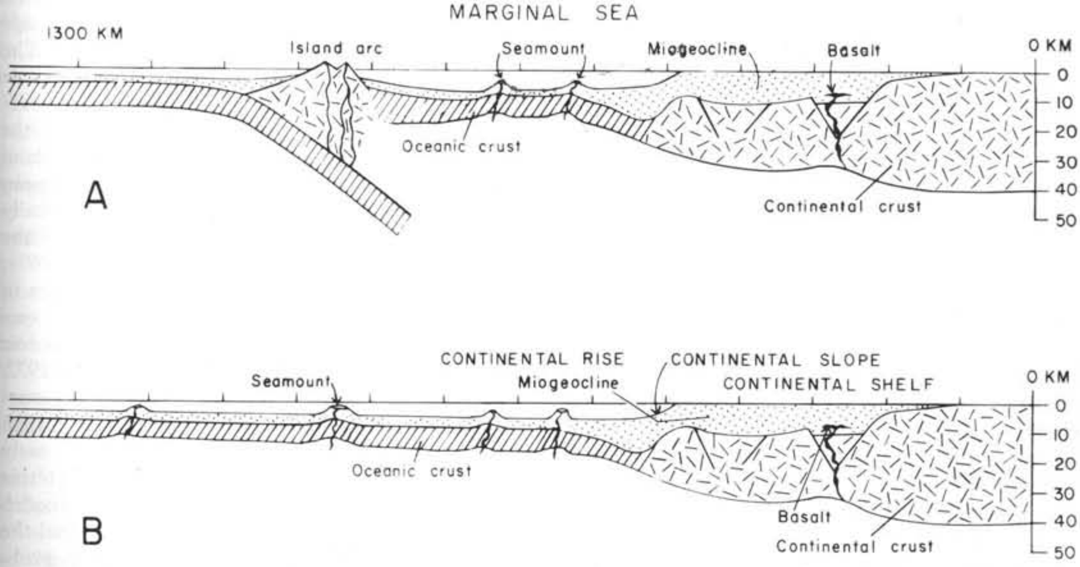


FIG. 14.—Models of lower Paleozoic and uppermost Precambrian Cordilleran geosyncline.

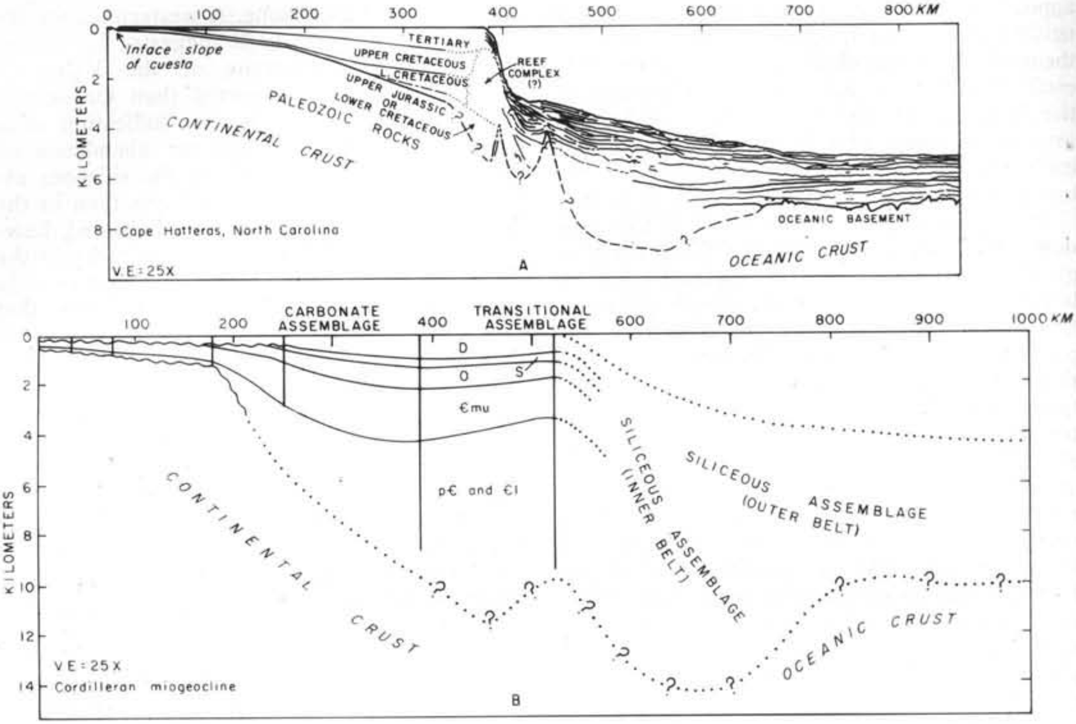


FIG. 15.—Cross sections comparing Cordilleran miogeocline with present-day miogeocline along continental margin of eastern North America. *A,* Section at Cape Hatteras, North Carolina, from Emery and others (1970, fig. 38). *B,* Cordilleran miogeocline showing distribution of assemblages. Based on figure 6 (orientation reversed). Letter symbols same as on figure 6. Dotted lines are hypothetical.

geocline, although uppermost Precambrian and Lower Cambrian quartzite generally contains some feldspar and a few quartz granules and pebbles whereas quartzite of the Valmy does not. The arkosic debris of the Harmony could have been derived from upfaulting and erosion of Precambrian crystalline rocks along the front of the miogeocline, second cycle from older (upper Precambrian?) arkosic deposits on the front of the miogeocline, or from an unknown Precambrian crystalline source, perhaps a westward extension of the craton lying either north or south of the Great Basin.

In the stable continental margin model (Atlantic type), the Cordilleran miogeocline is considered to be a continental terrace deposit with a broad continental shelf and a relatively narrow continental slope bordered on the west by a continental rise and, in turn, by an oceanic basin. In this model, rocks of the carbonate and transitional assemblages are continental shelf deposits, rocks of the transitional and siliceous assemblages represent progressively deeper water deposits to the west on the outer shelf, on the continental slope and rise, and in the ocean basin. The model is patterned after the Mesozoic and Cenozoic miogeocline along the eastern border of the North American continent (fig. 15). This feature is similar in size and shape and contains somewhat similar deposits. The stable continental margin model does not allow for an island arc system to the west. Accordingly, lower Paleozoic assemblages in the Klamath Mountains either are not true island arc deposits, as has been suggested by Hopson (1973), or, if they are, have been tectonically emplaced subsequent to the development of the miogeocline described here. Hamilton (1969), for example, has suggested that assemblages in the Klamath Mountains were tectonically emplaced by underflow of the Pacific in Mesozoic or Cenozoic time. Burchfiel and Davis (1972, fig. 3) speculate that the Klamath Mountains assemblages were related to an island arc system far removed from North America in early Paleozoic time, although they believe that this model is less likely than a marginal sea model.

The arkosic deposits of the Harmony and the mature quartzite of the Valmy present problems in the stable margin model, as in the marginal sea model. Either deposit could be derived from uplift and erosion of blocks along the front of the miogeocline or by slumping along the front, as was suggested in the marginal sea model, or, as was suggested by Churkin (1974), sand of the Valmy may have been derived from the shelf. Alternatively, the source area may have been a microcontinent rifted away from North America (an analogous situation today would be Madagascar), but evidence of such rifting is lacking. Still another possibility is that units like the Harmony and Valmy are not related to deposition on or near the North American continent but rather were deposited on the margin of another continent or an island arc system unrelated to North America and were tectonically emplaced (obducted) on the North American continent during the Late Devonian and Early Mississippian Antler orogeny or subsequent orogenies. This hypothesis is difficult to evaluate but is discounted because of the absence of a major serpentine belt that would mark the suture line at a continent-continent (or a continent-island arc) join, although small scattered bodies of Alpine-type serpentinite (Poole and Desborough, 1973) do occur. A further argument against such tectonic emplacement is that reconstructed facies changes, particularly in Ordovician rocks, seem gradual, indicating a progressive change in lithologic type within a single basin of deposition rather than an abrupt change related to tectonic telescoping of two unrelated facies.

The strongest argument in favor of the stable margin model is the similarity of the Cordilleran miogeocline to the Atlantic coast miogeocline (fig. 15), known to be an accumulation along a tectonically stable margin subsequent to a time of continental separation (Dietz and Holden, 1970). The remarkable lateral extent and lithologic uniformity of lower Paleozoic and uppermost Precambrian strata of the Cordilleran miogeocline indicate a long interval (perhaps as much as 500 my) of tectonic stability in western North America. Such stability is difficult to understand if a subduction zone and island arc system existed to the west. Perhaps the major tectonic activity was confined to the island arc region and North America, which lay landward of the marginal sea, was protected from direct interaction with an oceanic plate farther west. Nonetheless, we would expect that some energy, during part of the 500 my interval, would be transferred to the continent, either by collapse of the marginal sea or by the formation of a subsidiary subduction zone along the edge of the main continental mass.

The tectonic evolution of the Great Basin during the early Paleozoic and latest Precambrian may involve elements of both the stable continental margin model and the marginal sea model. An appealing interpretation is that the Great Basin was a stable continental margin during most of latest Precambrian and early Paleozoic time and that a subduction zone and marginal sea did not develop until late in the early Paleozoic, perhaps in the Devonian. According to this interpretation, Ordovician and Silurian rocks in the Klamath Mountains (Irwin, 1966; Churkin, 1974; Hopson, 1973) and Silurian and related rocks in the northern Sierra Nevada (McMath, 1966) are largely oceanic deposits unrelated to an island arc system, whereas younger rocks, such as the Devonian of the Klamath Mountains, which include thousands of meters of volcanic flows and pyroclastics, are island arc deposits.

The development of the Cordilleran miogeocline, which rests on Precambrian crystalline rocks and Precambrian supracrustal detrital rocks, marks a distinct change in the tectonic setting of western North America (Stewart, 1972). The pregeosynclinal supracrustal detrital rocks occur in platform facies and in major west- and northwest-trending aulacogens within which 3000–6000 m (10,000–20,000 ft) of strata accumulated. The miogeocline cuts across these older structures. If aulacogens are offshoots of geosynclines, as generally believed (Salop and Scheinmann, 1969, p. 586; Hoffman, 1971), then a pre-Cordilleran geosyncline may have been present off North America, but we know of no evidence that would indicate the trend or position of such a geosyncline, if it existed.

A rifting event appears to be the most likely way to form the Cordilleran geosyncline. This rifting could involve a major continental separation (Stewart, 1972) or the reshaping of the margin by rifting and migration of a continental fragment westward. In either case, volcanic rocks in uppermost Precambrian and Lower Cambrian strata in the eastern Great Basin, which are unique for Paleozoic and early Mesozoic rocks of the eastern Great Basin, are believed to mark zones of extension in the continental basement that tapped sources of basalt in the mantle below. These volcanic rocks may be related to extensional events similar to those that formed grabens and led to the eruption of basaltic rocks in the Triassic Newark Group in the eastern United States. The Triassic grabens, presumed to be related to the breakup of the supercontinent of Pangaea and opening of the Atlantic basin (Dietz and Holden, 1970), lie about 320 km (200 mi) inland of the continental slope, a distance comparable to that of upper

Precambrian and Lower Cambrian volcanic rocks of the Great Basin east from the inferred Cordilleran continental margin. According to this interpretation, the Wasatch line marks the eastern edge of continental extension and thinning during late Precambrian rifting.

The outer shelf of the Cordilleran miogeocline is characterized by carbonate shoals and reefs, thinning of sedimentary units, relatively abrupt changes in facies, and unconformities. Shoal or reeflike deposits included bioherms near the shelf edge in the Ordovician (Ross and Cornwall, 1961) and Silurian (Winterer and Murphy, 1960) and perhaps during most periods. The Precambrian Reed Dolomite (Nelson, 1962), for example, in Inyo County, California, changes facies from carbonate and quartzite into dominantly carbonate within a relatively short distance; this may indicate a nearness to a shoal or reef. Most of the rock systems are thinner along the miogeoclinal shelf edge (figs. 8–12), and some individual units, such as the Eureka Quartzite (Ketner, 1968) are missing or thin. Facies changes and unconformities, for example, those in Ordovician rocks in central Nevada (Webb, 1958; Lowell, 1960; Merriam, 1963; McKee and Ross, 1969), are also characteristics of the shelf edge. These shelf edge features may be related to a basement-ridge barrier similar to that on the eastern margin of the North American continent and elsewhere (Drake and others, 1959; Emery and others, 1970; Hedberg, 1970; Burk, 1968).

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