CAMBRIAN AND LATEST PRECAMBRIAN PALEOGEOGRAPHY AND TECTONICS IN THE WESTERN UNITED STATES

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ABSTRACT

The upper Precambrian and Cambrian rocks of the western United States are mostly shallow water deposits that formed on a broad shelf along the western margin of the North American craton. Relatively deep water deposits, some perhaps tectonically emplaced from distant depositional sites, crop out in a few areas near the western margin of the shelf.

The upper Precambrian and Cambrian shelf deposits are divided into three main lithogenetic sequences. These are, in ascending order: (1) a discontinuous sequence, locally as much as 3,000 m thick, of diamictite, mudstone, sandstone, conglomerate, and mafic volcanic rocks of Precambrian age; (2) a westward-thickening wedge, locally more than 6,000 m thick, of upper Precambrian and Lower Cambrian terrigenous detrital rocks of shallow-water, in part tidal, origin; and (3) a Middle and Upper Cambrian sequence of predominantly shallow water carbonate rocks generally about 2,000 m thick.

The rate of accumulation of strata in the Cordilleran geosyncline was apparently greatest in Early Cambrian and presumably also in late Precambrian time and decreased progressively through Ordovician and perhaps younger early Paleozoic time. The rate of accumulation and inferred subsidence appears to decline exponentially with a time constant of about 50 m.y. Accumulation and subsidence rates due to thermal contraction in the Cretaceous to Holocene Atlantic and Gulf shelves of the United States are also exponential with a similar time constant, as are subsidence rates for oceanic crust spreading away from a mid-oceanic ridge. In these cases, thermal contraction followed an uplift related to heating at a spreading center. The exponential decrease in the rate of subsidence in the Cordilleran miogeocline suggests a similar history of thermal contraction after some initial heating event, perhaps related to fragmentation and reshaping of the western margin of the United States by rifting in the late Precambrian. The diamictite and volcanic sequence may have been deposited in a rift valley formed during an early stage of the rifting; the terrigenous detrital sequence may be related to erosion of a thermally expanded area near the rift, and the carbonate sequence to widespread shallow-water shelf sedimentation after destruction of the thermally uplifted area by cooling and thermal contraction and by erosion.

INTRODUCTION

The upper Precambrian and Cambrian rocks of the western United States consist of as much as 10,000 m

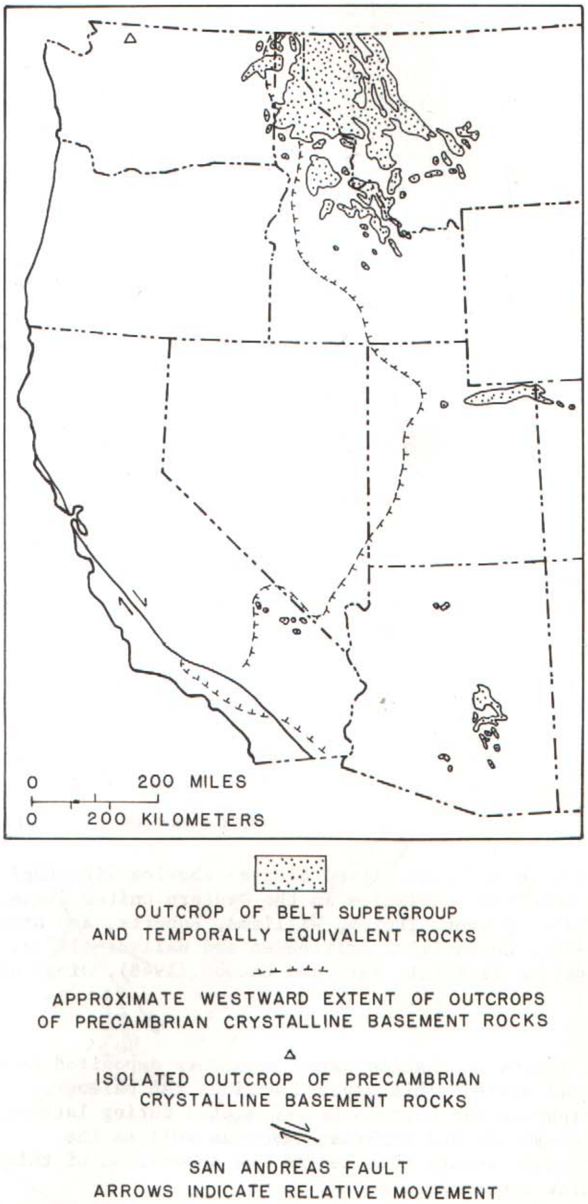


Figure 1. Distribution of crystalline basement rocks (2,400-1,450 m.y.) and Belt-age (1,450-900? m.y.) unmetamorphosed sedimentary and volcanic rocks in the western United States. Based in part on King (1976) and Armstrong(1975).

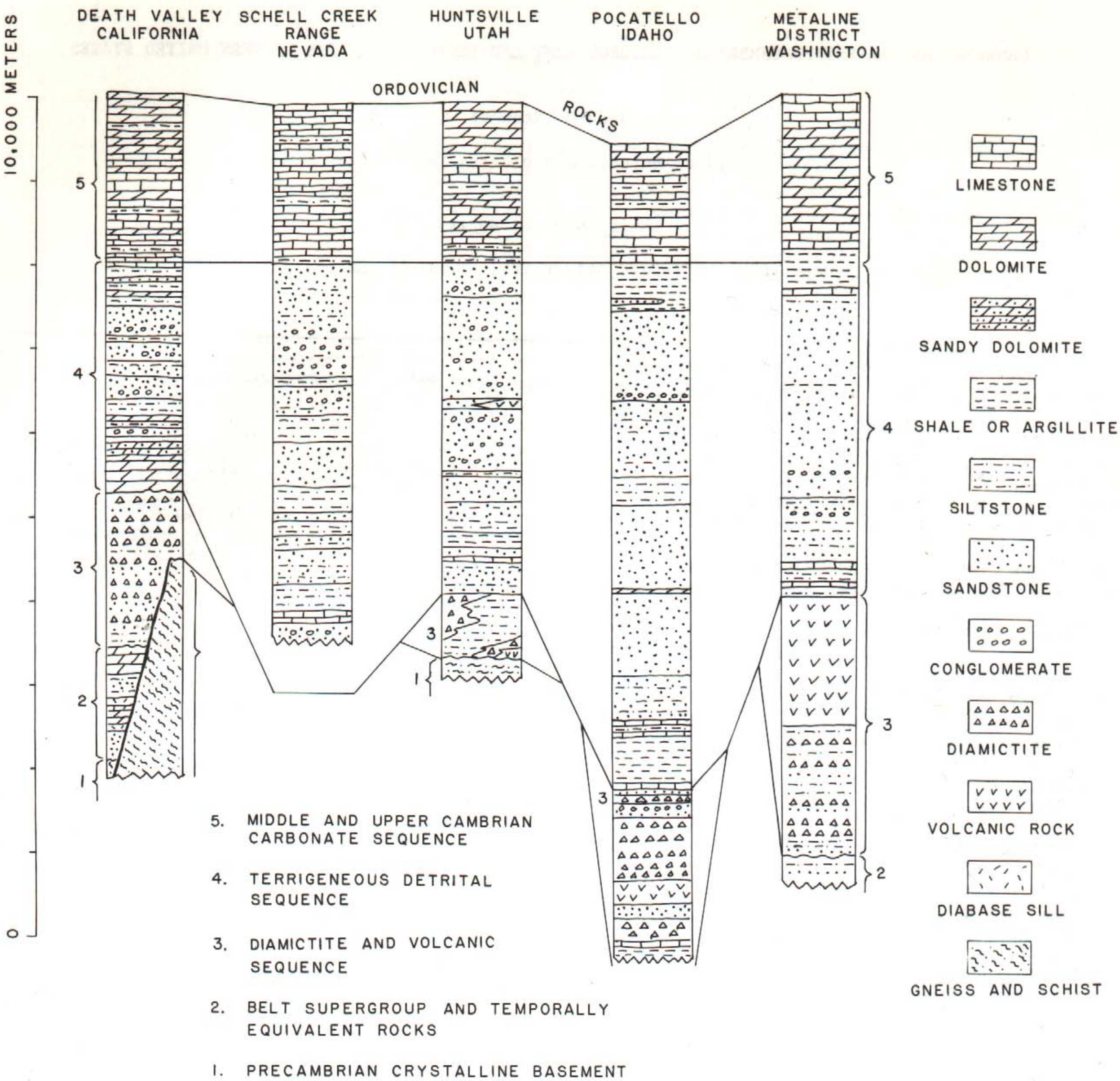
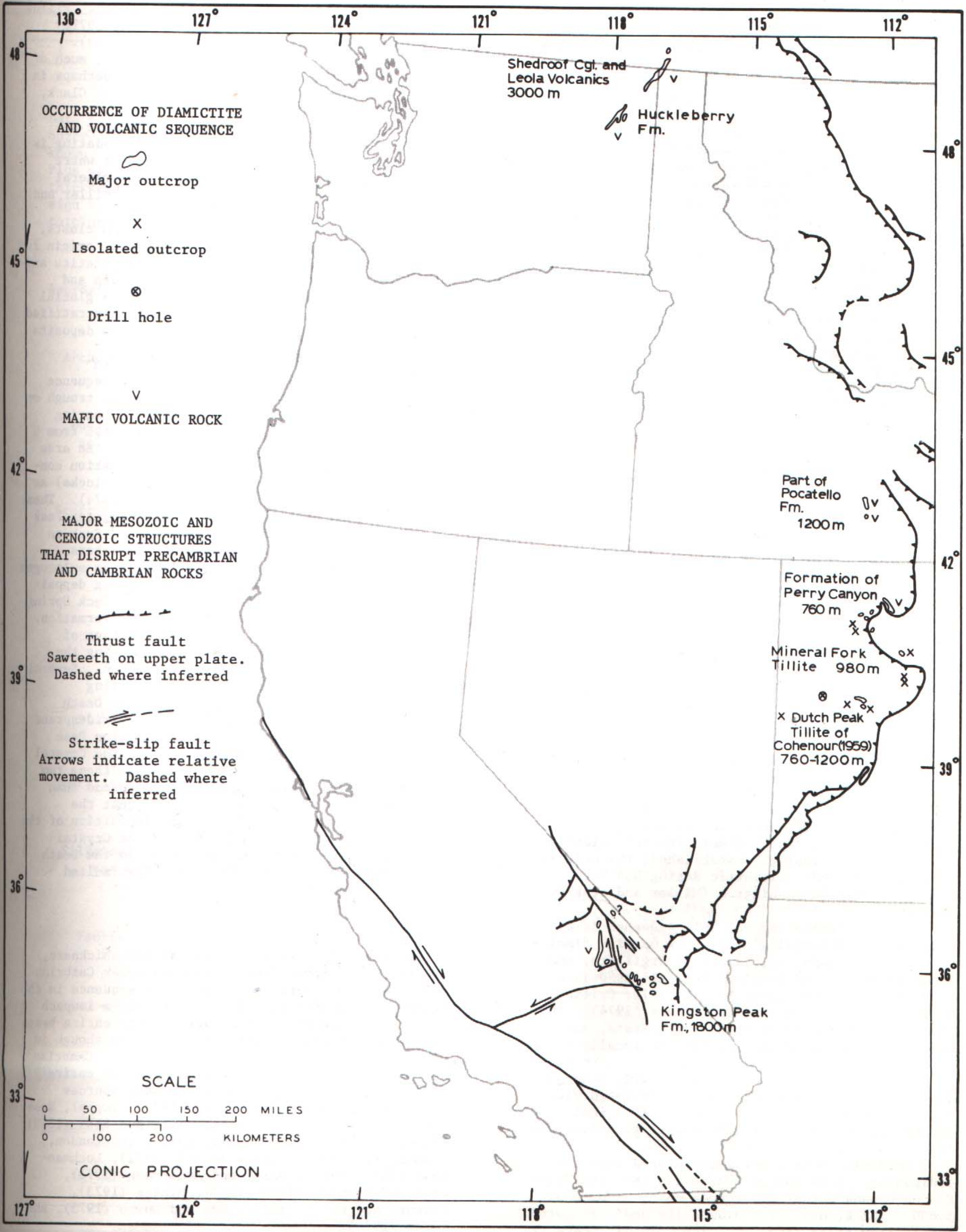


Figure 2. Generalized diagram showing lithologic types and correlation of late Precambrian and Cambrian strat at selected localities in the western United States. Sources of data: Death Valley, Hazzard (1937), Stewart (1970), Wright, Troxel, Williams, Roberts, and Diehl (1974); Shell Creek Range, Young (1960), Hose and Blake (1976); Huntsville, Crittenden and Wallace (1973), Sorensen and Crittenden (1976); Pocatello, Trimble (1976); Metaline district, Park and Cannon (1943), Dings and Whitebread (1965), and Yates (1976).

of generally shallow water sediments deposited on a broad shelf. This paper describes the paleogeography of the western United States during late Precambrian and Cambrian time, as well as the tectonic events that led to the deposition of this thick sequence of rocks.

This report focuses on three major sequences of Precambrian and Cambrian rocks: an upper Precambrian diamictite and volcanic sequence (900? to 800? m.y.), an uppermost Precambrian and Lower Cambrian terrigenous detrital sequence (650? to 540 m.y.), and a Middle and Upper Cambrian carbonate sequence (540-500 m.y.). Rocks older than these sequences consist of Precambrian crystalline metamorphic and intrusive rocks (about 2,400 to 1,450 m.y.), the oldest rocks in the western United States, and of relatively unmetamorphosed sedimentary and volcanic rocks of the Belt Supergroup and temporally equivalent rocks (Fig. 1). The Belt Supergroup and temporally equivalent rocks occur in deep epicratonic troughs, some containing as much a

Figure 3. Distribution and maximum thickness of Precambrian (900?-800? m.y.) diamictite and volcanic sequence in the western United States. Based on many sources, including Cohenour (1959), Wright and Troxel (1967), Crittenden, Scheaffer, Trimble, and Woodward (1971), Miller and Clark (1975), and Yates (1976).



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20,000 m of sediment (Harrison and others, 1974), or are widespread, relatively thin shelf deposits. More detailed information on the crystalline basement rocks and the Belt Supergroup and temporally equivalent rocks is given elsewhere (Harrison and others, 1974; King, 1976).

DIAMICTITE AND VOLCANIC SEQUENCE (900? TO 800? M.Y.)

The diamictite and volcanic sequence consists of as much as 3,000 m of diamictite, mudstone, argillite, sandstone, and conglomerate, and locally thick units of volcanic rocks. The highly distinctive diamictite (a rock composed of large clasts set in a finer grained matrix) is considered to be of glacial origin. The sequence was grouped by Stewart (1972) with the overlying terrigenous detrital sequence but is here described separately because of its possible tectonic importance.

In some areas, the diamictite and volcanic sequence rests unconformably on crystalline basement rocks, whereas in other areas it rests unconformably, for the most part, on the Belt Supergroup and temporally equivalent rocks. Near the United States-Canada border, the Belt Supergroup and equivalent Purcell Supergroup of Canada were uplifted and mildly folded during the East Kootenay orogeny (White, 1959) before deposition of the diamictite and volcanic sequence.

In the western United States, the diamictite and volcanic sequence is recognized in the northeastern Washington and northern Idaho area, as well as in southern Idaho, north-central and westernmost Utah, and eastern California (Figs. 2 and 3). In each of these areas, it is the only unit in the Precambrian and Paleozoic sequence that contains diamictite, and in most of these same areas it is the only unit that contains abundant volcanic rocks. These characteristics suggest that its correlation from region to region in the western United States is reasonable (Crittenden and others, 1972; Stewart, 1972). In addition, the diamictite and volcanic sequence occurs at approximately the same stratigraphic position in each area; that is, above the Belt Supergroup or rocks correlated with it on the basis of radiometric dating or the types of algal structures, and below terrigenous detrital rocks that contain Lower Cambrian fossils in the upper part. The correlation is also reasonable if the diamictites represent glacial deposits formed during a relatively short glacial epoch. Nonetheless, correlation of the diamictite and volcanic unit is not conclusively established; the unit is discontinuous and radiometric dating has been completed only in Washington (Miller and others, 1973).

The diamictite and volcanic sequence is a laterally intertonguing complex of massive diamictite, boulder mudstone, laminated argillite, and minor sandstone and conglomerate, all associated with mafic volcanic rocks (Aalto, 1971; Crittenden and Wallace, 1973; Wright and others, 1974). The diamictites commonly contain large clasts, many of which are a meter or more across, of locally derived rocks. Crittenden and Wallace (1973) report dropstones, as large as 2 m in size, showing penetrative deformation of enclosing laminae. Striated clasts occur but are not abundant. In thick basinal sequences, the diamictites form well-graded turbidites.

Volcanic rocks occur interlayered with the sedimentary rocks and in the region near the United States-Canada border occur in a unit, as much as 2,800 m thick, overlying diamictite and associated sedimentary rocks (Fig. 2). The volcanic rocks are primarily greenstone and consist of flows, breccias, and minor amounts of tuff (Crittenden and others, 1971; Miller and Clark, 1975; Yates, 1976). Pillow greenstones are reported locally (Aalto, 1971; Stewart, 1972; Crittenden and Wallace, 1973). Although the greenstone is highly altered, much of it seems to have been a basalt that was perhaps in part tholeiitic (Stewart, 1972; Miller and Clark, 1975; Yates, 1976).

The age of the diamictite and volcanic sequence is poorly known. The only direct dating is of greenstone in northeastern Washington; where potassium-argon ages on whole rocks and mineral separates are between 827 and 918 m.y. (Miller and others, 1973).

The presence of dropstones, striated clasts, and massive diamictite suggests a glacial origin for many of the sedimentary rocks in the diamictite and volcanic sequence (Aalto, 1971; Crittenden and Wallace, 1973). Aalto (1971) indicates a glacial marine environment along a coast, with unstratified subglacial till inland and turbidity flow deposits oceanward.

In the Death Valley region of eastern California, the diamictite and volcanic sequence apparently accumulated in a fault-bounded trough or rift valley basin. Here the diamictite-bearing Kingston Peak Formation ranges in thickness from 0 to over 1,800 m within a few kilometers. An area close to the presumed margin of the formation contains gigantic blocks (presumably slide blocks) as much as 300 m long (Wright and others, 1974). These relations seem best explained if the Kingston Peak was deposited in a fault-bounded basin. Such an origin has been proposed by Wright and others (1974), although they call this feature an aulacogen and suggest that it first developed during deposition of the Belt-age Crystal Spring and Beck Spring Formations underlying the Kingston Peak Formation. They indicate, however, that the main stage of subsidence in the aulacogen occurred during the deposition of the Kingston Peak Formation. Several units in the Crystal Spring and Beck Spring Formations are remarkably uniform in the Death Valley area and perhaps were originally widespread shelf deposits. In addition, the Kingston Peak Formation contains debris derived from the Crystal Spring and Beck Spring, indicating that these two formations were at one time more widespread than their present outcrop area. We feel that the evidence for a trough prior to the deposition of the Kingston Peak is equivocal and that the Crystal Spring and Beck Spring are preserved in the Death Valley area only because they were downfaulted

Figure 4. Distribution, generalized thickness, and facies of upper Precambrian and Lower Cambrian (650?-540 m.y.) terrigenous detrital sequence in the western United States. East of the 300-m isopach line, the thicknesses shown are for the entire basal Cambrian terrigenous detrital unit, even though in places this unit probably includes Middle Cambrian rocks and locally may be predominantly or entirely Middle Cambrian in age. Based on many sources including Beutner and Scholten (1967), Ruppel, Ross, and Schleicher (1975), Hobbs, Hays, and Ross (1968), Miller (1970), Stewart (1970, 1974), Crittenden, Schaeffer, Trimble, and Woodward (1971), Lochman-Balk (1971, 1972), Oriel and Armstrong (1971), Crittenden and Wallace (1973), Hintze (1973), Stewart and Poole (1975), Zen and Dutro (1975), and Yates (1976).

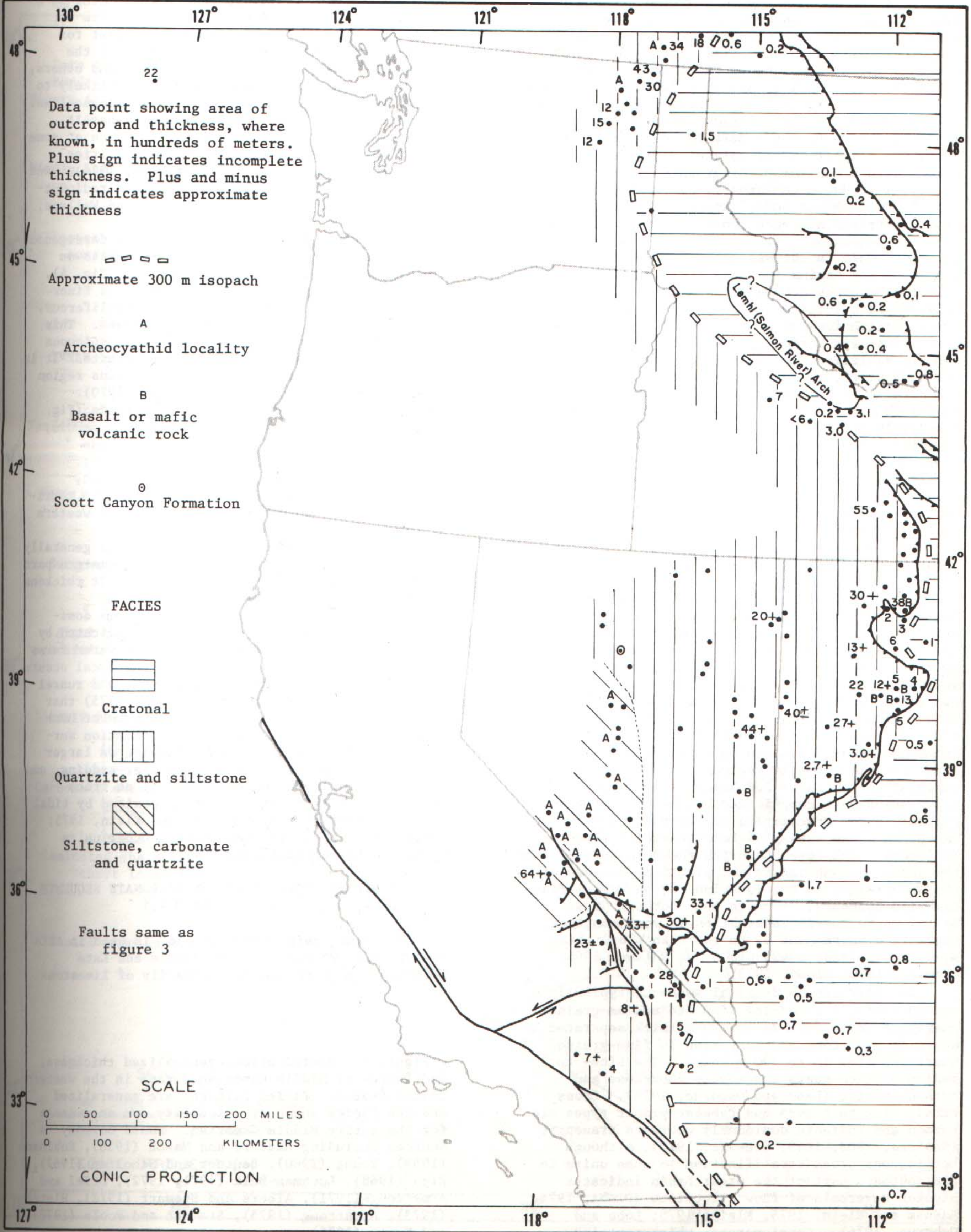


Figure 4. Caption on adjoining page.

during development of the fault-bounded basin and preserved below the Kingston Peak Formation.

Late Precambrian normal faulting has been described also in southern Canada near the international boundary by Price and Lis (1975). Although the exact relation of the diamictite and volcanic sequence to the faulting is not described, the diamictite and volcanic sequence is locally thick in southern Canada and its distribution and thickness trends could be fault related. Price and Lis (1975) indicate that the Belt-Purcell Supergroup, below the diamictite and volcanic sequence, was offset with a maximum stratigraphic separation of 13 km before it was unconformably overlapped by so-called "Eo-Cambrian beds," apparently the terrigenous detrital sequence above the diamictite and volcanic sequence.

Elsewhere in the western United States, the origin of the basins in which the diamictite and volcanic sequence accumulated is uncertain. In Utah, the diamictite locally fills a 400-m deep basin (Crittenden and others, 1952) in the upper part of the Big Cottonwood Formation (a unit temporally equivalent to the Belt Supergroup). This basin, however, is entirely erosional, and the same subunits in the Big Cottonwood Formation can be mapped on either side of the basin (Crittenden and others, 1952). Except for Death Valley and the Canadian areas described above, faulting is not known or suspected to be a factor in the development of the basins in which the diamictite and volcanic sequence was deposited, although evidence of such faulting would be difficult to see elsewhere because of the sparsity of outcrops of rocks of this age.

TERRIGENOUS DETRITAL SEQUENCE (650? TO 540 M.Y.)

Upper Precambrian and Lower Cambrian rocks in the western United States consist of a west-thickening wedge of terrigenous detrital rocks composed of quartzite and siltstone and minor conglomerate (Figs. 2 and 4). This entire sequence of upper Precambrian and Lower Cambrian rocks is considered together here, because the sequence is lithologically similar, both within individual stratigraphic sections and in different regions. Although the sequence can be divided into distinctive formation in some regions, interregional correlations of these formations are not complete. Correlations are complicated by the lack of fossil control in the Precambrian and by the sparsity, except in a few areas, of Lower Cambrian fossils. To describe the Lower Cambrian rocks separately from the upper Precambrian is not practical because of the uncertainty in the location of the Precambrian-Cambrian boundary. Even in the southern Great Basin, where fossil information is most complete, there is disagreement as to where to place this boundary (Nelson, 1962; Cloud and Nelson, 1966; Stewart, 1970; Cloud, 1973).

The terrigenous detrital sequence typically consists of cliff-forming fine- to medium-grained quartzite units from 25 to 100 m thick separated by units of siltstone and very fine to fine-grained quartzite from 15 to 300 m thick. The fine- to medium-grained rocks are mostly subarkose and orthoquartzite (Lobo and Osborne, 1976). Cross strata of both trough and tabular planar types are common and indicate dominantly westward transport (Seeland, 1968, 1969; Stewart, 1970), although herringbone cross stratification in some units in the southern part of the Great Basin indicates bipolar reversals of flow directions (Diehl, 1974; Barnes and Klein, 1975, Klein, 1975; Lobo and Osborne, 1976). Conglomerate, which contains pebbles of quartz and quartzite, is sparsely distributed in the quartzite. Limestone and dolomite are present in layers less than a meter to a few hundred meters thick. Individual units of quartzite, siltstone, or carbonate persist for hundreds of kilometers along the trend of the miogeocline (Stewart, 1970; Crittenden and others, 1971), but are less persistent and more likely to change facies across the trend of the miogeocline. Thin mafic lava flows (Fig. 4), mostly basalt, occur in the terrigenous detrital sequence at some localities in Nevada and Utah (see localities listed in Stewart, 1972; also Stewart, 1974). One mafic volcanic breccia in Utah is dated radiomettrically as 570 m.y. old (Crittenden and Wallace, 1973).

A somewhat different facies of the terrigenous detrital sequence occurs in the Great Basin in eastern California and southern Nevada (Fig. 4). Here the terrigenous detrital sequence is finer grained, richer in carbonate, more fossiliferous, and thicker than the sequence to the east. This facies consists mostly of siltstone, conspicuous carbonate units, and fine-grained quartzite. It is best developed in the Inyo-White Mountains region of California (Nelson, 1962; Stewart, 1970). Fossils include trilobites, archeocyathids (Fig. 4), pelecypods, echinoderms, pelmatozoan debris, *Hyolithes*, *Salterella*, *Skolithus*, and algae (Stewart, 1970; Palmer, 1971; Alpert, 1976; Gangloff, 1976; Nelson, 1976). In contrast, fossils generally are extremely rare in the terrigenous detrital sequence elsewhere in the western United States.

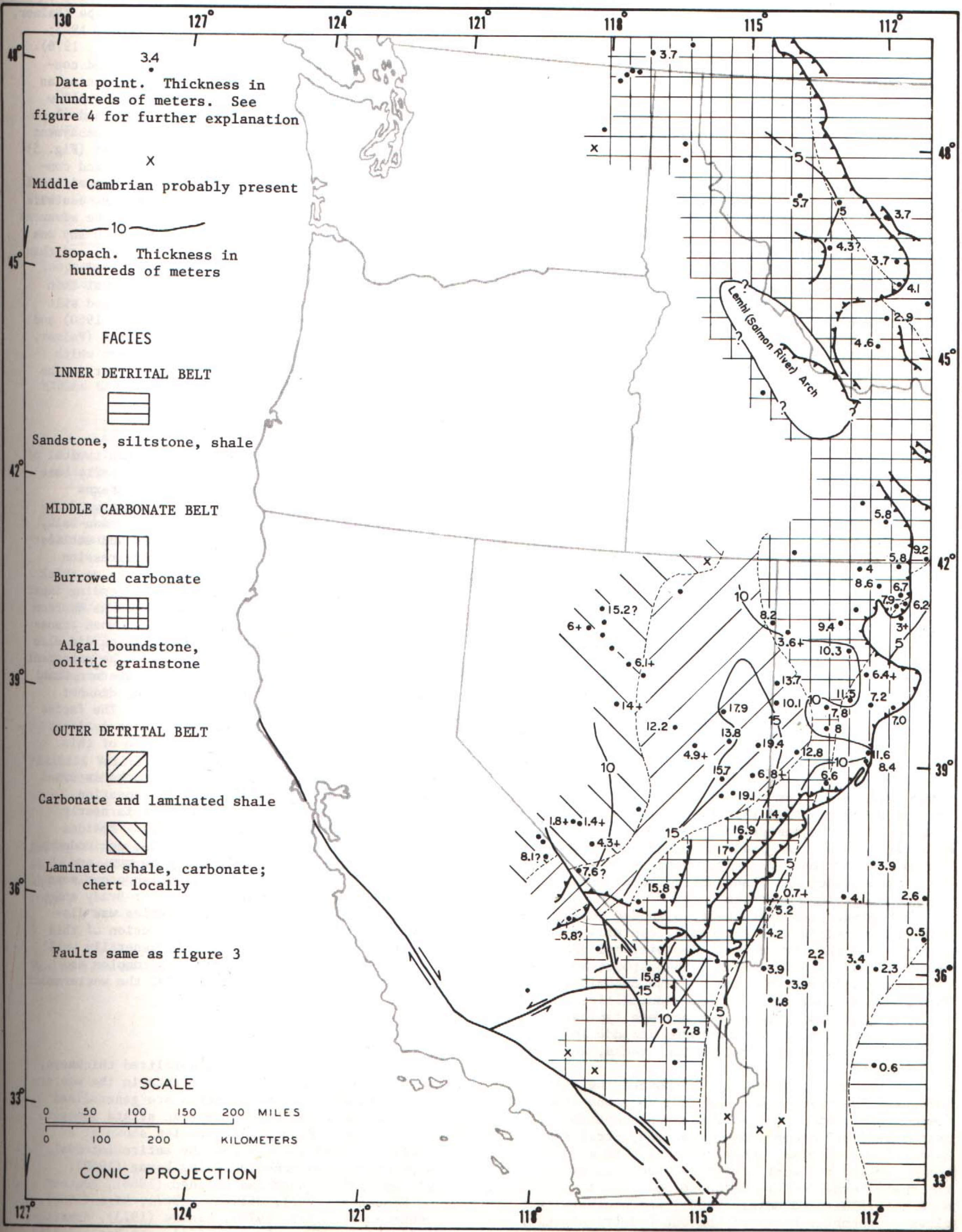
The terrigenous detrital sequence is generally only a few hundred meters thick in the eastern part of the western United States (Fig. 4). It thickens westward to locally over 6,000 m.

The terrigenous detrital sequence is dominantly of shallow-water origin as is indicated by the abundance of such presumed shallow-water forms as algae and archeocyathids and by the local occurrence of mudcracks, raindrop imprints, and runzel marks (Barnes and Klein, 1975; Klein, 1975) that indicate local exposure. The occurrence of herringbone cross stratification, reactivation surfaces, superimposition current ripples on larger current ripples, flaser and lenticular bedding, and other sedimentary structures as well as trace fossils indicate that sediments deposited by tidal currents (Stewart, 1970; Barnes and Klein, 1975; Klein, 1975) occur in some units. Other units appear to be subtidal sand bodies.

MIDDLE AND UPPER CAMBRIAN CARBONATE SEQUENCE (540 TO 500 M.Y.)

The name "carbonate sequence" is used in this report (Fig. 2) for rocks of Middle and Late Cambrian age which consist primarily of limestone

Figure 5. Distribution, generalized thickness, and facies of Middle Cambrian strata in the western United States. Facies patterns are generalized and the facies shown at a locality are an average for the entire Middle Cambrian. Based on many sources including Hazzard and Mason (1936), Robinson (1959), Young (1960), Beutner and Scholten (1967), Rigo (1968), Lochman-Balk (1971, 1972), Oriel and Armstrong (1971), Albers and Stewart (1972), Hintze (1973), Armstrong (1975), Stewart and Poole (1975), and Yates (1976).



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and dolomite. These rocks represent a marked change in lithology from the underlying terrigenous detrital sequence of late Precambrian and Early Cambrian age. The Middle and Upper Cambrian rocks of the carbonate sequence are described separately below.

Middle Cambrian rocks

Middle Cambrian rocks consist mostly of dolomite and limestone with some shale and quartzite. During the Middle Cambrian, the transgression of the sea onto the craton that began in the late Precambrian continued. The thickness of these rocks followed the same general pattern as underlying rocks, with sediments thin in the east and thickening towards the west (Fig. 5); but the total thickness is commonly less. In the western part of the region the thickness of the Middle and Upper Cambrian rocks together is only a third that of the upper Precambrian and Lower Cambrian rocks.

Middle Cambrian carbonate and shale are moderately fossiliferous and, as a result, correlations and dating are better than in the Lower Cambrian. Trilobites, brachiopods, algae, and ichnofossils are the main fossil types; mollusks, conodonts, and sponges occur more rarely. Dating is based on established trilobite zones, but facies changes complicate the problem of age assignment. Boundaries of some trilobite zones originally thought to be time boundaries may actually be facies boundaries with fossil differences due to change in biofacies rather than time (Lochman, 1957; Robison, 1976; Palmer and Campbell, 1976). Although complicated by this facies problem, control on the Middle Cambrian-Upper Cambrian boundary generally is good, except in western and northern exposures where structures are complex. The base of the Middle Cambrian is more difficult to locate because the Lower Cambrian sections are so sparsely fossiliferous. Here correlations are based on similarities of lithic content and sequences of units, and the boundary is only approximately located in many areas. Sedimentation was generally continuous across the Lower-Middle Cambrian boundary.

Palmer (1960, 1971) showed that three distinct lithofacies belts (inner detrital, middle carbonate, and outer detrital) could be recognized for the Upper Cambrian of the western United States. Robison (1960) applied this concept to the Middle Cambrian. In the western United States, the inner detrital belt of the Middle Cambrian occurs only in Montana and Arizona (Fig. 5), where sand, silt, and clay, thin bedded and glauconitic (Lochman-Balk, 1971), are typical deposits.

The middle carbonate belt, although dominantly thick-bedded to massive carbonate, contains a wide variety of rocks with complex depositional patterns (Kepper, 1972). Stromatolitic boundstone, mottled carbonate, thin-bedded limestone and dolomite, and thick-bedded bioclastic, oolitic, oncolitic, and pelleted grainstone and wackestone are typical. Organic and sedimentary structures include animal burrows, ripple marks, desiccation cracks, and some crossbedding.

In the outer detrital belt, toward the west, laminated shaly siltstone and lime mudstone are common, and grainstone and wackestone contain graded bedding, slump structures, and intraformational conglomerates. Chert generally is more common toward the west. Volcanic rocks occur in the Shwin Formation of the Shoshone Range in central Nevada (Gilluly and Gates, 1965), which otherwise closely resembles rocks of the outer detrital belt.

The lithofacies patterns are interpreted to be the result of a shallow shelf in eastern areas followed to the west in turn by carbonate shoals, a deeper water open shelf, and a basinal slope (Palmer 1971; Kepper, 1972, 1976; Cook and Taylor, 1975; Taylor and Cook, 1976; Palmer and Campbell, 1976). The position and shape of the shoal changed continuously throughout this time span, producing an interfingering pattern with rocks of the shallow shelf. The position of the deeper shelf and slope remained more stable, although the broad embayment of deeper facies rocks into the shoal area (Fig. 5) did not begin until late Middle Cambrian and continued into early Late Cambrian time. Depositional patterns were also affected by the general eastward transgression of the sea. As the shoreline advanced toward the east, the deposition of sand in any one place gave way gradually to silt and clay, and then to carbonate (Lochman, 1957; Scholten, 1957; Lochman-Balk, 1971; Palmer, 1971). Several thin widespread deposits of terrigenous sand and silt occur in Middle Cambrian rocks (Robison, 1960) and may be related to local nearshore uplifts (Palmer, 1956) or to a temporary marine regression which allowed sediment to be transported farther to the west than usual (Kepper, 1972).

Upper Cambrian rocks

The pattern of carbonate deposition typical of the Middle Cambrian continued into the early Late Cambrian (Dresbachian), with facies patterns similar to those shown in Figure 5. Eastward transgression of the sea continued (Lochman-Balk, 1972). During the latest Dresbachian and earliest Franconian stages, however, a brief regression occurred, and shale and sandstone layers as thick as 60 m were deposited, temporarily extending inner detrital belt sedimentation as far west as eastern Nevada (Lochman, 1957; Palmer, 1971). When transgression resumed, carbonate sedimentation did also and was more widespread than before. The embayment of silty rocks in eastern Nevada and western Utah that characterized the Middle Cambrian did not exist after earliest Franconian time. The facies patterns shown on Figure 6 are for post-earliest Franconian time, after the destruction of this embayment.

Correlation of Upper Cambrian rocks is based generally on trilobite and conodont zonation, although dating based on brachiopods is nearly established (Rowell and Brady, 1976). Besides trilobites and brachiopods, mollusks, echinoderms, algae, and sponges are relatively common, and conodonts, graptolites, and jellyfish occur more rarely (Lochman, 1957; McBride, 1976; Brady and Rowell, 1976). The base of the series was discussed under the Middle Cambrian section of this report. The top of the series is generally well defined except in the structurally complex and relatively unfossiliferous rocks of the westernmost

Figure 6. Distribution, generalized thickness, and facies of Upper Cambrian strata in the western United States. Facies patterns are generalized and shown only for Upper Cambrian strata younger than lowermost Franconian. Facies shown at a locality are an average for the entire interval. Based on many sources including Young (1960), Kellogg (1963), Hotz and Willden (1964), Beutner and Scholten (1967), Gilluly (1967), Rigo (1968), Lochman-Balk (1971, 1972), Hintze (1973), Armstrong (1975), Stewart and Poole (1975), and Yates (1976).

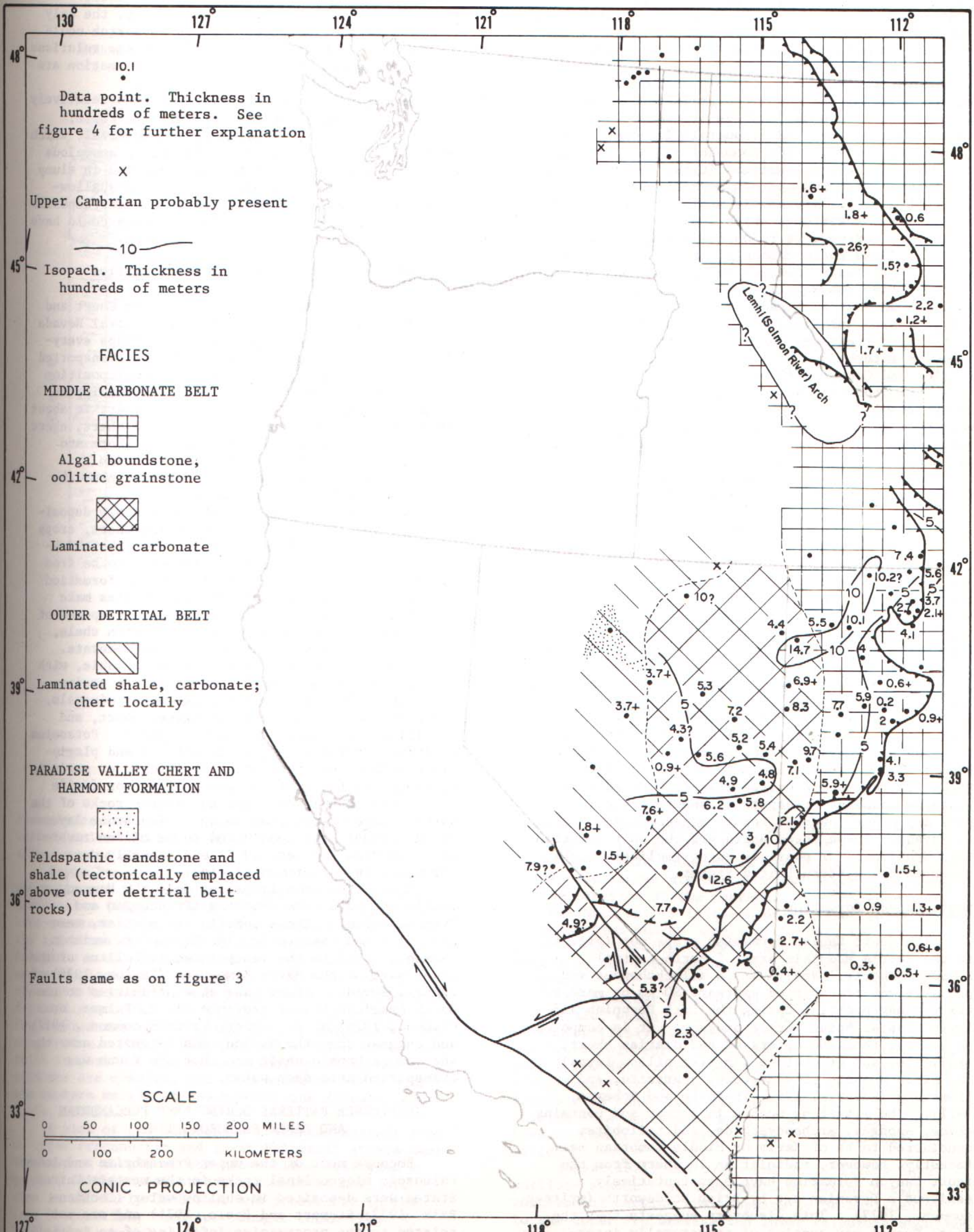

Figure 6. Caption on adjoining page.

exposures.

During Late Cambrian time after the earliest Franconian, the deposits of the inner detrital belt and the restricted shelf were east of the area of Figure 6 because of continued eastward transgression of the sea. In western Montana and central Utah characteristic deposits are medium- to thick-bedded skeletal and oncolitic grainstone, commonly with trough crossbeds, massive or crossbedded oolitic grainstone, and stromatolitic boundstone (Brady and Rowell, 1976; Lohmann, 1976). Much of the rock has been dolomitized (Bick, 1966). In westernmost Utah and eastern Nevada, post-Dresbachian rocks are mainly medium- to thick-bedded lime wackestone, which are characteristically so burrowed as to obliterate depositional fabric. These rocks commonly crop out as massive ledges. Bioclasts in them are unbroken and unsorted (Brady and Rowell, 1976). Nodular chert is common. In central Nevada, Upper Cambrian rocks are laminated dark-colored lime mudstone and turbidity-current and debris-flow grainstone deposits interbedded with mudstone and siltstone (Taylor and Cook, 1976).

This depositional pattern is interpreted to be the result of a narrow shoal bordered on the west by a deeper water shelf, with sedimentation below wave base, bordered in turn by the continental slope. The shelf was locally emergent during the middle Late Cambrian in western Montana, central Idaho, and central Utah (Lochman, 1957).

UNUSUAL CAMBRIAN ROCKS

Three formations of Cambrian or partly Cambrian age in north-central Nevada are so different lithologically from any other Cambrian units in the western United States that they are described separately here. One of these, the Scott Canyon Formation, is considered part of the siliceous and volcanic (eugeosynclinal) assemblage of Nevada and is somewhat similar lithologically to Ordovician, Silurian, and Devonian eugeosynclinal rocks in Nevada. The Paradise Valley Chert and Harmony Formation, although they have been classified as parts of the eugeosynclinal assemblage (Stewart and Poole, 1974), are difficult to relate to other rocks in the western United States. The Harmony contains abundant feldspathic sandstone and is not like any other lower Paleozoic eugeosynclinal unit in Nevada. These three units are the only Cambrian eugeosynclinal, or possibly eugeosynclinal, rocks known in the western United States.

Scott Canyon Formation

The Scott Canyon Formation crops out (Fig. 4) in the Battle Mountain area of north-central Nevada (Roberts, 1964). It is allochthonous and probably was transported eastward 80 km or more to its present position during the Late Devonian and Early Mississippian Antler orogeny. It is composed of several thousand meters of radiolarian chert, argillite, and greenstone (in part pillow lavas) and minor amounts of sandstone, quartzite, and limestone (Roberts, 1964; Theodore and Roberts, 1971). The limestone occurs in lenses and contains algae, sponges, archeocyathids, and trilobites considered to be of Early or Middle Cambrian age. Recently, however, radiolarian in chert from the Scott Canyon Formation have been tentatively assigned a Devonian age by Brian Holdsworth (written commun., 1977). This discovery suggests that the Scott Canyon may consist of structurally interleaved rocks of at least two ages, a structural relation commonly mapped in other eugeosynclinal terranes in Nevada (Gilluly and Gates, 1965; Stanley and others, 1977). Conceivably, the only Cambrian part of the Scott Canyon Formation could be the dated limestone, but the exact age relations of various parts of the Scott Canyon Formation are not presently known.

The Scott Canyon Formation may be a relatively deep water oceanic deposit (Stewart and Poole, 1974). If so, the contained fossil material, which is indicative of allow-water origin, is anomalous and must be explained either as occurring in slump deposits or in structurally interleaved shallow-water layers in a predominantly deep-water deposit. Alternatively the shallow-water deposits could have been on the flanks of oceanic volcanoes.

Paradise Valley Chert and Harmony Formation

The Upper Cambrian Paradise Valley Chert and Harmony Formation crop out in north-central Nevada (Fig. 6). They are in places, and perhaps everywhere, allochthonous and probably were transported tectonically eastward into their present position during the Late Devonian and Early Mississippian Antler orogeny. The Paradise Valley Chert is about 100 m thick and made up of thin-bedded chert, chert breccia, and some shale and limestone (Hotz and Willden, 1964). It is early Late Cambrian (Dresbachian) in age and crops out only near the center of the Hot Springs Range.

The Harmony Formation, which probably depositionally overlies the Paradise Valley Chert, crops out in north-central Nevada (Fig. 6). The thickness of the Harmony has been estimated to be from 600 to 1,300 m, although the top of the formation is not exposed and structural complexities make thicknesses uncertain. The Harmony is composed of feldspathic sandstone and siltstone, with shale, minor limestone, and minor pebble conglomerate. The composition of the sandstone is variable, with quartz constituting 50-80 percent of the grains, feldspar 10-30 percent, and mica, heavy minerals, and lithic clasts (mostly quartzite, chert, and argillite) the remainder (Suczek, 1977). Potassium feldspars (orthoclase and microcline) and plagioclase both occur, indicating a plutonic or gneissic source rock; radiometric dating of zircon (Jaffe and others, 1959) indicates the source rocks of the Harmony were Precambrian in age. Sandstone layers in the Harmony are considered to be midfan turbidite deposits (Suczek, 1977) and probably were deposited on a continental rise.

Trilobites from limestone date the Harmony as middle and late Late Cambrian (Franconian and Trempealeauan). These fossils are similar, taxonomically and texturally, (M. E. Taylor, written commun., 1977) to the redeposited tribolites of the lower part of the Hales Limestone (Taylor, 1976) of central Nevada. These taxa show affinities to the North American faunal province (A. R. Palmer, oral commun., 1976; M. E. Taylor, written commun., 1977) and suggest that the Harmony was deposited near the shallow carbonate shelf and that the fauna was transported into deep water.

SUBSIDENCE PATTERNS DURING LATE PRECAMBRIAN AND EARLY PALEOZOIC TIME

Because most of the upper Precambrian and lower Paleozoic miogeoclinal rocks in the western United States were deposited in shallow water (Lochman-Balk, 1972; Stewart and Poole, 1974) and are not related to the progressive infilling of an initially deep basin, the accumulation of the thick upper

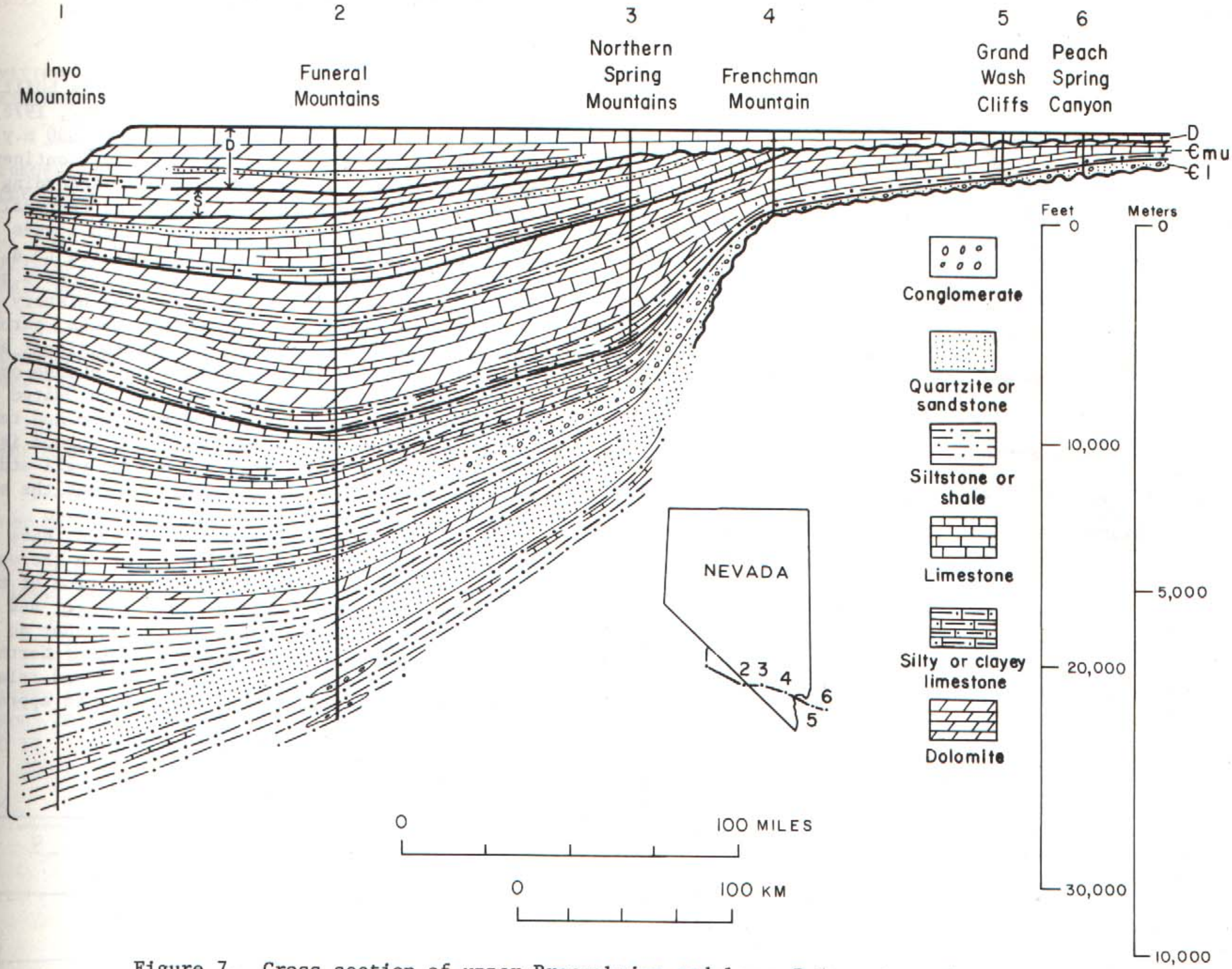

Figure 7. Cross section of upper Precambrian and lower Paleozoic rocks in the southern Great Basin (after Stewart and Poole, 1974). Symbols: p€ and €l, upper Precambrian and Lower Cambrian; €mu, Middle and Upper Cambrian; O, Ordovician; S, Silurian; D, Devonian.

Precambrian and lower Paleozoic section in the western United States must be due to subsidence over a long period of time.

Upper Precambrian and lower Paleozoic rocks generally are thickest in the western part of the miogeocline (Fig. 7), and thus subsidence must have been greatest there also. The eastward decrease in the thickness of upper Precambrian and lower Paleozoic rocks is caused mainly by thinning and eastward pinchout of individual units in the uppermost Precambrian and Lower Cambrian terrigenous detrital sequence. These units thin abruptly to a few hundred meters along the eastern margin of the miogeocline. Middle Cambrian to Devonian strata are thick in parts of the miogeocline, but thickness patterns are complex, and some units are thin near the western margin of the miogeocline (Stewart and Poole, 1974).

A plot of the cumulative thickness against time of upper Precambrian and lower Paleozoic strata near the western margin of the miogeocline indicates a greater thickness of strata accumulated per unit time in the Early Cambrian, and presumably in the late Precambrian, than for younger rocks (Fig. 8). This plot is based on the White-Inyo Mountains section, where paleontological control on the position of the Precambrian-Cambrian boundary is best in the western United States. Even so, the location of this boundary is speculative, and thicknesses shown are based on two interpretations of the location of the contact. The thickness of the Devonian strata in the plot is based on extrapolation from nearby sections, because the upper part of the Devonian strata in the White-Inyo Mountains has been removed by pre-Mississippian erosion.

The greatest problem with the plot concerns the time scale, which is based on the Geological Society of London (1964) standard. For upper Precambrian and lower Paleozoic rocks, the standard is based on only a few data points and, therefore, might be significantly in error. Thus, the plot shown in Figure 8 is tentative, although the general conclusion of a decreasing rate of accumulation and thus decreasing subsidence with time is considered valid.

Sleep (1971) and Sleep and Snell (1976) have outlined a model for subsidence on Atlantic-type continental margins that appears to be applicable to the upper Precambrian and lower Paleozoic sequence of the western United States. Such margins are considered to have been created when a rift and

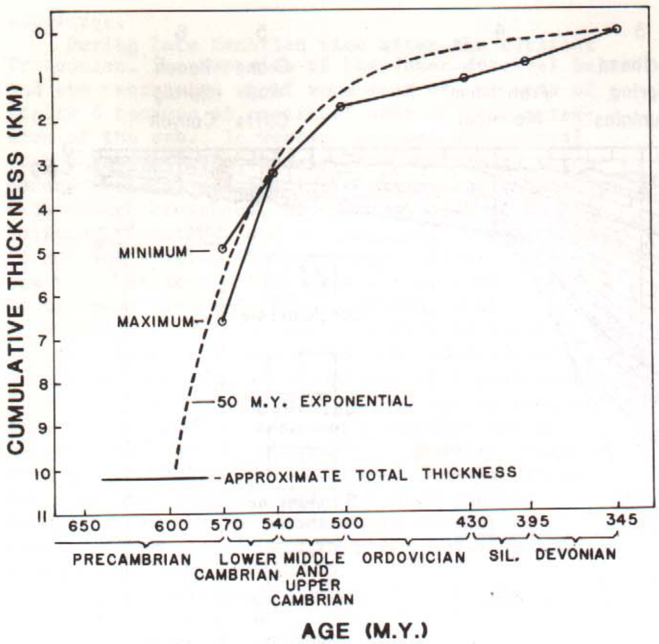

Figure 8. Cumulative thickness with time of upper Precambrian and lower Paleozoic rocks in the White-Inyo Mountains of eastern California.

spreading center develop within a continent and the fragments drift apart as a new ocean basin forms (Dewey and Bird, 1970). The Atlantic margin, for example, formed during the Mesozoic rifting and fragmentation of the Pangaean supercontinent (Dietz and Holden, 1970) when, as the result of the rift and a spreading center, North America drifted away from Europe and Africa as the newly formed Atlantic ocean basin opened. In such a system, mantle upwelling during the rifting and along the spreading center causes high heat flow that results in thermal expansion of the lithosphere and uplift at the surface. As the continent moves away from the source of the thermal anomaly at the spreading center, the lithosphere cools, contracts, and consequently subsides. The rates of subsidence in this system, as determined both for oceanic crust as it moves away from the spreading center and for the Cretaceous to Holocene sequence in the Atlantic and Gulf coasts of the United States, are similar and decline exponentially with a time constant of about 50 m.y. Subsidence follows the equation $u = u_o \exp(-at)$, where (u_o) is the initial uplift, (u) is the elevation at time (t) after the initial uplift, and (a) is a constant. The time constant is equal to $1/a$.

A similar exponential rate of subsidence is probable for the upper Precambrian and lower Paleozoic rocks of the western United States (compare observed subsidence rate with 50-m.y.-exponential curve in Fig. 8). The relations shown on Figure 8 indicate that the oldest exposed sedimentary rocks in the White-Inyo Mountains may have been deposited about 600 m.y. ago, provided that the model of subsidence following a thermal uplift is used. Even if the original thickness of the Precambrian rocks was much greater than that now preserved and if the 50-m.y.-exponential curve is recalculated to fit a minimum rate of subsidence, the time of initial late Precambrian sedimentation in the White-Inyo Mountains probably could be no more than about 650 m.y. ago.

TECTONIC MODEL

The tectonic model proposed here is built on the hypothesis proposed by Stewart (1972, 1976) that late Precambrian rifting approximately 850 m.y. ago created a new, or at least a reshaped, continental margin along western North America. According to this concept, the Belt Supergroup and temporally equivalent rocks in the western United States formed prior to this new margin and are related to a different tectonic style. This concept does not preclude the possibility of a complex history for the Belt and temporally equivalent rocks, perhaps involving both extensional and compressional events. The main point is that a significant change occurred in the tectonic setting of the western United States after deposition of the Belt Supergroup and temporally equivalent rocks and that this change is similar to that which occurred during well studied rifting events such as that which formed the eastern margin of North America in the Mesozoic.

Some geologists, on the other hand, have proposed that the main rifting event in the western United States was prior to the deposition of the Belt Supergroup and related rocks (Monger and others, 1972; Burke and Dewey, 1973), that an ocean margin existed along the western edge of North America during deposition of the Belt and related rocks, and that this ocean margin was in approxi-

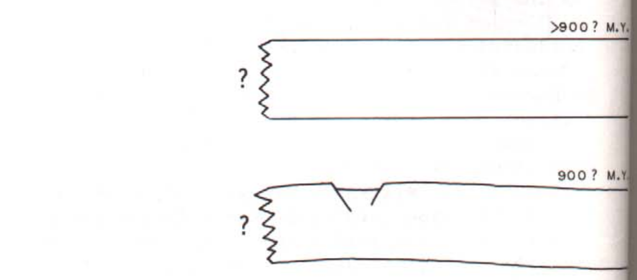

Figure 9. Diagram showing a model of the late Precambrian and Cambrian development of the western United States.

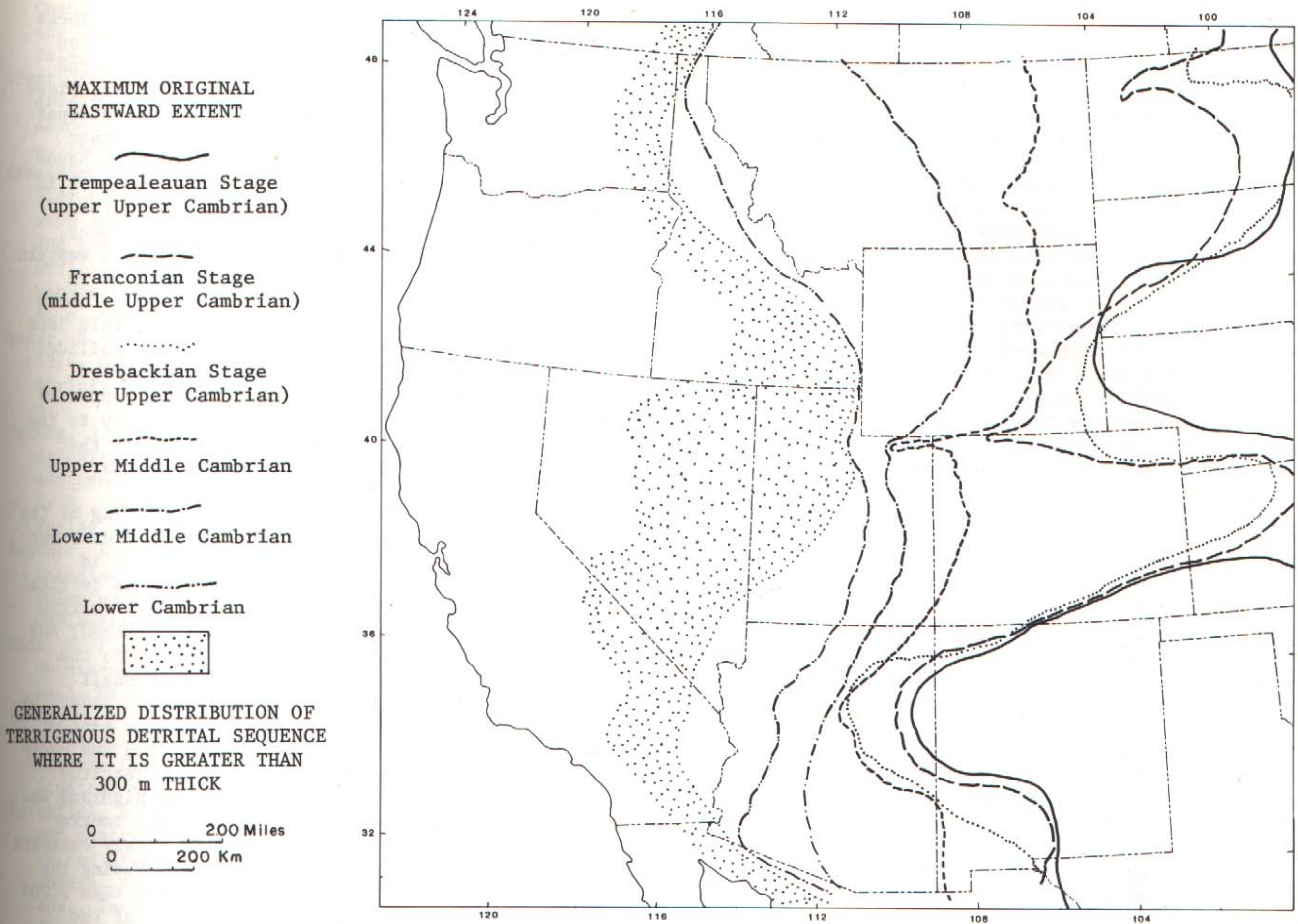

Figure 10. Distribution of miogeoclinal upper Precambrian and Lower Cambrian terrigenous detrital sequence and maximum original eastward extent of Middle and Upper Cambrian stages in the western United States (in part after Lochman-Balk, 1972).

mately the same position as it was for later Cordilleran miogeoclinal deposits (Gabrielse, 1972; Monger and others, 1972; Seyfert and Sirkin, 1973; Harrison and others, 1974; Harrison and Reynolds, 1976; King, 1976). The relative merits of the pre-Belt and post-Belt rifting hypotheses have been discussed previously by Gabrielse (1972), Monger, Southes, and Gabrielse (1972), Burchfiel and Davis (1975), and Stewart (1976) and will not be reviewed here. Some geologists (Burchfiel and Davis, 1975; Dickinson, 1977) have discussed the possibility that rifting has occurred in both pre-Belt and post-Belt times, thus incorporating elements of both hypotheses.

The model proposed here (Fig. 9) differs slightly from that proposed by Stewart (1972, 1976) in that it incorporates the concept that the diamictite and volcanic sequence may have accumulated locally in rift-valley basins that formed prior to the main stage of rifting. The timing of these events is poorly known, but the rift valleys may have formed from 800 to 900 m.y. ago and the main rifting event not until perhaps 650 m.y. ago. Knut Bjørlykke (oral commun., 1976) has noted that many rift valleys, such as the Oslo graben in Norway, formed at a significantly earlier time than the main continental rifting. He suggests that these rift valleys formed along lines of weakness that were later to become the main site of continental rifting.

In the model proposed, the terrigenous detrital sequence is related to erosion of the initial rift bulge (uplifted area) that was created by thermal expansion during the continental rifting. Such uplift is well known along the Afro-Arabian rift system (Baker and others, 1972; Kinsman, 1975; Lowell and others, 1975). The cessation of deposition of terrigenous detritus in the miogeocline is related to the destruction of the bulge by thermal contraction due to cooling and by erosion. After the bulge had been lowered to a certain level, the ocean transgressed eastward (Fig. 10) into cratonic areas. As a consequence, coarse detrital material was trapped in coastal areas on the craton and the miogeoclinal shelf was an area of carbonate deposition relatively free of detritus.

PALEOGEOGRAPHY

The dominant paleogeographic pattern in the western United States during the late Precambrian and Cambrian was a broad shallow-water shelf extending along the western margin of the North American craton. Land areas lay to the east and a deep-water ocean basin to the west. The shelf sediments are miogeoclinal deposits similar to those that accumulated during the Cretaceous and Cenozoic along the eastern margin of North America (Stewart and Poole, 1974). Most of the preserved upper Precambrian and Cambrian rocks in the western United States are shallow-water sediments that comprise the miogeocline. The eugeosynclinal Scott Canyon Formation, which may be only partly Cambrian in age, includes abundant chert and greenstone and is interpreted to be a deep-water ocean basin deposit formed west of the shelf and tectonically emplaced over the shelf sediments presumably during the mid-Paleozoic Antler orogeny. Some Middle and Upper Cambrian units contain abundant carbonate slump breccia and relatively deep water trilobites (Taylor and Cook, 1976) and probably accumulated on the upper part of the continental slope between the shelf to the east and the lower slope and upper continental rise to the west. Middle and Upper Cambrian units generally are shaly and cherty near the western margin of the miogeocline, and these units may also have accumulated near the shelf edge or on the upper part of the continental slope.

Cambrian paleogeography west of the continental slope and rise is unknown. The oldest rocks to the west are Ordovician in the northern Sierra Nevada, Ordovician in the Klamath Mountains, Devonian in central and eastern Oregon and Ordovician in northwestern Washington and adjacent parts of British Columbia. The oldest Ordovician rocks in the Klamath Mountains and northwest Washington are ophiolite complexes (Hopson and Mattinson, 1973), which clearly indicate a post-Cambrian origin of most of this westernmost region of the United States.

In east-central Idaho, an unusual feature of the Cambrian paleogeography is a topographically high area, called the Lemhi arch (Figs. 4-6) by Sloss (1954) and the Salmon River arch by Armstrong (1975) that separates the Cordilleran miogeocline from the shelf to the east. Within this arch, Ordovician rocks locally rest unconformably on rocks temporally equivalent to the Belt Supergroup (Beutner and Scholten, 1967; Ruppel and others, 1975; Ruppel, 1976). Precambrian crystalline basement rocks presumably were exposed in the arch during Cambrian time and are a probable source for some feldspar-rich detrital units in north-central Idaho and perhaps elsewhere.

The source of the feldspathic detrital material in the Harmony Formation (Upper Cambrian) of north-central Nevada is conjectural. All other units of comparable age in Nevada are composed predominantly of carbonate and shale. In northern Utah and southeastern Idaho, however, shelf rocks contain a feldspar-rich quartzite unit, the Worm Creek Quartzite Member of the St. Charles Limestone (Haynie, 1957; Armstrong and Oriel, 1965; Trimble and Carr, 1976). This unit is Late Cambrian, the age of the Harmony. In east-central Idaho, the Clayton Mine Quartzite, which is of uncertain age but lies below rocks of Middle Ordovician age and apparently above structurally separated rocks of Middle Cambrian age, is also feldspathic (Hobbs and others, 1968). Thus, the sandstone of the Harmony possibly was derived from the same source area as the Worm Creek and Clayton Mine quartzites. Such a source might have been Precambrian rocks exposed in the Lemhi (Salmon River) arch in east-central Idaho or on the craton to the east of Idaho or northern Utah. If so, detrital material may have been transported across the shelf in Idaho and from there southward in deeper water west of the shelf into north-central Nevada. Alternatively, the sand of the Harmony could have been derived from rocks of the uppermost Precambrian and Lower Cambrian terrigenous detrital sequence exposed along the continental slope, although in general these older detrital rocks do not contain sufficient feldspar to be a source for the Harmony. Still another possibility is that Precambrian crystalline basement rocks exposed on the continental slope or in some area of local uplift (Erickson and Marsh, 1974) could have been a source for the Harmony. Finally, the Harmony could have come from a western source area separated from North America by an ocean basin or a marginal sea. The Harmony would have been emplaced on North America when this land (perhaps a microcontinent or island arc) collided with North America. The concept of a foreign source for the Harmony, however, is not supported by the affinity of the fauna in the Harmony to the North American faunal province, a relation that suggests that the Harmony was deposited near the North American shallow shelf.

In summary, the paleogeographic setting of the diamictite and volcanic sequence is not clear, but this sequence may have accumulated in part in rift valleys. By the time the terrigenous detrital sequence was deposited, however, the shelf margin of the western United States was well defined, and this paleogeographic pattern continued into the Devonian. If the establishment of this shelf margin was related to rifting, a highland caused by thermal expansion probably existed to the east of the miogeocline in the latest Precambrian and Early Cambrian time and was a source of much detritus. By the end of Early Cambrian time, the highland was largely destroyed by erosion and thermal contraction, allowing the sea to transgress slowly eastward over areas that were previously land. During the Late Cambrian, seas extended 600 to more than 1,000 km (Fig. 10) east of the eastern margin of the miogeocline, and much of the western United States was a shallow sea.

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