Geologic Excursions in Stratigraphy and Tectonics: From Southeastern Idaho to the Southern Inyo Mountains, California, via Canyonlands and Arches National Parks, Utah

GUIDEBOOK — PART II

The Geological Society of America
Rocky Mountain and Cordilleran Sections Meeting
Salt Lake City, Utah
May 2 - 4, 1983

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UTAH GEOLOGICAL AND MINERAL SURVEY
a division of
Utah Department of Natural Resources and Energy

Special Studies 60 **May 1983**

GEOLOGIC EXCURSIONS IN STRATIGRAPHY AND TECTONICS

Contents

Editor's Note: The papers contained in this Guidebook were solicited by the organizers of the GSA Rocky Mountain and Cordilleran Sections and have been edited and given a common format; however, their style and content have not been formally reviewed by the Utah Geological and Mineral Survey.

Utah Geological and Mineral Survey Special Studies 60, 1983
Guidebook Part 2 — GSA Rocky Mountain and Cordilleran Sections Meeting
Salt Lake City, Utah — May 2 - 4, 1983

UPPER PROTEROZOIC DIAMICTITES AND VOLCANIC ROCKS OF THE POCATELLO FORMATION AND CORRELATIVE UNITS, SOUTHEASTERN IDAHO AND NORTHERN UTAH

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REVIEW OF GEOLOGY OF THE UPPER PROTEROZOIC POCATELLO FORMATION

ABSTRACT

Lying at the base of the exposed Upper Proterozoic strata of the Bannock Range, Idaho-Wyoming thrust belt, the Pocatello Formation contains rocks that record volcanism and glaciomarine sedimentation in a newly rifted ocean basin. The Scout Mountain Member composes the bulk of the Pocatello Formation and contains strata which become more texturally mature upward. Two diamictite horizons occur in the Scout Mountain Member in the northern Bannock Range, with the upper one interpreted as a subaqueous lodgement tillite. The lower diamictite gradationally overlies fragmental mafic volcanic rocks of the Bannock Volcanic Member and is interpreted as a subaqueous mass flow deposit in a tectonically active environment. Other lithologies in the Bannock Volcanic Member suggest extrusions and subsequent fragmentation of lava in shallow water. The top of the Scout Mountain Member and overlying upper member, Pocatello Formation contains a fining upward (inferred transgressive) sequence that may record a eustatic sea level rise coincident with melting of glacial ice and waning of Late Proterozoic glaciation from Alaska south to Death Valley.

INTRODUCTION AND REGIONAL SETTING

The Pocatello Formation is the oldest rock unit exposed in the Idaho-Wyoming thrust belt, and occurs in the Bannock Range from the Snake River Plain near Pocatello south to the Utah border west of Preston (Figure 1). It occupies a similar stratigraphic position and is lithologically comparable to the formation of Perry Canyon, the Mineral Fork Formation, and the Sheeprock Group of northern and western Utah (Crittenden and others, 1971, 1983; Christie-Blick, 1982).

The Pocatello Formation and homotaxial units to the south contain mafic volcanic rocks and diamictite, which are unique among strata of the Cordilleran miogeocline (Stewart, 1972, 1982; Stewart and Suczek, 1977). The mafic volcanics are assigned to the Bannock Volcanic Member of the Pocatello Formation and contain pillow lava, porphyritic basalt, volcaniclastic diamictite breccia, intrusive rocks, and much foliated greenstone (Trimble, 1976; Link, 1982). The diamictite belongs to the Scout Mountain Member, Pocatello Formation, and contains pebble to boulder-size clasts of a variety of rock types, floating in a sandy to muddy matrix (Anderson, 1928; Ludlum, 1942, 1943; Crittenden and others, 1971; Trimble, 1976; Link, 1981, 1982, 1983). Other strata of the Scout Mountain Member consist of dirty sandstone, siltstone, conglomerate, quartzite, limestone, and dolomite. The generally accepted model for deposition of the Pocatello Formation was developed by Stewart (1972, 1982; Stewart and Suczek, 1977) and involves continental rifting with associated mafic volcanism and coarse clastic sedimentation. The non-volcanic diamictites in Idaho and Utah have long been thought to be glacial tillite or flow tillite, as they can contain large and heterolithologic clasts which are occasionally striated, as well as sparse dropstones and till pellets (Blackwelder, 1910, 1925, 1932; Hintze, 1913; Ludlum, 1942, 1943; Calkins and Butler, 1943; Crittenden and others, 1952, 1971, 1983; Varney, 1976; Blick,

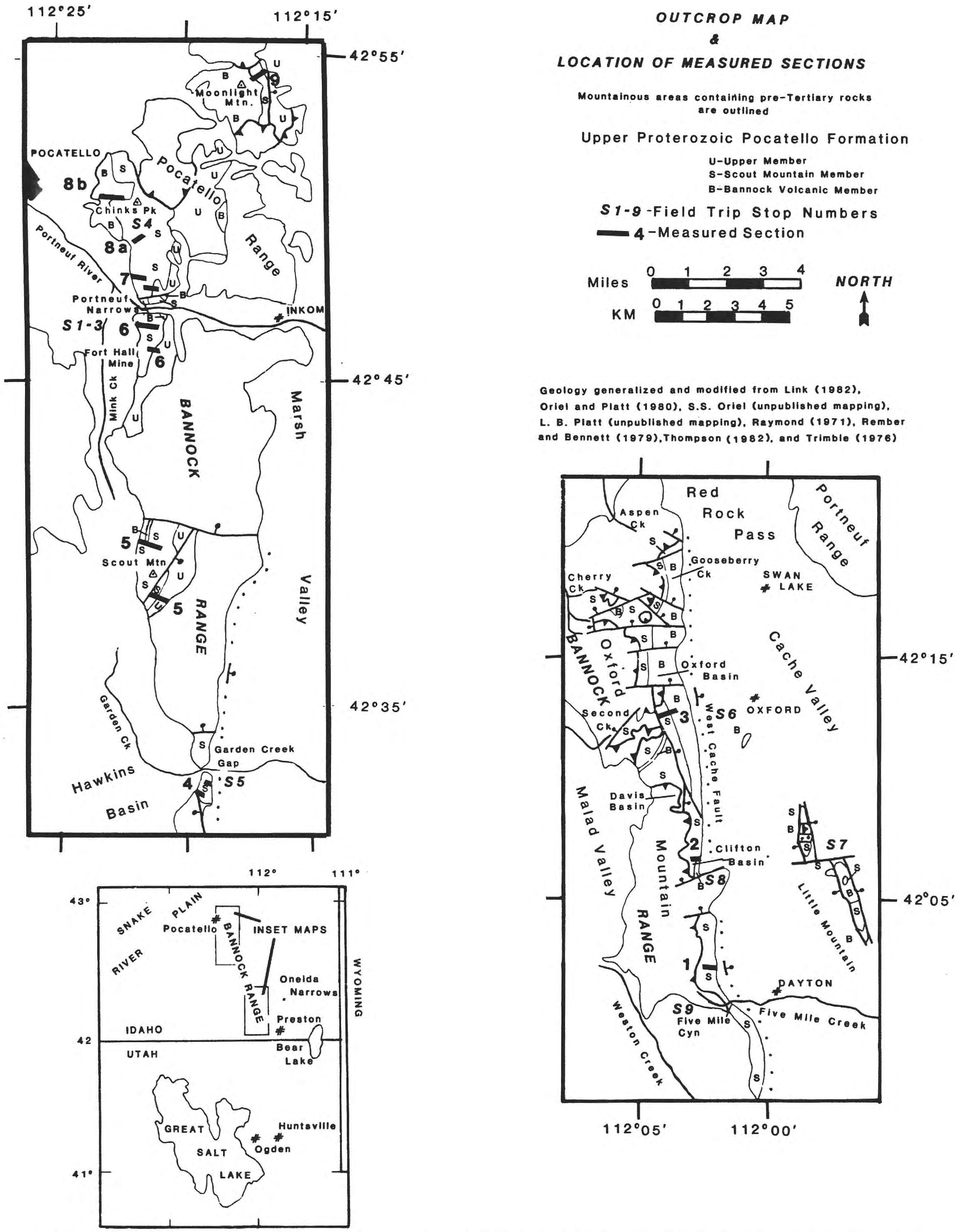

Figure 1. Generalized outcrop map of the Pocatello Formation. Locations of measured sections are shown. Sections are compiled in Figure 3.

1979; Ojakangas and Matsch, 1980; Link, 1981, 1982, 1983). The Utah diamictites and the Mineral Fork Formation in particular, contain more abundant glacial indicators than the Scout Mountain Member. A minority viewpoint (Condie, 1967; Schermerhorn, 1974) stresses the active tectonic setting and paucity of definitive glacial criteria and proposes that the diamictites are non-glacial subaqueous mass flows within a turbidite environment. Clasts in diamictite are both extrabasinal (quartzites, gneisses, and schists) and intrabasinal (siltstones and volcanics).

The association of diamictite and mafic volcanic rocks near the base of strata of the Cordilleran miogeocline from California to Alaska has been emphasized by Stewart (1972), Crittenden and others (1971), and Cristie-Blick and others (1980). This lithostratigraphic similarity is the main basis for correlating the Pocatello Formation with those upper Proterzoic strata. Diamictites occur in upper Proterozoic sequences worldwide (Harland, 1964; Crowell and Frakes, 1970; Crawford and Daily, 1971; Schermerhorn, 1974; John, 1979; Frakes, 1979; Hambrey and Harland, 1981).

Although no radiometric dates have been obtained from the Pocatello Formation, it is generally thought to be Late Proterzoic (900 to 570 m.y.) in age (Harrison and Peterman, 1980). This assignment is based on its stratigraphic position conformably below a siltstone and quartzite sequence ("Brigham Group")[1] which contains middle Cambrian fossils near the top (Figure 2), and also on homotaxial correlation with similar strata in northeastern Washington. There, a diamictite-bearing unit, the Shedroof Conglomerate, underlies volcanic greenstone dated at 827 to 918 m.y. by K-Ar analysis of plagioclase, pyroxene, and whole rock samples (Miller and others, 1973). A somewhat younger age (less than 770 m.y.) can be inferred for similar diamictite-bearing strata of the Windermere Supergoup in the Mackenzie Mountains of the northern Canadian Cordillera (Armstrong and others, 1982), based on Rb-Sr isochrons obtained from diabase that intrudes strata of the underlying Mackenzie Mountains Supergroup.

In a review paper correlating diamictite units in northern Utah and southeastern Idaho, Crittenden and others (1983) describe two diamictite intervals separated by up to 1,000 m of non-glacial deposits in allochthonous parts of the northern Utah thrust belt. These diamictites are interpreted to represent separate glacial episodes. The diamictites of the Scout Mountain Member and the upper diamictite horizon in northern Utah are correlated and inferred to represent the second glaciation because they are intercalated with and overlie mafic volcanic rocks. In the Utah autochthon only one diamictite is present and is also correlated with this second glaciation.

STRATIGRAPHY AND PETROLOGY OF THE POCATELLO FORMATION

The Pocatello Formation, metamorphosed to greenschist facies, crops out in three structural blocks (Figure 1): 1) on Oxford Mountain of the southern Bannock Range, 2) from Garden Creek Gap north to Portneuf Narrows in the northern Bannock Range, and 3) on Chinks Peak and Moonlight Mountain in the northernmost Bannock (Pocatello) Range, north of Portneuf Narrows. A simplified composite stratigraphic section is shown in Figure 2. It must be emphasized that the Pocatello Formation is a metamorphic rock unit that is discontinuously exposed in a structurally complex area, which has been affected by both Mesozoic thrusting of the Sevier orogenic belt and by Tertiary basin and range volcanism and complex normal faulting.

The Pocatello Formation is now recognized to contain three members (Figures 2 and 3). The originally defined phyllitic lower member (Crittenden and others, 1971, p. 583; Trimble, 1976, p. 13) occurs in only one locality, north of Portneuf Narrows (measured section 7, Figures 1 and 3). It has since been shown to be structurally overturned (Link and others, 1980), and is now recognized to be instead the upper member, which consists of gray to black phyllitic argillite and shale, at least 600 m thick (Trimble, 1967, p. 127). The upper member reaches its highest metamorphic grade in the small canyon north of Portneuf Narrows where it was originally thought to be the "lower member."

The Scout Mountain Member consists of at least 900 m of detrital and carbonate strata and comprises the lowest exposed beds in the Pocatello area. The base is not exposed. The Bannock Volcanic Member is lenticular within the Scout Mountain

[1]The term "Brigham Group" refers to strata assigned to the Brigham Quartzite in the northern Bannock Range by Anderson (1928); in the Portneuf Range of southeastern Idaho by Oriel and Armstrong (1971) and Lindsey (1982) (Figure 2). The quotation marks are to alert the reader that ""Brigham Group"" as used in this sense in southeastern Idaho contains slightly different formations than the "Brigham Group" as defined in northern Utah (Walcott, 1908; Crittenden and others, 1971; Sorensen and Crittenden, 1976a).

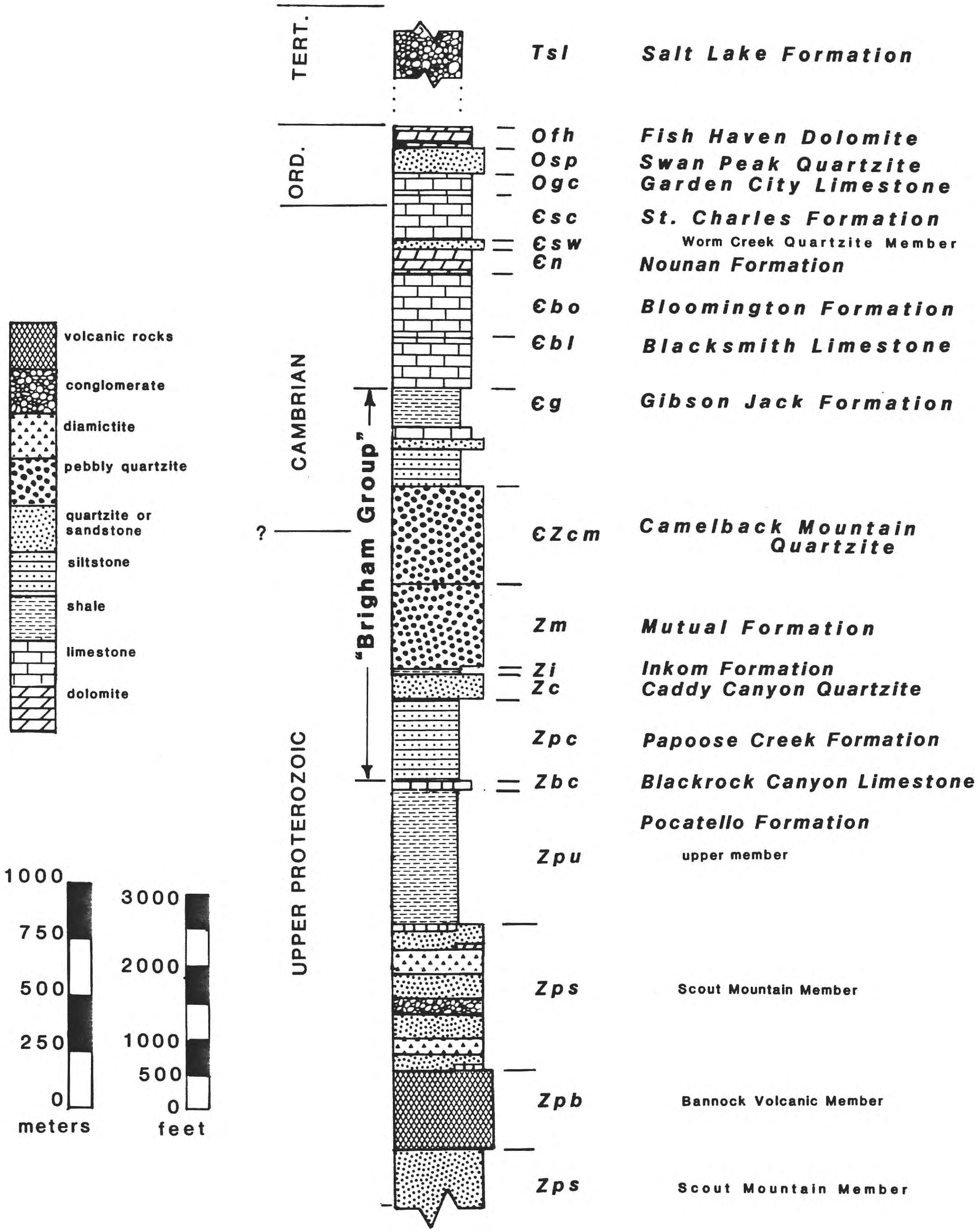

Figure 2. Generalized stratigraphic section for the Bannock Range. Thicknesses are those characteristic of Oxford Mountain in the southern Bannock Range.

Member and comprises up to 400 m of green, mafic, plagioclase-porphyritic lava, pillow lava, intrusive rocks and volcaniclastic diamictite breccia. Volcaniclastic diamictites containing recrystallized chloritic fragments in a fine-grained green or gray matrix, and assigned to the Bannock Volcanic Member, grade into diamictites containing both volcanic and quartzite clasts. Placing a contact between the two units is sometimes arbitrary. This is particularly true immediately north and south of Portneuf Narrows. Diamictite containing any terrigenous quartz grains or extrabasinal stones belongs to the Scout Mountain Member by definition, but detection of the isolated quartzite cobble in a sea of greenstone breccia is frustrating and often futile. An origin for some of these quartz-free volcaniclastic rocks as pillow breccia (deWit and Stearn, 1978) has been proposed by G. D. Harper (oral communication, 1982). Pillow breccias form when basaltic pillow lava is broken up and redeposited by waves or other ocean currents.

As documented by Link (1982, 1983, Figure 3, this paper), the Scout Mountain Member of the Pocatello Formation contains an anomalous and possibly faulted pebbly quartzite at the base, overlain by a lower siltstone, a lower diamictite unit, a sequence of lithic wackes containing graded beds, siltstones with local limestones, a quartzite-cobble conglomerate, more sandstones, an upper diamictite, a horizon of dolomite and dolomite-chip breccia, and a fining-upward sequence of sandstone, siltstone, and limestone. This latter sequence is interpreted as being transgressive. It is overlain by dark shale of the upper member, Pocatello Formation.

Except for the lower quartzite unit, there is a decrease in percent of lithic fragments and increase in textural maturity upward in sandstones of the Scout Mountain Member. With a few exceptions, the principal lithic fragments are siltstone and volcanic rocks in the lower part of the section. Orthoquartzite and granitoid plutonic and metamorphic rocks become prevalent above the cobble conglomerate. Glacial erosion of the land surface probably contributed to this change from intrabasinal siltstone to extrabasinal quartzite-dominated lithic assemblages near the horizons of the cobble conglomerate and upper diamictite units. Sandstones near the top of the member are subarkosic and quartz wackes and subarkosic arenites (terminology of Folk, 1980). The orthoclase to plagioclase feldspar ratio increases upward through the Scout Mountain Member.

"Brigham Group" sandstones above the Pocatello Formation are chiefly quartz arenites and subarkosic arenites (Lindsey, 1982). The principal lithic fragments in "Brigham Group" sandstones are polycrystalline quartzite derived from basement rocks and orthoquartzite. These trends in sandstone petrology are thought to represent unroofing of Archean basement of the North American continent following rifting.

SEDIMENTARY FACIES AND DEPOSITIONAL ENVIRONMENT

Interplay of basaltic volcanism, active faulting, and glaciation is recorded in the strata of the Pocatello Formation. Anderson (1928) first recognized the volcanic component, while Ludlum (1942) proposed that glaciation and syndepositional faulting occurred. Recent studies by Stewart (1972), Trimble (1976) Crittenden and others (1971), and Link (1982; 1983) have emphasized glaciomarine sedimentation in a newly rifted ocean basin floored with continental crust.

The lower quartzite unit of the Scout Mountain Member, which only occurs west of Scout Mountain (measured section 5 on Figures 1 and 3) was probably deposited in a shallow marine setting. This unit is poorly exposed and may be faulted into place. If it can be proven to lie depositionally below lithic sandstones of the main part of the Scout Mountain Member, the shallow water sedimentation it records must have occurred before rifting, volcanism, and uplift of nearby source areas. Thus this unit could have been deposited under tectonically quiet conditions similar to those prevailing during deposition of the middle Proterzoic Belt Supergroup to the north in Idaho and Montana, and the Uinta Mountain Group and Big Cottonwood Formation to the south in Utah. It may even be a partial age equivalent of these units, although this probably cannot be proven by radiometric dating.

The lowest exposed rocks in the Chinks Peak area and in the southern Bannock Range are graded sandstone and the lower diamictite (Scout Mountain Member), pillow lava, and breccia (Bannock Volcanic Member). These strata record volcanism and tectonically influenced sedimentation in what, because of lateral continuity, is interpreted to have been a marine setting. A lacustrine setting is also possible. The volcanic rocks may have been deposited in shallow water, in which the pillow lavas were reworked. Some of the sandstones are turbidites. Nearby source areas containing siltstone and volcanic rocks (lateral equivalents of the Scout Mountain Member) were uplifted and supplied sediment to

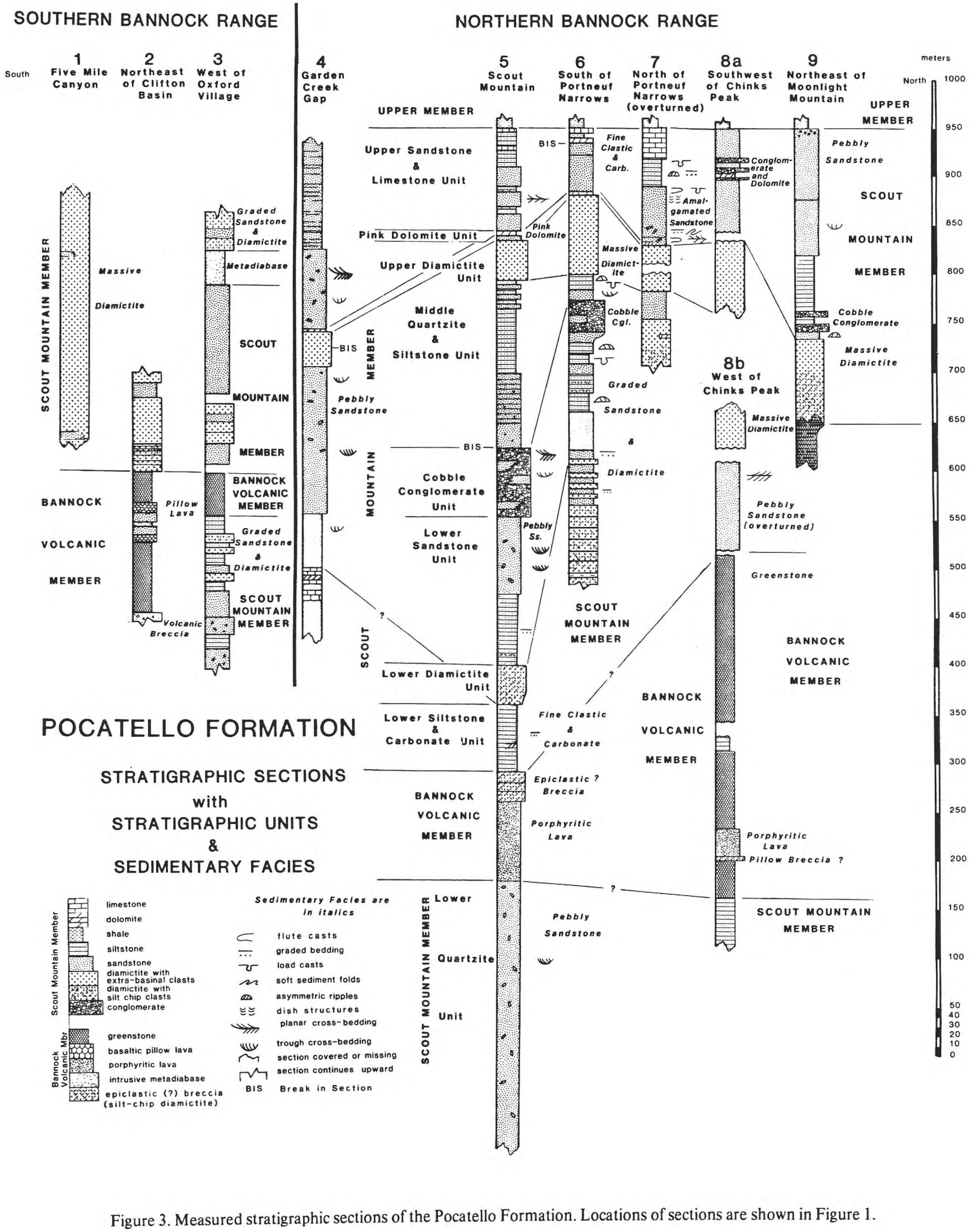

Figure 3. Measured stratigraphic sections of the Pocatello Formation. Locations of sections are shown in Figure 1.

what was probably a deepening basin. Inferred glacial sediment input of rounded extrabasinal cobbles and boulders was present, though minor, and the lower diamictite was probably depostied by subaqueous mass flows in a turbidite environment rather than directly from glacial ice. Above the lower diamictite is a non-glacial turbidite sequence (the lower sandstone unit).

The cobble conglomerate, upper diamictite, and associated sandstones were probably derived chiefly from glacially transported debris. They can be interpreted to have been deposited in shallower water than the underlying graded sandstones, perhaps reflecting a drop of sea-level due to storage of water in glacial ice. The cobble conglomerate can be interpreted both as a shallow marine deltaic or beach deposit and as a submarine channel deposit fed by glacial meltwater streams laden with coarse debris. The upper diamictite appears massive, and shows no signs of having been reworked by currents. It is interpreted as subaqueous lodgement tillite deposited near the grounding line (line of buoyancy) of a large ice sheet that extended at least 35 km north to south in the northern Bannock Range and may have extended south to the Utah border or even farther (Crittenden and others, 1983).

Pink silty dolomite and dolomite-chip breccia occur above the upper diamictite in both Idaho and Utah (Crittenden and others, 1983). Dolomites overlie many upper Proterzoic diamictites worldwide (Schermerhorn, 1974). The dolomite in the Scout Mountain Member is wavy laminated and coarsely crystalline. It is most often found as clasts in a sedimentary breccia of dolomite flakes up to 50 cm long in a coarse subarkosic matrix. Breccia occurs in all sections except Garden Creek Gap (section 4, Figures 1 and 3).

Pleistocene and Holocene carbonates overlying tills have recently been documented by Bjorlykke and others (1978) and Parker and others (1981). Permian carbonates in glacial marine rocks of Tasmania are described by Rao (1981).

One interpretation of the dolomite involves local retreat of the ice in the northern Bannock Range followed by isostatic rebound. This would have allowed carbonate deposition in a shallow water or lacustrine setting. The breccias are interpreted as tidal channel or storm deposits. Dolomitization was presumably post-depositional.

General sea level rise caused by regional or even worldwide melting of ice sheets is recorded in the fining upward sequence at the top of the Scout Mountain Member and overlying shale of the upper member. This transgression appears to be represented from Death Valley to Alaska and, if the result of glacial eustacy, is useful as a time line (Christie-Blick and others, 1980; Young, 1982).

SOUTHERN BANNOCK RANGE

The stratigraphically lowest exposures in the southern Bannock Range (measured sections 1 to 3 in Figures 1 and 3) contain Bannock Volcanics and graded sandstone with thin diamictites at the base. These rocks are overlain by a massive diamictite which is correlated with the upper diamictite in the northern Bannock Range. The upper part of the section is eliminated by a younger-on-older fault. Because of their association with volcanic rocks, all diamictites of the Pocatello Formation are correlated with the upper diamictite of the formation of Perry Canyon southeast of Brigham City, Utah (Crittenden and others, 1983).

REMAINING PROBLEMS

The stratigraphy and general depositional framework of the Pocatello Formation have been established. Because of limited exposure and complex deformation, some conclusions of Link (1982; 1983) are highly inferential. Problems for future research include:

1. Provenance of Extrabasinal Clasts—The vast majority of clasts in diamictite of the Scout Mountain Member are pink, white, grey, and black orthoquartzites. Similar rocks occur in three stratigraphic units. The first is the Lemhi Group and Salmon Supergroup of the Belt miogeocline in central Idaho (Ruppel, 1975). Second, these lithologies are present in Cretaceous and Paleocene conglomerates in Jackson Hole, Wyoming, which were derived from an inferred Proterozoic quartzite terrane in eastern Idaho named the Targhee Uplift (Love, 1973). Third, similar quartzites occur in fault slices above archean gneiss in the Albion Range of central southern Idaho (Armstrong, 1968; Compton and others, 1977; Crittenden, 1979; Compton and Todd, 1979; Miller, 1980, 1983). Middle Proterozoic quartzites of the Big Cottonwood Formation and Uinta Mountain Group in Utah (Crittenden and Wallace, 1973) are less vitreous and more poorly sorted than most clasts in the Pocatello Formation and have been inferred to not have been a source rock (Link, 1982).

No detailed comparison of quartzite or granitoid clasts in the Scout Mountain Member with rock units in other areas has been attempted. Without proof of the provenance of extrabasinal clasts, paleogeographic models of the Pocatello Formation will remain speculative.

2. Stratigraphy, Petrology, Geochemistry, and Geochronology of the Bannock Volcanic Member—Structural complications and generally poor outcrop have precluded detailed study of the Bannock Volcanic Member in the Pocatello area. Some of the structural complexity has been resolved by LeFebre (1983). Detailed study of the Bannock Volcanics, both in the field and by geochemical techniques, is now in order.

3. Stratigraphic Position of the Lower Quartzite Unit—The sole exposure of the lower quartzite unit of the Scout Mountain Member west of Scout Mountain is possibly faulted into place. Petrographically the lower quartzite unit is more similar to quartzites of the "Brigham Group" than to the Scout Mountain Member. If an original stratigraphic position at the base of the Pocatello Formation can be verified, the lower quartzites of the Belt Supergroup.

4. Structural problems in the northern Bannock Range. Mapping of the area near Chinks Peak is being compiled (LeFebre, 1983), but the Moonlight Mountain area to the north and the area south of Fort Hall Mine still warrant detailed study.

ACKNOWLEDGEMENTS

I became familiar with the Pocatello Formation in the course of my Ph.D. dissertation research at the University of California, Santa Barbara. The study was supervised by John C. Crowell and Steven S. Oriel, and to them I owe many thanks. The research was supported by National Science Foundation Grant EAR77-06008 (John C. Crowell, principal investigator). Base maps, aerial photographs, and field equipment were provided by the U.S. Geological Survey through Steven S. Oriel and M. H. Hait Jr. The preparation of this field guide was supported by Faculty Research Grant No. 509 from Idaho State University.

I also wish to acknowledge discussions and field trips with Robert C. Bright, Nicholas Christie-Blick, Max D. Crittenden Jr., Eugene W. Domack, Victor Gostin, Michael Green, G. D. Harper, Dennis D. Kurtz, George B. LeFebre, Julia M. G. Miller, H. Thomas Ore, Lucian B. Platt, John H. Stewart, Braden J. Thompson, Donald E. Trimble, and Marjorie H. A. Ulin.

Alex Fischer and Vicky Zimmerman provided invaluable help in drafting and typing. H. T. Ore, G. B. LeFebre, and W. R. Hackett offered valuable comments on an earlier draft of this paper.

ROAD LOG FOR UPPER PROTEROZOIC DIAMICTITE-BEARING STRATA, POCATELLO, IDAHO TO BRIGHAM CITY, UTAH

This two day field trip will examine the Pocatello Formation in the Bannock Range of southeastern Idaho and the formation of Perry Canyon in the northern Wasatch Mountains near Brigham City, Utah. Figure 4 is a map of the field trip route. On the first day we will take two strenuous climbs in the Pocatello area to examine stratigraphic sections of the Scout Mountain and Bannock Volcanic Members, Pocatello Formation. The second day will cover more territory and involve several stops in the Pocatello Formation in Marsh and Cache Valleys, Idaho, and the formation of Perry Canyon in Perry Canyon southeast of Brigham City.

FIELD TRIP ROAD LOG: FIRST DAY

Pocatello Area

Mileage		Description
Incremental	Cumulative	
0.0	0.0	Assemble at Quality Royale parking lot at Pocatello Creek Road and Interstate 15. Pocatello Creek is incised into loess-covered benches of Quaternary gravel. Turn left out of parking lot and head southeast on I-15 (turn left under freeway). Refer to Figure 5 for a generalized geologic map of the Pocatello area. U.S. Geological Survey Bulletin 1400 (Trimble, 1976) describes this area.
3.5	3.5	Road cut on the left is in gravels and tuffs of the Mio-Pliocene Starlight Formation. We are travelling southeast toward Portneuf Narrows (Figure 6), cut through the Bannock Range by the ancestral Portneuf River. Our first stop will be just north of the narrows on the west side of the ridge. It is probably that the early Pleistocene Bear River flowed north and west from near Soda

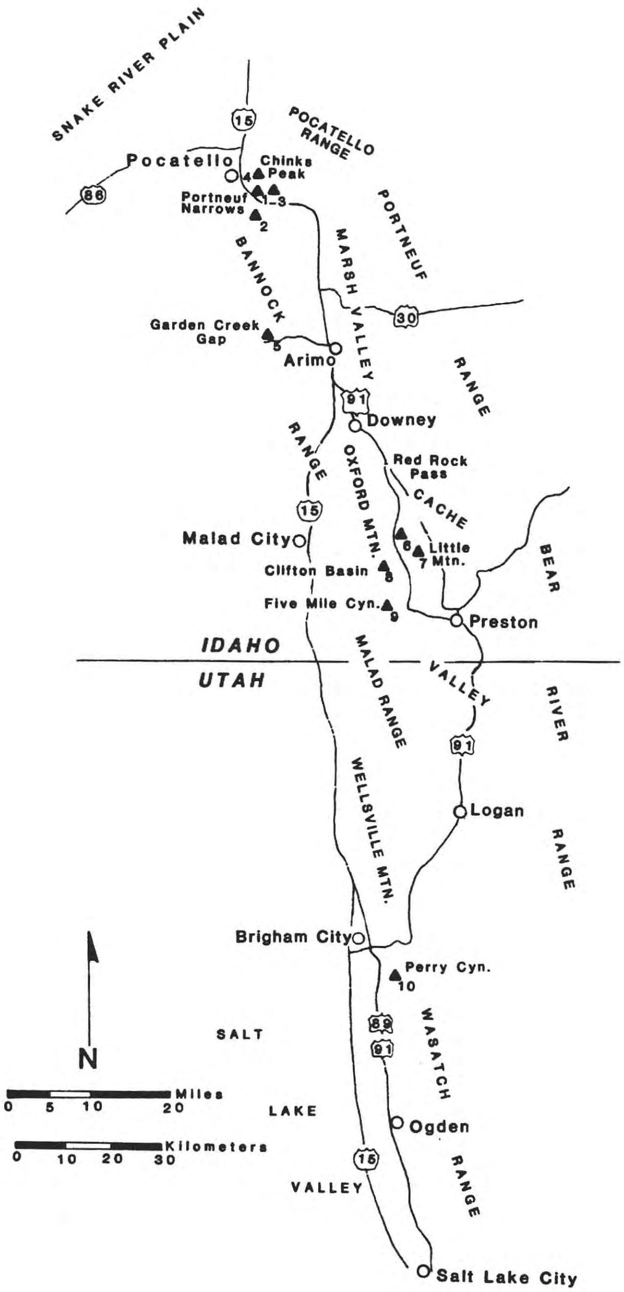

Figure 4. Stop locations for two day field trip to upper Proterozoic diamictites of southeastern Idaho and northern Utah.

Springs, through the north end of Gem Valley, and down the present course of the Portneuf River (Bright, 1963; Malde, 1968; Ore, 1983). It may have initially crossed the Bannock Range at Portneuf Narrows while flowing on Pliocene alluvial and tuffaceous sediment. Volcanism of the Gem Valley lava field probably dammed ancestral Bear River (Bright, 1960; 1963; Mabey and Oriel, 1970), and diverted it to the south into the Great Salt Lake. The modern Portneuf River is the superposed and beheaded remnant of the Pleistocene Bear River.

The two northern arms of the Bannock Range can been seen from the freeway. Kinport Peak (7,222 ft) is the highest point in the southwestern arm to the right. Kinport and nearby slopes are underlain by generally east-dipping quartzites and shales of the upper Proterzoic-Cambrian "Brigham Group" (Crittenden and others, 1971; Trimble, 1976; Sorensen and Crittenden, 1976a; see Figure 2).

The northeastern part of the Bannock Range (Figure 7, to the left) is also known as the Pocatello Range and is underlain mainly by the upper Proterozoic Pocatello Formation. Chinks Peak (6,791 ft) forms the rounded summit with radio towers on top. Figure 8 is a map of the Chinks Peak area. Stop 4 of the field trip is on the summit of Chinks Peak.

The main north-trending ridge of the Bannock Range extends south from Portneuf Narrows and merges with the trend of the Wasatch Range in Utah. All of our field trip stops will be along this trend. Scout Mountain (8,700 ft) is the peak to the south-southeast.

4.6 8.4 Take Portneuf Area exit 63 to the right, bear right onto frontage road. Bear right at stop sign and proceed under freeway.

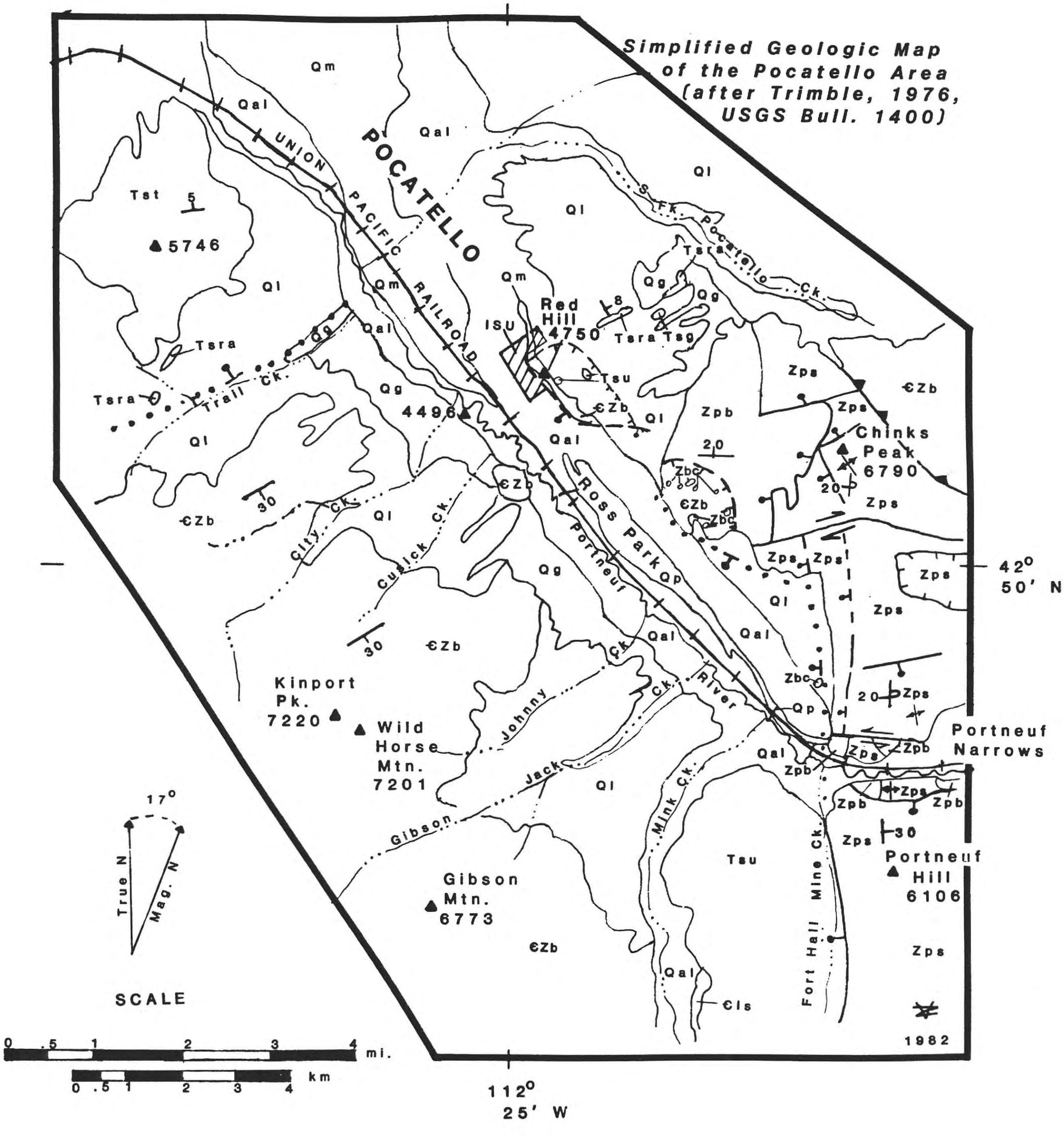

Figure 5. Generalized geologic map fo the Pocatello area. Stratigraphic units are as follows: Qal-Alluvium, Ql-Loess, Qm-Michaud Gravel (Lake Bonneville flood deposits), Qp-Portneuf Basalt lava flow, Qg-Gravels, Tsu-Starlight Formation, gravels, Tst-Starlight Formation, trachyandesite lava, Cls-Cambrian limestones, CZb-"Brigham Group," mostly quartzites, Zps-Pocatello Formation, Scout Mountain Member, Zpb-Pocatello Formation, Bannock Volcanic Member.

Figure 6. View looking east to Portneuf Narrows from south of Ross Park. Portneuf Hill is the high point south of the Narrows. Field trip STOP 1 is indicated on left, immediately above the Portneuf Lava flow. Traverse of field trip STOP 2 is shown on right side of picture.

0.7 9.1 Pull off on right just beyond the freeway underpass. We are on the Inkom 7½-minute quadrangle. **STOP 1**. Bannock Volcanic Member north of Portneuf Narrows. Figure 9 is a geologic map of this area.

Several lithologies are present on the southwest flank of the ridge above the freeway. These include greenstone breccia, volcaniclastic sandstone and probably pillow lava. Up the ridge to the northeast, breccia of the Bannock Volcanic Member grades upward into silt-chip diamictite with sparse quartzite clasts of the lower diamictite unit, Scout Mountain Member.

Turn around and proceed back under freeway.

0.2 9.3 Bear right at stop sign across railroad tracks and Portneuf River.

0.2 9.5 Cross Portneuf Road (stop sign).

0.6 10.1 Bear left, avoiding road to sanitary landfill, and drive up Fort Hall Mine Canyon a hundred meters to locked gate. Park. We are on the Inkom 7½-minute quadrangle. Figure 10 shows the canyon we will ascend.

STOP 2. West slope of Portneuf Hill, east of Fort Hall

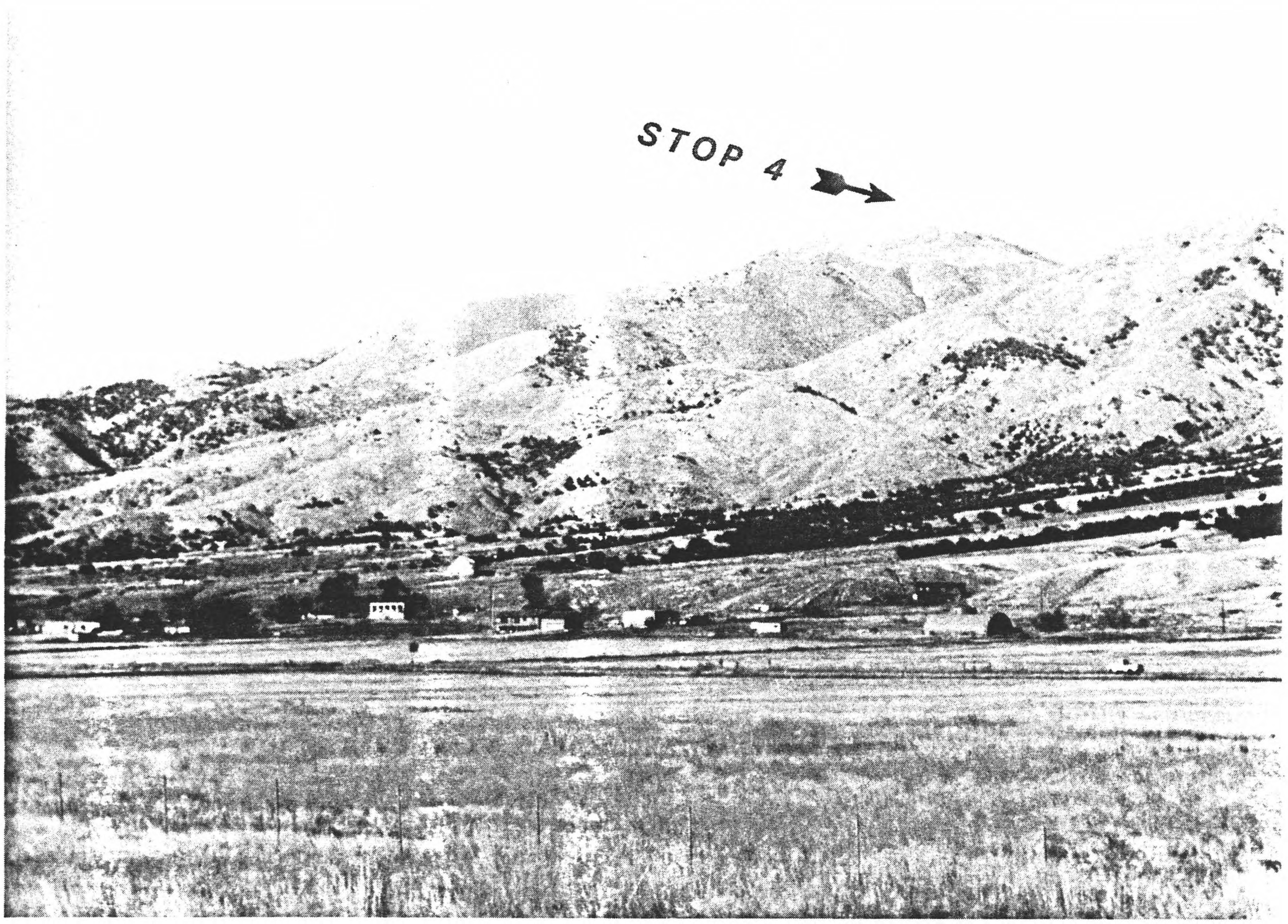

Figure 7. Chinks Peak looking north from I-15. The Scout Mountain Member makes up the summit and upper slopes. It is apparently overturned. Juniper-covered slopes in left background are underlain by the Bannock Volcanic Member. Location of field trip Stop 4 is shown at top of hill.

Mine Canyon. Much of the lower part of the Scout Mountain Member is exposed in the first and second canyons south of the locked gate. Figure 9 is a geologic map of this area.

The easiest access is up faint trails on the loess-covered alluvial fan surface south of the first canyon mouth. Drop south into the second canyon near the top of the alluvial bench, and proceed up-section to juniper-covered steep slopes, then cut north into the head of the first canyon and proceed up to the ridge summit. Figure 11 contains a measured stratigraphic section on this hillside. The section is condensed as section 6 of Figure 3.

The lowest exposed beds belong to the lower diamictite unit of the Scout Mountain Member, Pocatello Formation (1d on geologic maps, Figures 8 and 9). Near the head of the fan surface, schistose gray, green, and brown diamictite crops out among the junipers. Cleavage dips west (downslope) at about 30° and should not be confused with bedding, which dips northeast (into the slope) at about 50°. Bedding is difficult to find in this lower diamictite, but fine sandstone interbeds are present. Do not confuse thin

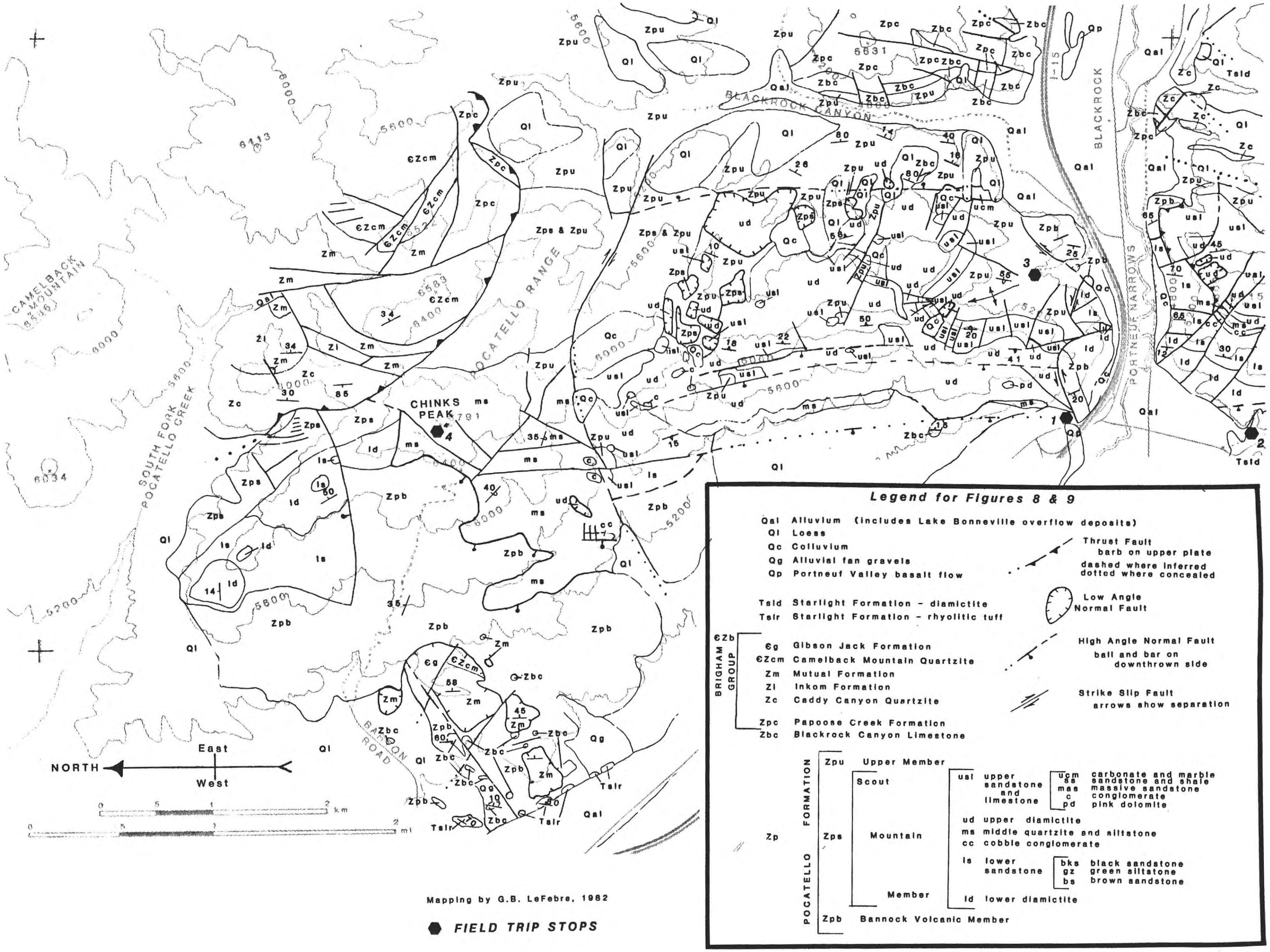

Figure 8. Geologic map of the Chinks Peak area, Pocatello Range. Note: NORTH IS TO THE LEFT. Stratigraphic units are shown on the legend. Mapping of the area east of Blackrock Canyon and of large klippe east of Chinks Peak are modified from Trimble (1976). Base map is from USGS Inkom and Pocatello South 7½-minute quadrangles.

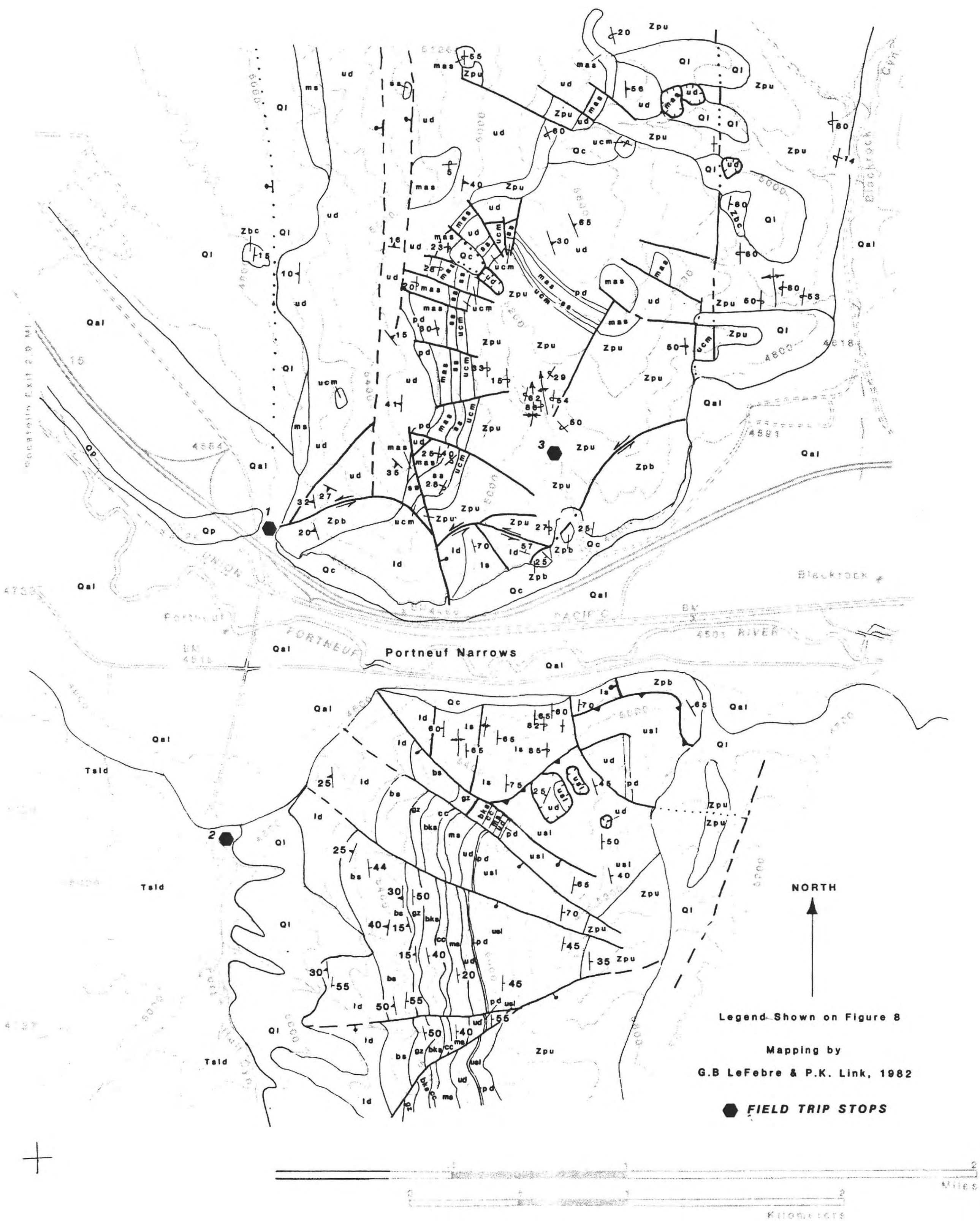

Figure 9. Geologic map of the Portneuf Narrows area. Legend is shown in Figure 8.

beds with liesegang bands! The clasts in this lower diamictite are dominantly recrystalized chloritic chips which were probably originally either pelitic rocks of the Scout Mountain Member or fine-grained volcanic rocks of the Bannock Volcanic Member that were ripped up and redeposited. Extrabasinal quartzite and granitic clasts are also present. Is there any hint of a glacial sediment input into this unit? What other mechanisms could provide the mix of clast types?

Northward toward Portneuf Narrows this lower diamictite becomes schistose and greener and appears to grade downward into volcaniclastic breccia of the Bannock Volcanic Member, Pocatello Formation. The diamictite is mapped as Scout Mountain Member if any extrabasinal quartzite pebbles are present. Often, however, finding the boundary is frustrating or impossible.

The lower diamictite grades upward into the lower sandstone unit (ls on maps). The lower part of this sandy sequence is generally brown (bs on maps), the middle green and silts (gz), and the upper part maroon to black (bks). Graded beds, load structures, and floating clasts are present. With what depositional environment are we dealing? Are there enough graded beds to convince you these are turbidites? What other environments are possible? Does this coarsening upward sequence represent strata deposited during a fall in sea level? The cobble conglomerate unit (cc on Figures 8 and 9) gradationally overlies dark sandstone (bks) containing isolated pebbles. The conglomerate appears massive though local sandy lenses occur. Most clasts are quartzitic. The matrix is a coarse sandstone. At least two depositional environments are possible for the cobble conglomerate: Submarine channel or shallow marine to beach. Laterally this conglomerate persists for 15 km south to Scout Mountain, but it is absent as a persistent unit north of Portneuf Narrows.

Above the cobble conglomerate is the middle quartzite and siltstone unit (ms). This unit is laterally variable and thickens drastically on Scout Mountain.

The upper diamictite (ud), still with west-dipping cleavage, forms a dark band just below the summit. It contains a heterogeneous clast assemblage and appears unbedded. A few conclusively glacially striated quartzite cobbles have been found here (Link, 1982, Figure 14). This diamictite is laterally persistent from Protneuf Narrows south to Garden Creek Gap (25 km).

Link (1982; 1983) interprets the upper diamictite as a subaqueous lodgement tillite deposited below a bouyant ice sheet. What other depositional mechanisms could explain the crude bedding, heterogeneous clast assemblage, and lateral continuity?

About 1 m of sedimentary breccia containing pink dolomite (pd) chips and flakes up to 50 cm long overlies the upper diamictite with an abrupt and probably disconformable contact. Along our traverse, the dolomite is preserved *in situ*, while about 100 m to the south the breccia is present. This dolomite overlies the

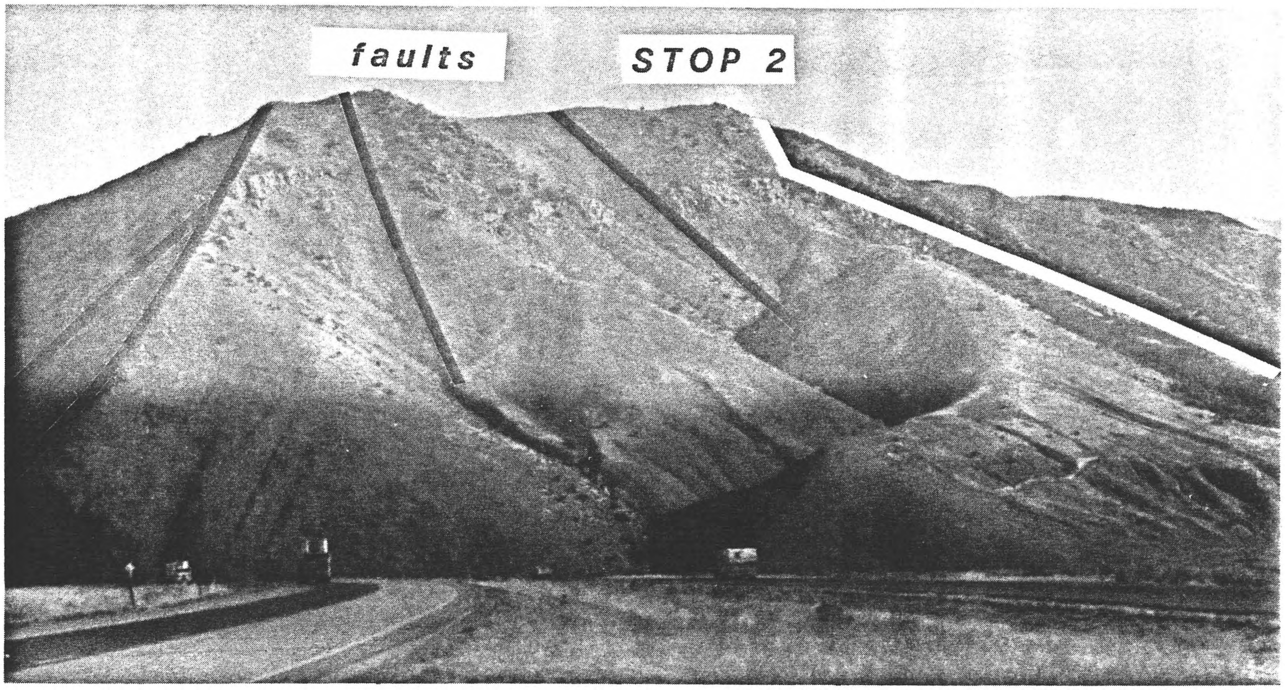

Figure 10. Portneuf Hill, south of Portneuf Narrows, looking south-east from I-15. Traverse of field trip Stop 2 is shown. Cliff-forming unit near the top of the hill is east-dipping cobble conglomerate, broken by southeast-trending minor faults (shown). A major fault truncates the cobble conglomerate at the north end of the ridge.

upper diamictite north of Portneuf Garden Creek Gap.

The upper part of the hill is underalin by tan sandstone interbedded with thin carbonates (usl on geologic maps). The gray shale of the upper member. Pocatello Formation (Zpu), occurs on the southeast side of the hill in a poorly exposed dip slope and along the crest of the hill about 1 km south of our traverse. This upper part of the Scout Mountain Member above the upper diamictite is well exposed in structurally overturned beds north of Portneuf Narrows (field trip STOP 3).

The elevation of the top of the hill is 6,015 ft. Congratulations, you have just climbed 1,371 ft. The view east is toward Inkom and the Portneuf Range. The two-level basalt flow in the Portneuf River valley orginated about 80 km upstream in the Gem Valley lava field. This flow near McCammon was dated at 140,000 years by K-Ar (Armstrong and others, 1975). Earlier carbon 14 dates estimated the flow to be about 30,000 years old (Malde, 1968). The Lake Bonneville flood flowed over the top of the lava flow and removed it completely from the Portneuf Narrows area.

Return down the ridge to the cars, and return north toward the Portneuf River.

0.8	10.9	Cross Portneuf River and Union Pacific Railroad, turn right on freeway frontage road.
1.3	12.2	Turn left on Blackrock Road and go under freeway.
0.3	12.5	Take immediate left after going under freeway.
0.3	12.8	**STOP 3**. Overturned Pocatello Formation north of Portneuf Narrows (Figure 12). Figure 9

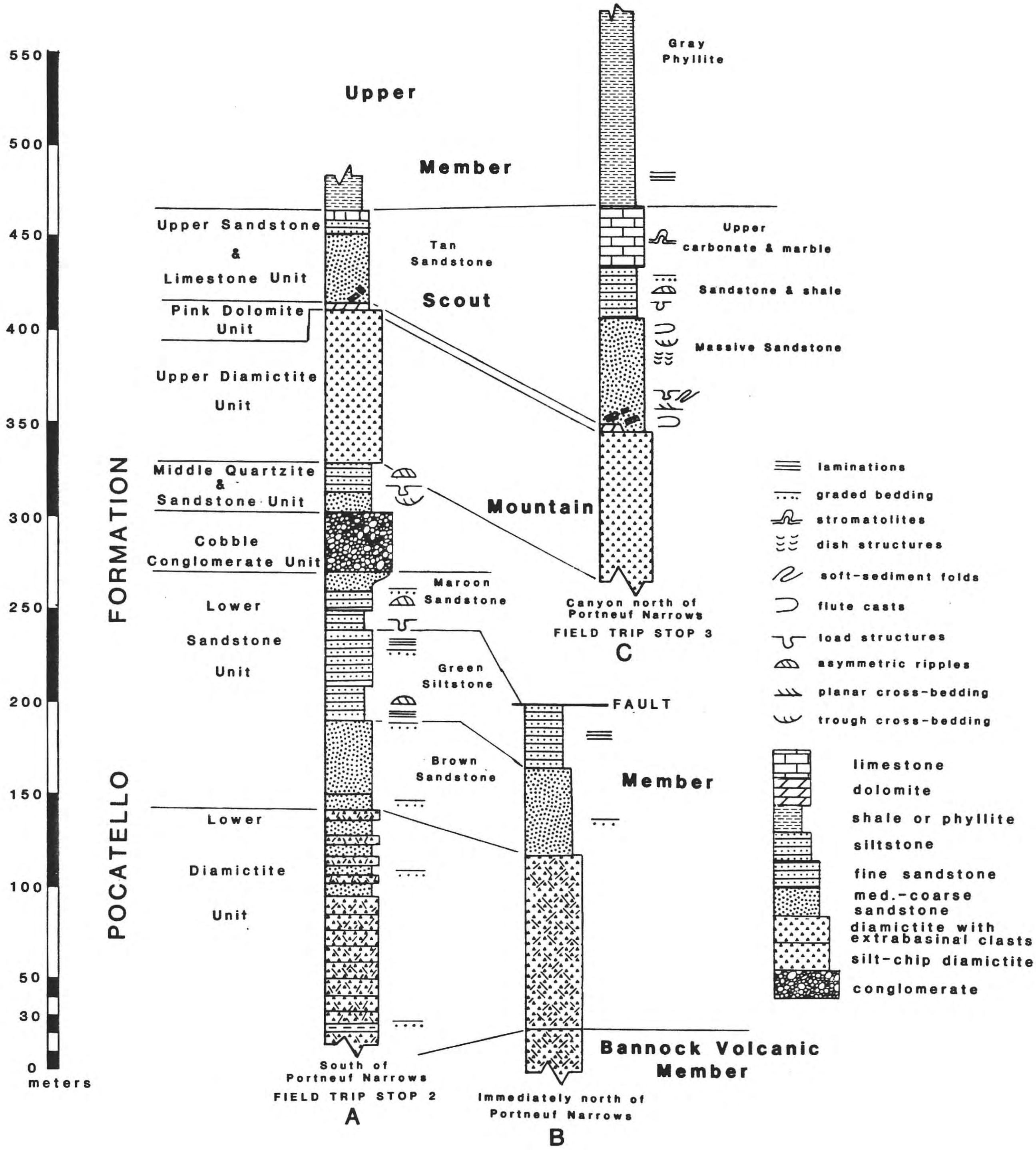

Figure 11. Stratigraphic sections of the Pocatello Formation in the Portneuf Narrows area (field trip STOPS 2 and 3).

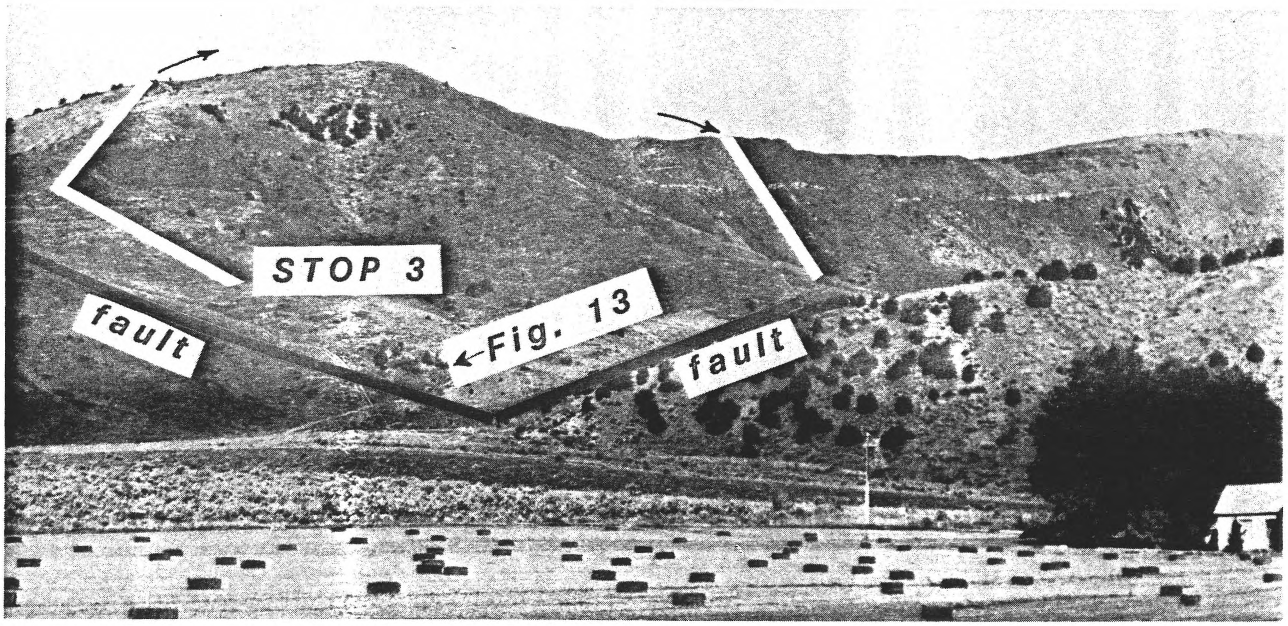

Figure 12. Northeast side of Portneuf Narrows looking northwest from the Portneuf River. Traverse of field trip STOP 3 is shown. Main ridge contains west-dipping over-turned strata of the upper part of the Scout Mountain Member and the upper member, Pocatello Formation. This is the original type area of the "lower member" of the Pocatello Formation. Juniper-covered slopes in the foreground are underlain by the Bannock Volcanic Member, and are separated from the main ridge by a major fault (shown). The fault is inferred to have left-lateral separation, and trends eastward near the base of the canyon. Location of Figure 13 is shown.

is a geologic map of this area. Figure 11 shows the stratigraphic section measured here. We are on the Inkom 7½-minute quadrangle.

The small canyon north and west of the vans was the type area and only exposure of the lower member, Pocatello Formation, defined by Trimble (1976, p. 13) and in Crittenden and others (1971, p. 583). This area has been shown to be structurally overturned (Link and others, 1980; Link, 1981, p. 739), and the dark shales originally placed in the lower member are recognized to instead belong to the upper member, Pocatello Formation as defined by Trimble (1976, p. 17-18) and in Crittenden and others (1971, p. 584-585).

We are in a structurally complex area, and most of the gullies contain faults (see Figure 9). A major fault separates greenstones of the Bannock Volcanic Member in the low hills near the vans from the main ridge of Scout Mountain and upper members to the north.

Cleaved phylite of the upper member, Pocatello Formation (Zpu), is tightly folded (Figure 13). The mesoscopic folds have shallow axes which plunge north. Some folds appear to verge to the west. At least three phases of deformation (cleavage formation, mesoscopic folding, and formation of kink bands) can be found in the phyllite. Folded cleavage in the upper member defines a north-plunging anticline in the main canyon to the north. This area is described in detail by LeFebre (1983).

The west-dipping overturned rocks of the upper part of the Scout Mountain Member are well exposed on the east-facing slopes of the main ridge. Unfortunately, view of the strata requires another steep climb of about 1,000 ft. We will be examining units above the upper diamictite, which underlies the crest and west side of the ridge. A measured section is shown in Figure 11, and condensed as section 7, Figure 3.

Proceeding uphill (downsection) one starts in dark shale and phylite of the upper member, Pocatello Formation. This grades stratigraphically down into a prominent white marble marker bed (ucm on Figure 9) (designated the top of the Scout Mountain Member by Link, 1982, p. 82) through a sequence of gray limestone and siltstone (upper sandstone and limestone unit of Figure 3). The marble contains laminated domal structures which resemble upside-down stromatolites.

Stratigraphically below the marble is a coarsening sequence of shale, siltstone, and sandstone (ss on Figure 9). Overturned sedimentary structures abound and include flaser beds, starved ripples, asymmetrical ripples, graded beds, scours, and load structures. The prominent brown cliff near the top of the hill is a subarkosic medium to coarse sandstone (mas on Figure 9) that contains large-scale through cross-beds, load and flute casts, parting lineation, slumped beds, and dish structures. Overturned cross-beds are best exposed on the southeast slope of the main ridge. This sandstone correlates with the sandstone at the top of the hill south of Portneuf Narrows.

Dolomite-chip breccia (pd) and black diamictite (ud) are exposed near the crest of the ridge. This ridge is a favorite hang-gliding departure point for local adventurers. The sequence exposed on this ridge is interpreted as representing a transgression by Link (1982, 1983). The dolomite-chip breccia was deposited in tidal channels which were covered by rapidly deposited beds of brown sandstone. Shallow marine conditions above storm wave base prevailed during deposition of the fining-upward siltstone sequence stratigraphically below the marble marker bed. The marble may be an open marine platform carbonate deposit. Further study of the possible stromatolites may better define the depositional environment. Quiet water conditions below storm wave base and with locally restricted circulation prevailed during deposition of the upper member, Pocatello Formation. These conditions are interpreted as the result of the post-glacial transgression.

What other interpretations are possible? In particular, how can the dolomite-chip breccia be best explained? Can the laminations in shale of the upper member be shown to be varves as proposed by Ludlum (1942, 1943)?

Return to the vans and to Blackrock Road.

0.3	13.1	Blackrock Road, turn right. Go under freeway.
0.3	13.4	Turn right onto frontage road, continue northeast on frontage road (South 5th Avenue).
5.2	18.6	Cross under I-15 freeway. Welcome to Pocatello.
1.0	19.6	Turn right on Barton Road, just past Farmers Insurance

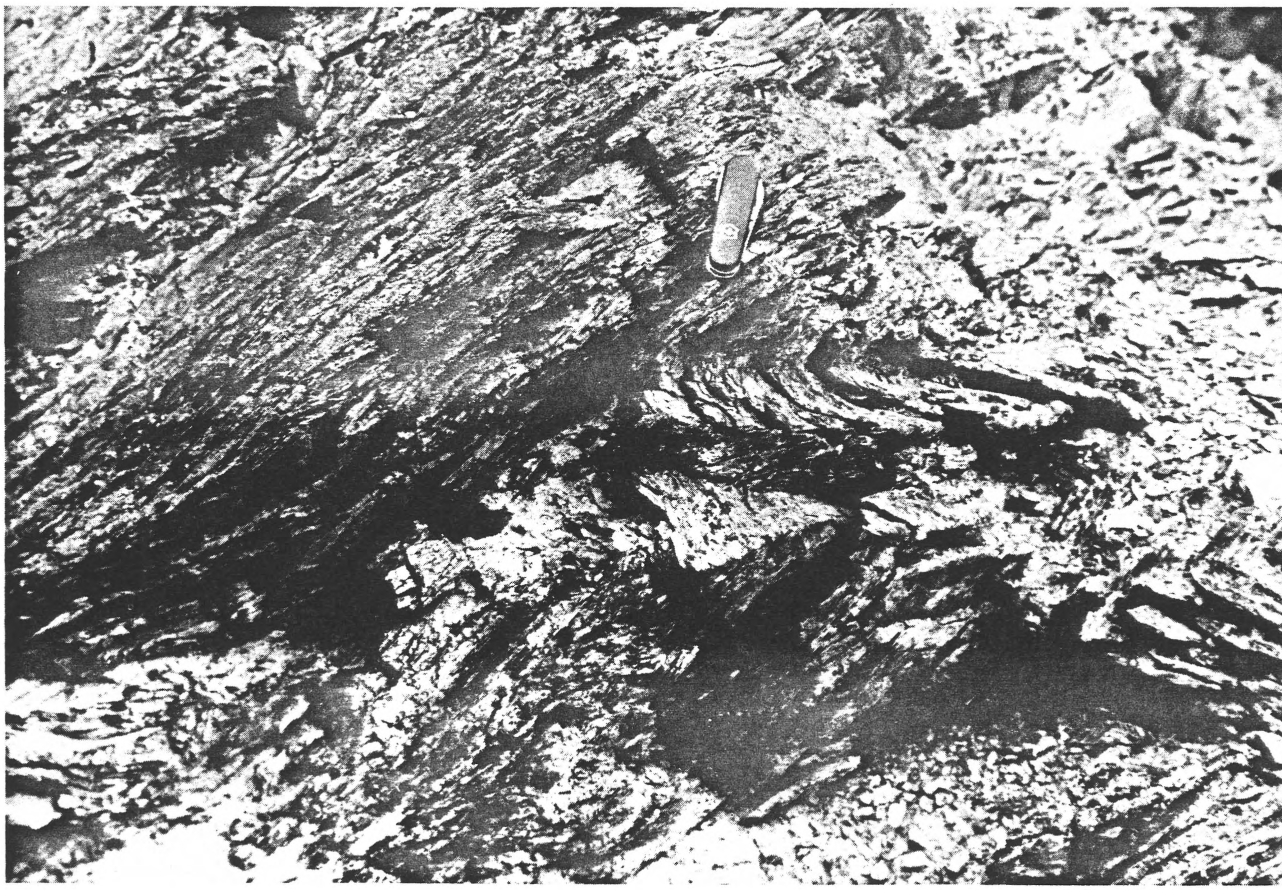

Figure 13. Tightly folded and cleaved shale of the upper member, Pocatello Formation, north of Portneuf Narrows. View is to the north; fold axes plunge to the north.

building. Proceed uphill towards Chinks Peak.

1.8 21.4 Bear right on main gravel road up the mountain. This is a good road as long as it is not snow covered. The Bannock Volcanic Member underlies the lower part of the ridge, below about 5,800 ft.

2.5 23.9 **STOP 4**. Summit Chinks Peak. This area is shown on Figure 7. Figure 8 is a geologic map of the area. We are on the Inkom 7½-minute quadrangle.

Beside the spectacular view east over the Pocatello Range to the Portneuf Range and west to the Snake River Plain, there are intriguing, and structurally overturned strata of the Pocatello Formation on this summit.

Diamictite (thought to be the lower diamictite unit, ld) is exposed about 200 m north of the radio facility. South and west of the summit occurs and anomalous cross-bedded and poorly lithified sandstone thought to correlate with the middle quartzite and siltstone unit (ms). The cross-beds appear to have been planar at first, but have been post-depositionally oversteepened. Can you convince yourself they are overturned?

Return to vehicles and to the Quality Royale on Pocatello Creek Road.

FIELD TRIP ROAD LOG: SECOND DAY
Pocatello, Idaho to Brigham City, Utah

Mileage Incre-mental	Mileage Cumu-lative	Description
0.0	0.0	Assemble at Quality Royale lot and proceed south on I-15.
8.4	8.4	Portneuf exit, the Portneuf Range and Mt. Bonneville loom to the east once we go through Portneuf Narrows. The Portneuf Range contains generally east-diping quartzitic strata of the "Brigham Group" and overlying lower Paleozoic carbonates (Corbett, 1978). The Pocatello Range contains east-dipping rocks of the "Brigham Group" (Anderson, 1928; Ludlum, 1942, 1943; Trimble, 1976).
8.0	16.4	Cross Portneuf River, near Inkom. The highway turns south into Marsh Valley. At the bend in the highway is a spectacular cut through columnar jointed basalt of the Portneuf lava flow. The rounded limestone hill east of the highway south of the Portneuf River is composed of Cambrian carbonates that dip west. It was kipuka, or island, in the Portneuf lava flow. Boulder deposits of the Lake Bonneville flood occur in road cuts west of this hill. Marsh Creek, through which flowed the Lake Bonneville flood, occupies the westside of Marsh Valley, while the Portneuf River is crowded against the east edge of the valley by the Portneuf lava flow. The peaks to the west are Scout Mountain and Old Tom Mountain of the Bannock Range. Elkhorn Peak is to the southwest, and Oxford Mountain is directly south. They are underlain by complexly faulted upper Proterzoic and lower Paleozoic strata. To the east are Mt. Bonneville and Haystack Mountain of the Portneuf Range.
8.7	25.1	U.S. Highway 30 exit to Lava Hot Springs and McCammon.
3.2	28.3	Jensen Road, exit 44.
3.6	31.9	Arimo dxit, turn right. Continue west on Arimo Road to Garden Creek Gap.
7.2	34.1	**STOP 5.** Garden Creek Gap. The stratigraphy here is shown in Figure 3, section 4. This picturesque canyon was cut by Garden Creek when it was following a course established on Pliocene valley fill. The fill has been completely removed, and the creek is superposed on bedrock. The bedrock here was recently assigned to the Scout Mountain Member in the course of regional mapping by the U.S. Geological Survey (L. B. Platt, oral communication, 1980) and subsequently studied by Thompson (1982). The east-dipping, slightly pebbly quartzite that makes up the cliffs in the gap is correlated with the cobble conglomerate unit south of Portneuf Narrows. The upper diamictite, overlain by dolomite, can be found at the top of the ridge about a mile south of the gap. We are on the Hawkins 7½-minute quadrangle. The quartzite of the Scout Mountain Member at Garden Creek Gap makes an excellent climbing rock. In the winter, ice accumulates in the big gully on the south side of the gap, and is used for ice climbing.

Turn around and proceed east back to I-15.

7.2 46.3 Bear right onto I-15 heading south.

4.5 50.8 U.S. Highway 91 exit, bear right, then turn left at stop sign toward Downey.

5.4 56.2 Entering Downey, population 645.

Oxford Mountain is the prominent peak to the south. The north end of Oxford Mountain was mapped by Link (1982; in press). the prominent break in slope below the Summit is caused by a younger-on-older fault which places "Brigham Group" strata on the Pocatello Formation.

South of Downey, U.S. Highway 91 parallels Marsh Creek, which now occupies the Lake Bonneville overflow channel. Gravel bars in the overflow channel and prominent truncated alluvial fans can be seen southeast of Downey.

7.3 63.5 Red Rock Pass. Bear right on County Road D4 (West Side Highway).

Red Rock Pass occupies the drainage divide between Marsh Creek (draining north to the Snake River) and the Lake Bonneville basin. It was through here, between 15,000 and 14,000 yrs ago (Scott and others, 1982, p. 3), that the Lake Bonneville flood poured (Malde, 1968). The dam for Lake Bonneville consisted of coalesced alluvial fans with sources on Oxford Mountain to the west and the Portneuf Range to the north and east. Truncated surfaces of the fans on Oxford Mountain are prominent west of the road south of Red Rock Pass. About 300 ft of downcutting occurred at Red Rock Pass during the flood, from the Bonneville shoreline (about 5,135 ft) to the Lake Provo level (about 4,800 ft).

3.6 67.1 Swan Lake Road, keep right. The village of Oxford is ahead and is to the right, the Bear River Range occupies the east side of Cache Valley to the south and east. From Oxford village southward, the West Side Highway travels on the Lake Provo shoreline (about 4,800 ft).

Lake Provo was the remnant of Lake Bonneville and emptied to the north through Red Rock Pass. The lake level dropped below the Provo shoreline about 13,500 years ago (Scott and others, 1982, p. 3). Great Salt Lake (elevation 4,202 ft) is the modern remnant.

The cliffs of the east face of Oxford Mountain loom west of the road. The Bannock Volcanic Member forms the lowest cliffs, with the overlying Scout Mountain Member just underneath the prominent color change from green to pink. That color change marks a younger-on-older flat fault with Cambrian-upper Proterozoic Camelback Mountain Quartzite ("Brigham Group") in the upper plate and Pocatello Formation in the lower plate.

3.9 71.0 Oxford village historical marker.

2.6 73.6 **STOP 6**. Boulders of the Bannock Volcanic Member. Just below the house west of the road, and along the fence east of the road, are boulders of greenstone which are representative of most of the litholo-

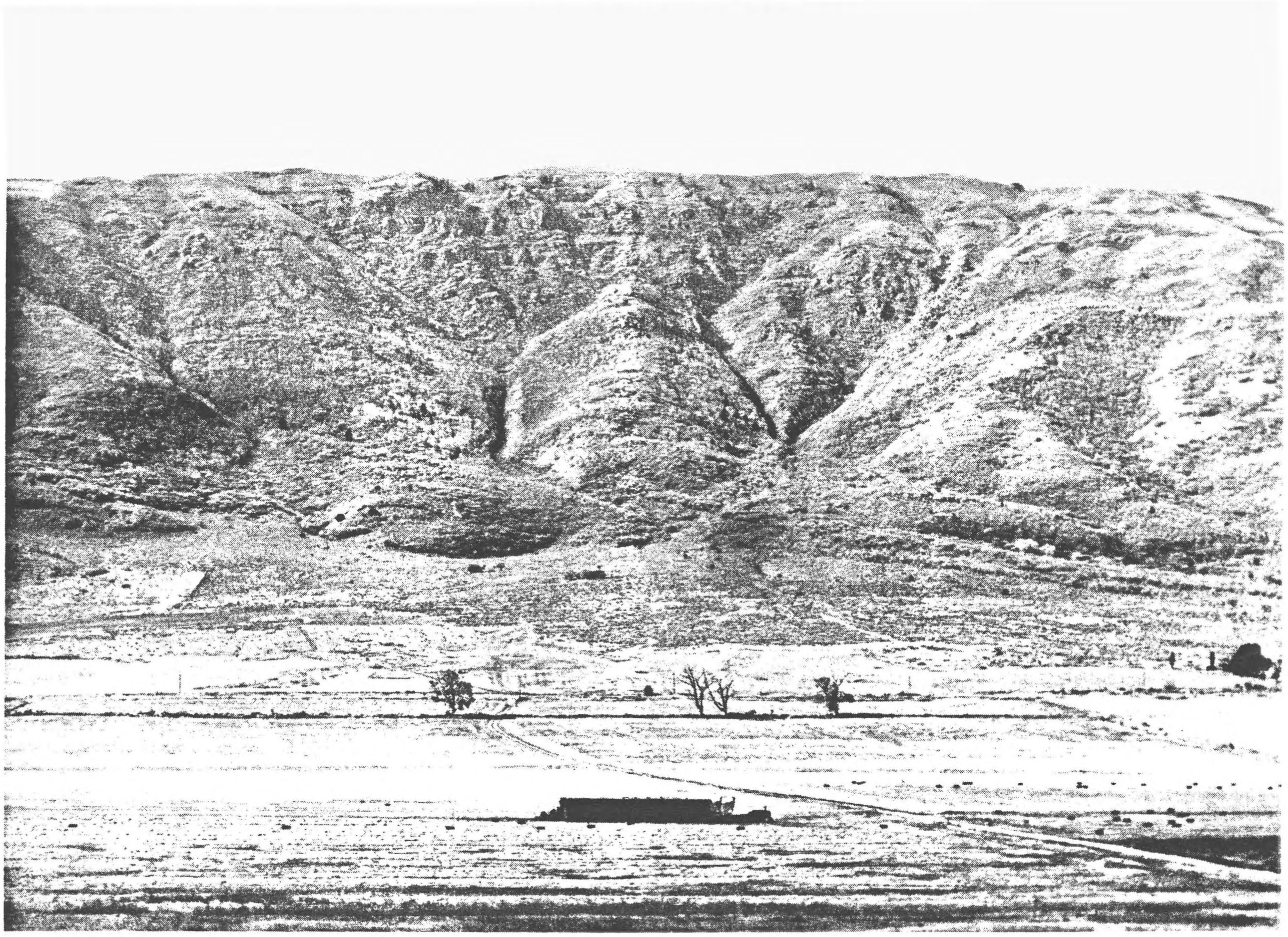

Figure 14. East side of Oxford Ridge north of village of Clifton. Base of cliffs is composed of the Bannock Volcanic Member. Volcanic component decreases upward and the upper portion of the cliffs is composed of the Scout Mountain Member. West Side Highway follows the Lake Provo shoreline behind the row of four trees. Field trip STOP 6 is at the extreme right of the photo.

gies present in the Bannock Volcanic Member on Oxford Mountain (Figure 14). We are on the Clifton 7½-minute quadrangle.

Among the rock types present are foliated fine-grained greenstone, chlorite-chip breccia, and coarse grained amphibolite that resembles a metagabbro.

The small round hill to the southeast is underlain by the Bannock Volcanic Member; foliation dips to the east. Access is via a private road 0.2 miles farther south on West Side Highway.

2.2 75.8 5600 N. Street in Clifton, Idaho. Turn left (east) toward Little Mountain (Figure 15) for field trip STOP 7.

1.8 77.6 Cross Union Pacific Railroad, turn left (north) on 4200 W. Street.

0.2 77.8 Turn right to Twin Lakes, keep left on north side of lake. If extended stop on Little Mountain is desired, cars can drive to the highest field on the northwest side of the mountain with the following directions.

0.7 78.5 Restrooms on lake shore.

0.8 79.3 Keep left, away from lake.

0.3 79.6 Keep left.

0.3 79.9 Keep left.

Figure 15. Little Mountain, looking southeast from north of the village of Clifton. Clifton benchmark (5,371 ft.) is at the summit, and is underalin by Scout Mountain Member. Cliffs below and to the right of the summit contain pillow lava fo the Bannock Volcanic Member. Prominent bench below the cliffs is the Lake Bonneville Shoreline. The Lake Provo shoreline is just above the level of the flat plowed fields in the foreground.

0.7	80.6	Turn right.
0.3	80.9	Cross Twin Lakes Canal; turn right.
0.3	81.2	End of dirt road in field, south of Twin Lakes. **STOP 7**. Little Mountain. We will walk about a mile south up the ridge. Figure 16 is a geologic map of the Twin Lakes-Little Mountain area. We are on the Banida 7½-minute quadrangle. Lithologies present include argillite, diamictite, and lithic sandstone (Scout Mountain Member) and foliated greenstone, breccia, and pillow lava (Bannock Volcanic Member). The best exposure of pillow lava is on the southwest flank of the peak containing the Clifton benchmark (5,371 ft) about 200 ft below the summit. Turn around and return to entrance to Twin Lakes recreation area.
3.6	84.8	Turn left on 4200 W. Street.
0.2	85.0	Take first right, cross tracks and proceed west on 5600 W. Street.
1.8	86.8	West Side Highway, turn left.
1.1	87.9	Pass Twin Lakes Mercantile on right and turn right on First South Street in Clifton.
0.2	88.1	Turn left on Cemetary Street. Follow this narrow dirt road to the south and west, passing stable on left, and around one sharp bend to the right as the road follows switchbacks up the hill.

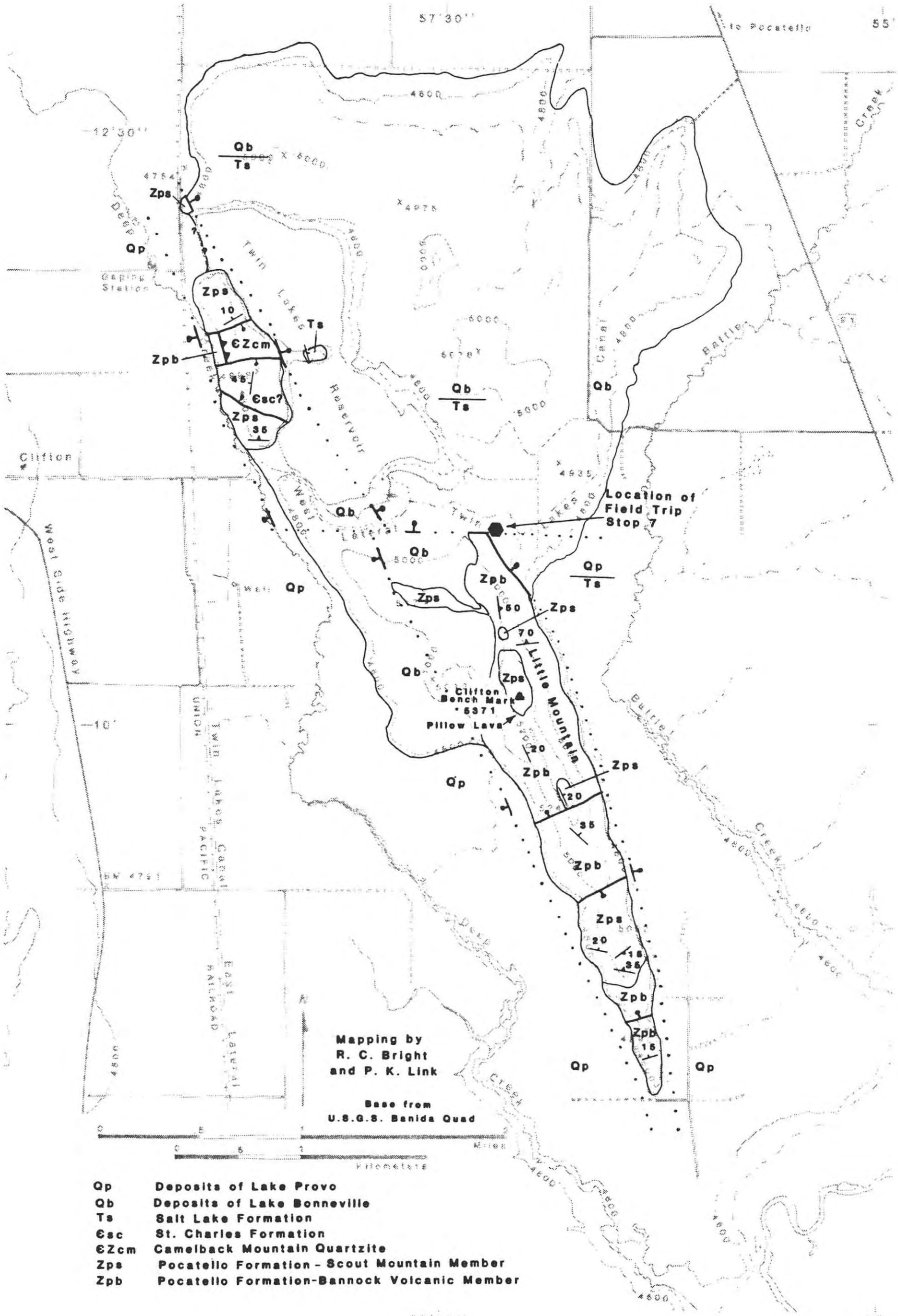

Figure 16. Geologic map of the Little Mountain area, east of Clifton, Cache Valley, Idaho.

2.8 90.9 Bear right through gate, following sign to "Lodge."

Geologists following this road log on their own can proceed 1.8 mi up the switchbacks to an outcrop of Bannock Volcanic pillow lava and greenstone on the west side of the road just below where it turns west into Clifton Basin. Be careful, for there is no place to park except on the road.

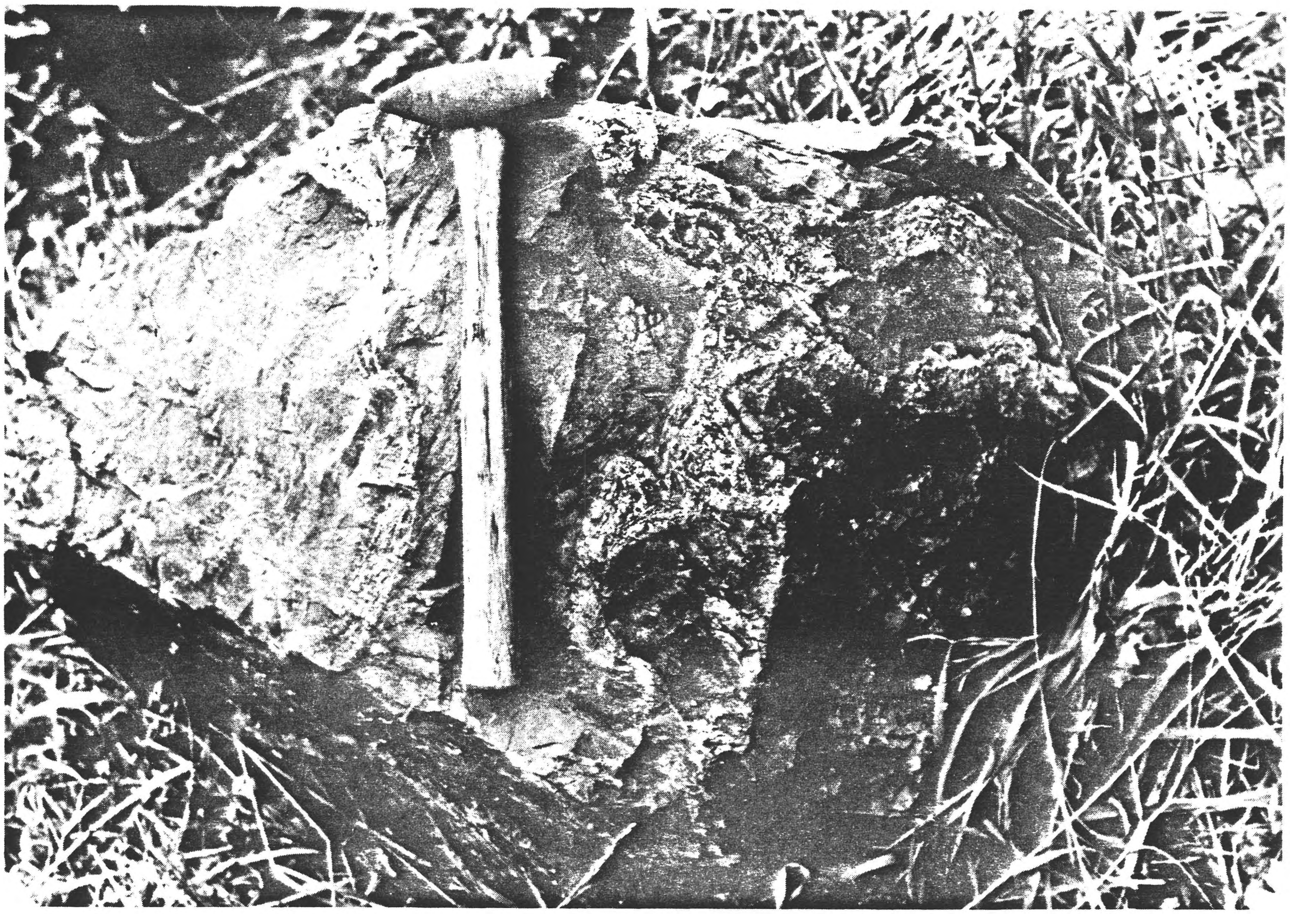

Figure 17. Boulder of pillow lava, Bannock Volcanic Member, at hunting lodge, east of Clifton Basin, field trip STOP 8. The pillows are composed of albite, epidote, and chlorite greenschist. Interpillow material is chlorite-quartz breccia.

0.8 91.7 **STOP 8**. Hunting lodge. This is private property (Lynn Davis, Clfton, Idaho, owner). Please be considerate. We are on the Clifton 7½ minute quadrangle.

Several boulders of pillow lava and probable pillow breccia can be found at the northwest corner of the driveway in front of the lodge (Figure 17). The adventuresome can examine the moss-covered cliffs on the south side of the creek where a variety of greenstone breccias crop out. A diabase dike intrudes the greenstone about halfway up the cliff face on the north side of the creek. A stratigraphic section measured here is shown as section 2, Figure 3.

Follow the one-way driveway east and south back to the main dirt road and return to West Side Highway in Clifton.

3.8 95.5 West Side Highway, turn right.

5.2 100.7 Turn right on Five Mile Road in village of Dayton (just before country store on left). Follow gravel road west up Five Mile Creek.

1.2 101.9 Keep left at fork.

0.3 102.2 Turn sharp left.

0.5 102.7 STOP 9. Five Mile Canyon, Weston Canyon 7½ minute quadrangle. Park in gravel pit on right and walk up gravel road. A partial section measured here is shown as section 1, Figure 3.

Diamictite of the Scout Mountain Member is well exposed in the canyon. Bedding is apparently nearly parallel with cleavage, dipping shallowly to the west. Hints of bedding include horizons of clasts and rare non-pebbly intervals. Clast types include quartzites, siltstones, mafic volcanics and granitic rocks.

Near the tip of the ridge on the south side of the creek and also north of the road at the west end of the narrow canyon, Cambrian Blacksmith Limestone lies with low-angle fault contact on diamictite of the Scout Mountain Member (Oriel and Platt, 1979). This represents a stratigraphic omission of 3 km of section. This is not a regional omission as the missing strata are present elsewhere on Oxford Mountain (Christie-Blick and Link, 1983).

Turn around and proceed back to West Side Highway in Dayton.

2.1 104.8 West Side Highway, turn right.

0.2 105.0 Bear left (east) toward Preston and the Bear River Range on the east side of Cache Valley.

The Bear River Range contains dominantly east-dipping upper Proterozoic and Paleozoic strata. The Paris thrust crops out on the east side of the range and places upper Proterozoic quartzite over Mesozoic rocks. The foothills in front (west) of the range are underlain by the Miocene-Pliocene Salt Lake Group which in Cache Valley has been divided into the Collinston Conglomerate (base), Cache Valley Formation, and Mink Creek Conglomerate (Adamson and others, 1955).

2.2 107.2 Sanitary landfill on left, we are descending toward the Bear River, which has cut terraces in Bonneville and Provo lake-bottom deposits.

2.2 109.4 Cross Bear River.

2.2 111.6 State Street (U.S. Highway 91) in downtown Preston, turn right. We will proceed south on U.S. Highway 91 along the east side of Cache Valley.

2.3 113.9 Whitney turn-off to the left.

Little Mountain west of Franklin, Idaho, lies to the southeast and contains prominent Lake Bonneville and Lake Provo shorelines.

4.5 118.4 Franklin city limits.

1.7 120.1 Utah state line.

3.6 123.7 High Creek Road to left.

The Malad Range forms the low mountains on the west side of Cache Valley while Wellsville Mountain looms to the southwest.

1.7 125.4 Richmond, Utah. We are travelling on the Lake Provo shoreline.

5.4 130.8 Entering Smithfield, Utah.

7.1 137.9 Cache Valley Mall in Logan, Utah.

2.5 140.4 Cross Logan River.

We are turning west and crossing to the southwest corner of Cache Valley. Wellsville Mountain is directly ahead. The geology of Cache County, Utah, is discussed by Williams (1958).

9.0 149.4 Intersection with Utah Highway 23.

We are heading up Wellsville Canyon and downsection in east-dipping beds. The sequence contains Pennsylvanian-Permian Oquirrh Formation at the top and continues down to upper Proterozoic quartzite of the "Brigham Group" near Brigham City.

7.0 156.4 Box Elder Summit, elevation 5,894 ft. The view south is toward the Wasatch Range in the area mapped by Sorensen and Crittenden (1976b).

2.9 159.3 Mantua turn-off to left.

1.6 160.9 Box Elder campground on left. We are on the Box Elder County map of Doelling (1980).

We are now in the type area of the Brigham Quartzite or Group first named by Walcott (1908, p. 9) and discussed by Sorensen and Crittenden (1976a). Here the "Brigham Group" contains the Caddy Canyon Quartzite at the base, and the Geertsen Canyon Quartzite at the top.

2.7 163.6 Brigham City limits.

We are heading southwest and down in elevation through gravel deposited in Lake Bonneville. As the vista of Great Salt Lake opens up, the Promontory Mountains are directly west, with the Blue Spring Hills to the northwest.

2.9 166.5 Turn left on U.S. Highway 89 at stoplight.

0.2 166.7 Enter town of Perry.

2.2 168.9 Osmond Lane (3000 S.), turn left.

0.5 169.4 Bear right at gravel pit.

0.7 170.1 End of road in Perry Canyon.

STOP 10. Figure 18 is a map of the area. We are on Willard 7½ minute quadrangle.

We will be examining east-dipping diamictite-bearing rocks informally designated the formation of Perry Canyon by Sorensen and Crittenden (1976b), and discussed by Blick (1979, p. 178-182), and Crittenden and others (1983). Diamictite on the north side of the creek in the narrow canyon belongs to the upper part of the formation and is overlain by arkosic quartzite, siltstone, and stromalitic limestone of the Maple Canyon Formation (Crittenden and others, 1971). The Maple Canyon Formation crops out, west of a normal fault, on the north side of Perry Canyon at the end of the dirt road.

The formation of Perry Canyon contains two horizons of diamictite separated by perhaps 525 m of poorly exposed sandstone and siltstone. The lower diamictite horizon crops out on the west face of the Wasatch Range south of Perry Canyon at an elevation of about 5,200 ft. Its examination requires a steep cimb of about 500 ft. The lower diamictite varies from 60 to 365 m thick and appears massive though intercalated siltstone is present locally. Clasts are heterogeneous and up to 3 m in diameter. Gneissic granite and quartzite are the dominant clast types; schist, metarhyolite, basalt(?), and carbonate clasts are also present. The upper diamictite in Perry Canyon consists of schistose and faulted, thickly bedded black diamictite with clasts up to 1 m in diameter. The clasts here are about 2/3 granite and gneissic rocks and 1/3 quartzite. Mylonitic zones

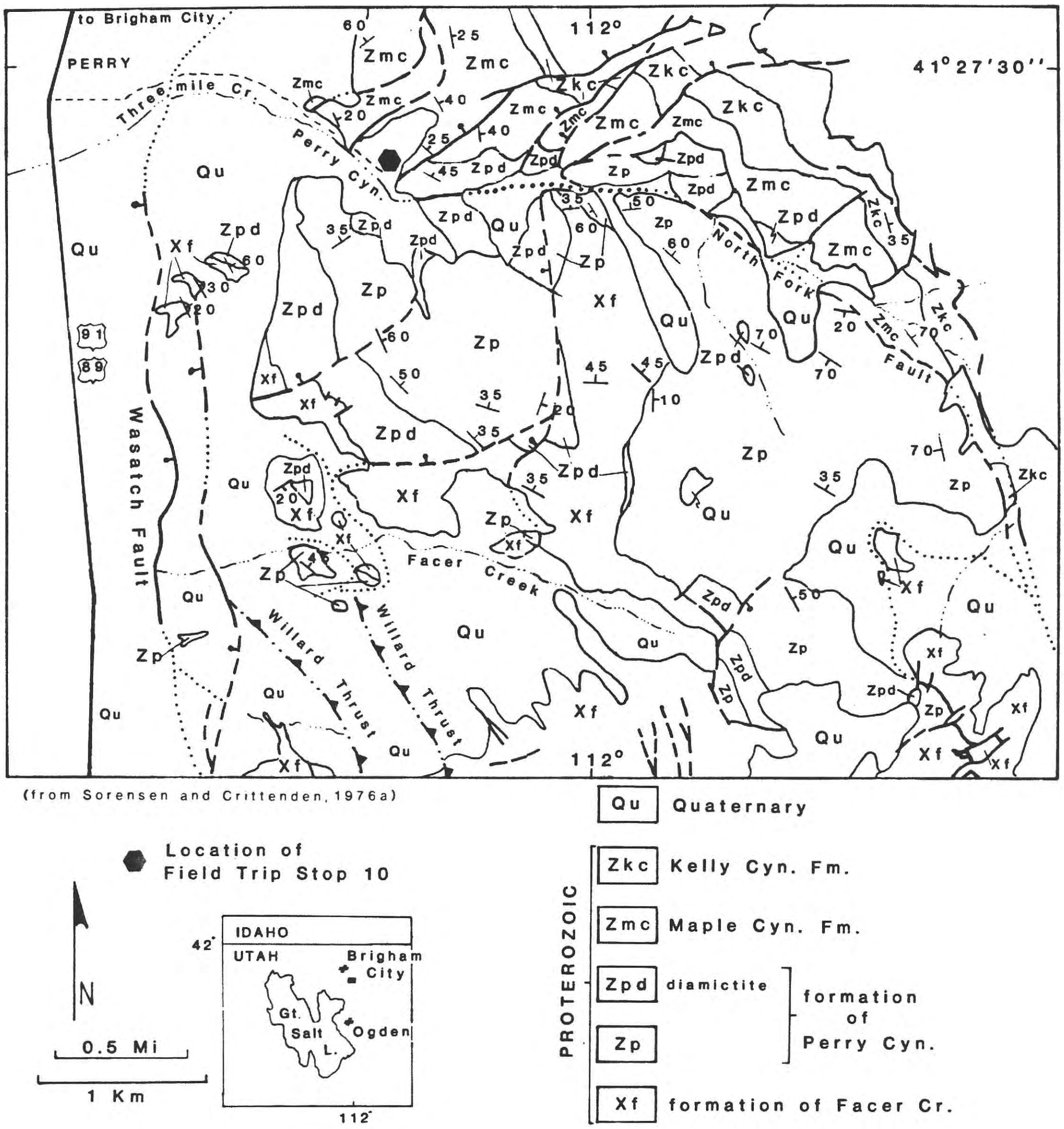

Figure 18. Geologic map of the Wasatch Front near Perry and Facer Canyons, southeast of Brigham City, Utah.

with extremely sheared clasts are present. It is this diamictite which we will examine and which is correlated with all diamictites of the Pocatello Formation by Crittenden and others (1983).

END OF ROAD LOG.

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Utah Geological and Mineral Survey Special Studies 60, 1983
Guidebook Part 2 — GSA Rocky Mountain and Cordilleran Sections Meeting
Salt Lake City, Utah — May 2 - 4, 1983

EVOLUTION OF EARLY MESOZOIC TECTONOSTRATIGRAPHIC ENVIRONMENTS—SOUTHWESTERN COLORADO PLATEAU TO SOUTHERN INYO MOUNTAINS

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LOWER MESOZOIC STRATIGRAPHY AND DEPOSITIONAL SYSTEMS, SOUTHWEST COLORADO PLATEAU

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INTRODUCTION

Lower Mesozoic strata are widely exposed across the western Colorado Plateau and vicinity in southern Utah, northern Arizona, and southern Nevada (Figure 1). This paper reviews previous work and provides an update of work in progress on the Moenkopi, Chinle, Moenave, and Kayenta Formations and lowermost Navajo Sandstone. The Wingate Sandstone is shown on Figures 2 and 3 but is not further discussed.

Although our data are brief and conclusions are somewhat generalized, the information is based on extensive past and ongoing research. The interested reader is referred to the references for a more extensive discussion of the units covered in this paper.

Although the age of the Navajo, Kayenta and Moenave Formations is somewhat uncertain, we follow the assignments of Peterson and Pipiringos (1979). Their evidence is based on palynomorphs found in the Whitmore Point Member of the Moenave Formation which suggest a Liassic (Early Jurassic) age.

MOENKOPI FORMATION

Regional Stratigraphy

The Lower Triassic Moenkopi Formation forms a west-thickening wedge of siltstone, mudstone, very fine-grained sandstone, gypsum, and local conglomerate (Figure 2). Carbonate content increases sharply to the west of the Hurricane Cliffs; redbed clastic lithofacies are dominant to the east. Cross-stratified sandstone is uncommon except in southeastern Utah (Blakey, 1974) and the Little Colorado River Valley, Arizona (McKee, 1954). Thicknesses

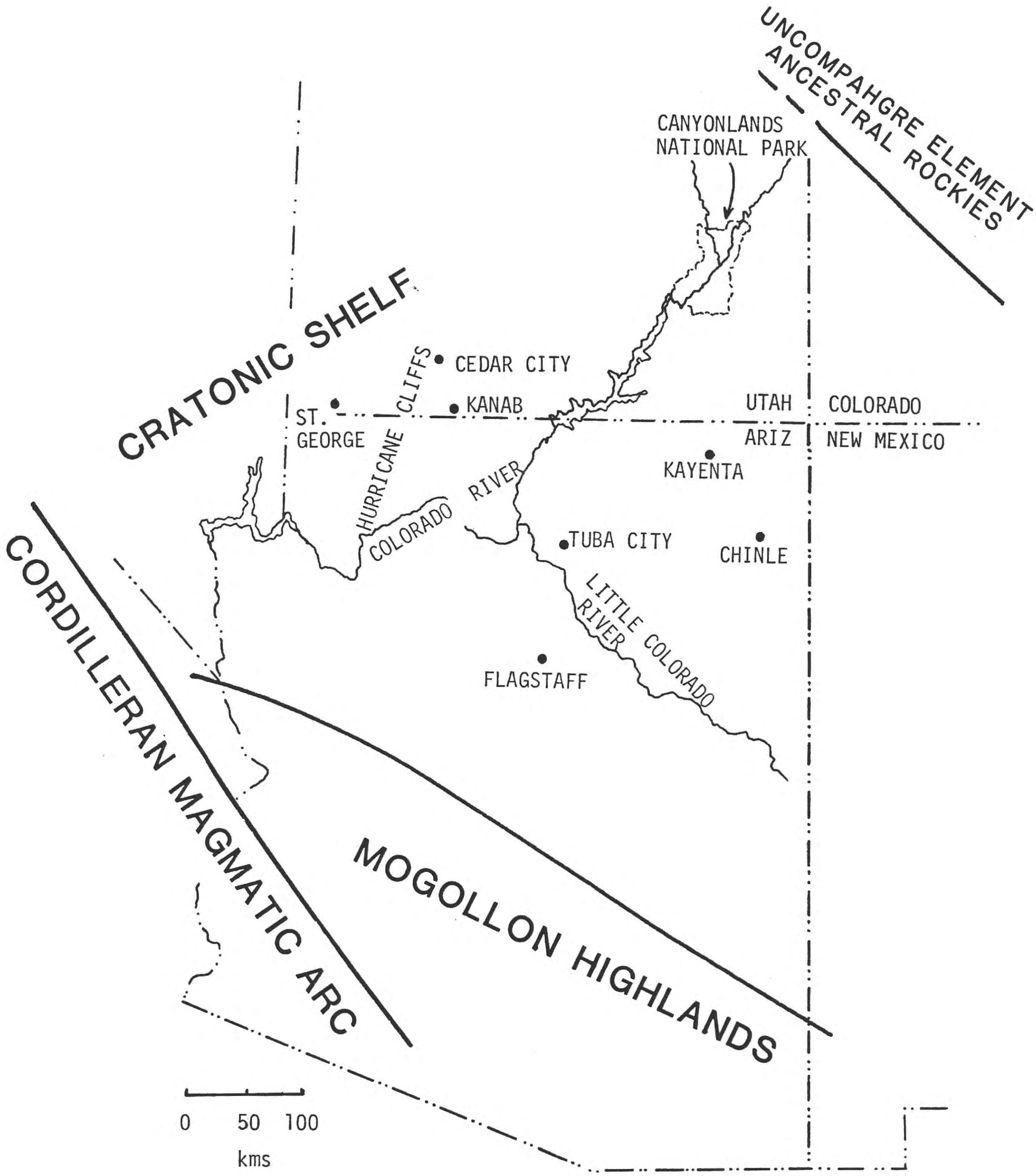

Figure 1. Map showing early Mesozoic tectonic features of Colorado Plateau and vicinity.

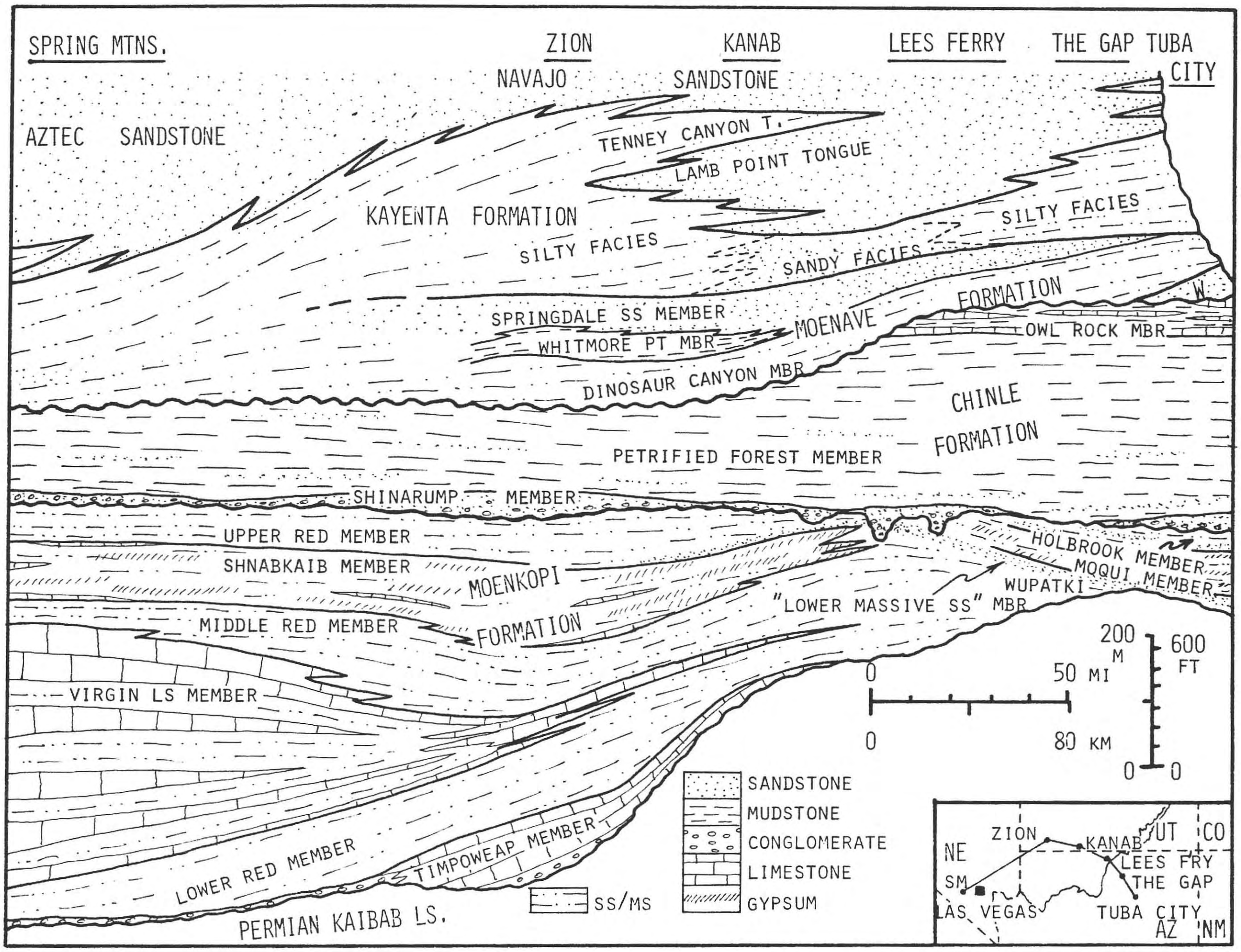

Figure 2. Restored generalized lithologic cross section of lower Mesozoic rocks from Las Vegas, Nevada, to Tuba City, Arizona. Sources of data: Harshbarger and others (1957), McKee (1954), Stewart and others (1972 a, b), Wilson (1958), and Wilson and Stewart (1967). W - Wingate Sandstone.

range from about 100 m in eastern Utah and Arizona to 600 m in the Zion-Las Vegas areas.

Nomenclature is subregional, reflects a rather complicated evolution of terminology, and comprises a mixture of formal and informal members (Figure 3). Regional detailed temporal correlation is hindered by broad areas of nonexposure between outcrop belts but is aided by two well-known ammonite zones, *Meekoceras* and *Tirolites.* Regional lithologic correlation is based on broad lithologic similarities and stratigraphic position (Blakey, 1974).

Facies

Pale reddish-brown sandy siltstone and mudstone dominate much of the Moenkopi Formation. Sedimentary structures include planar bedding, ripple lamination, ripple marks, and mud cracks, biogenic structures, salt crystal cracks, and locally abundant flaser bedding. Cross-stratified sandstone is abundant in southeastern Utah and the Little Colorado River Valley, Arizona. The former comprises sheets and lenses up to several meters thick with lobate sheet-like bodies increasing in abundance to the west and northwest (Blakey, 1974). The latter comprises cyclic sheet-like bodies with scoured bases in upward-fining sequences, and broad sheets with planar to low-angle stratification. Limestone and dolomite increase in percentage westward across the Moenkopi shelf and include fossiliferous packstone, oolitic grainstone, pelleted wackestone, stromatolitic boundstone, and birdseye dolomitic mudstone. Nodular gypsum is common in the Shnabkaib and Moqui Members and is associated with red and blue-green mudstone and dolomite.

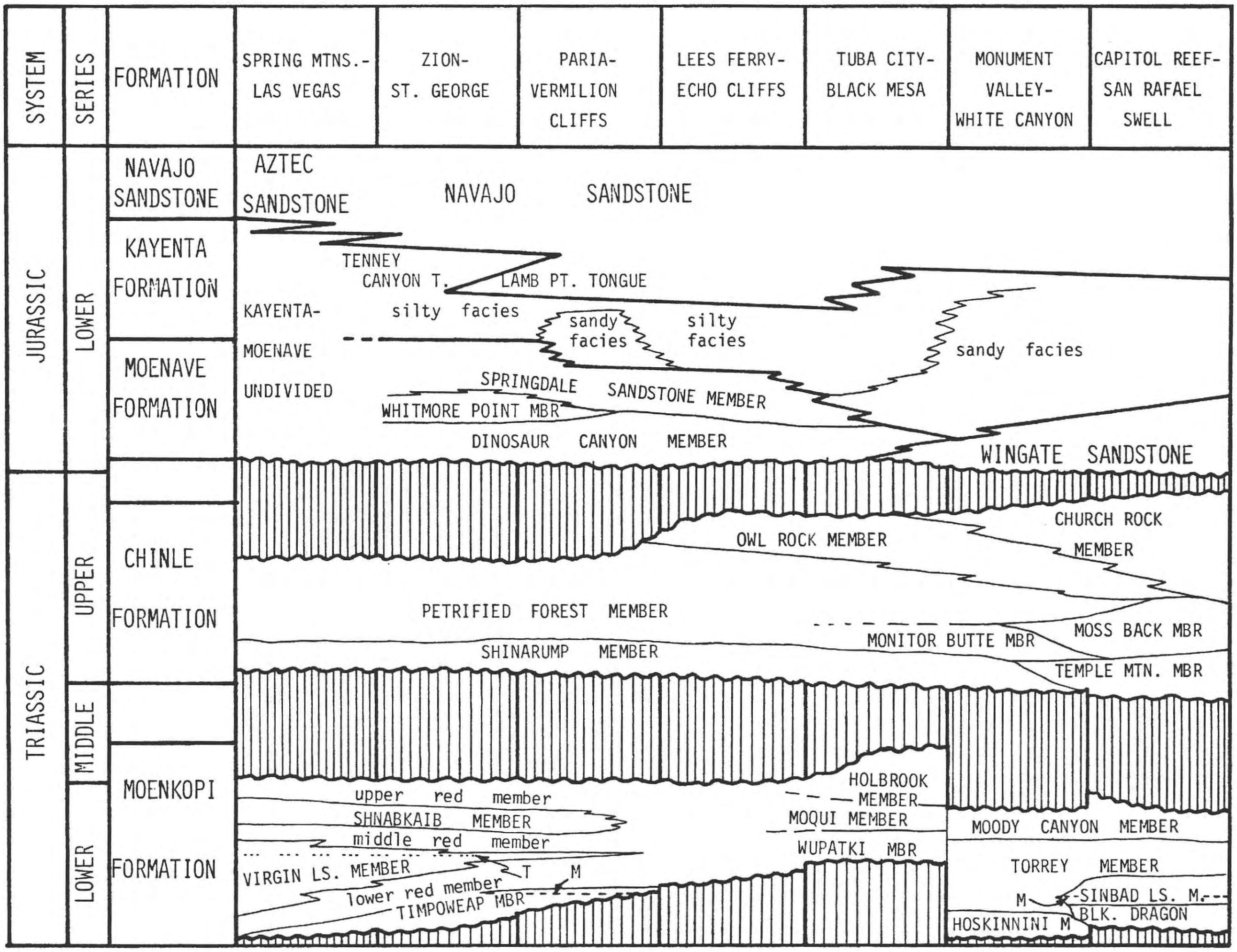

Figure 3. Age and correlation of lower Mesozoic rocks, western Colorado Plateau and vicinity. Age of Moenave, Kayenta, and Navajo after Peterson and Pipiringos (1979). M - Meekoceras zone; T - Tirolites zone.

More detailed descriptions of Moenkopi facies are given by McKee (1954), Stewart and others (1972a), Blakey (1974, 1977, 1979), and Baldwin (1973).

Paleogeography and Provenance

The Moenkopi was deposited during several transgressive-regressive cycles (Stewart and others, 1972a; Baldwin, 1973; Blakey, 1974; Abendshein, 1978). The Black Dragon and Timpoweap-Sinbad Members were deposited during the first cycle. The sequence is dominated by shoreline and shallow marine deposits (Blakey, 1974, 1979). The lower red, middle red, Torrey, and Virgin Limestone Members were deposited during the second cycle. The Virgin formed in clear-water shallow marine conditions (Auld, 1976) and the redbeds formed on broad coastal plains. A large delta prograded into the seaway from southeastern Utah (Blakey, 1974). The Shnabkaib, Moody Canyon, upper red, Wupatki and Moqui Members formed during a third cycle. Much of the lower massive sandstone of McKee (1954) was deposited on a wave-dominated coast (Baldwin, 1973). Much of the Shnabkaib represents restricted marine deposition as indicated by abundant gypsum. The Moqui Member probably formed in a broad sabkha to the east of the sea. The Holbrook Member is the youngest unit in the Moenkopi and is probably Middle Triassic (Michael Morales, Museum of Northern Arizona, personal communication, 1982). It was deposited by a major fluvial system that flowed northwest across the present Little Colorado River region. Equivalent units in the northwest, if ever deposited, have been removed by pre-Chinle errosion (Eugene Shoemaker, U.S. Geological Survey, personal communication).

Stewart and others (1972a) identified major

Moenkopi source terranes as the Mogollon Highlands and Ancestral Rockies. General sediment transport was to the northwest. An unknown amount of fine-grained detritus may have had a seaward source and may have been trapped in coastal environments.

CHINLE FORMATION

The Upper Triassic Chinle Formation was deposited across most of the Colorado Plateau and adjacent areas. The unit thins gradually in all directions from the Chinle area. Thicknesses generally range from 150-600 m. The Chinle comprises bentonitic mudstone, fine- to very coarse-grained sandstone, micritic limestone, and siliceous- and sedimentary-pebble conglomerate. Vertical and lateral changes are complicated and are described by Stewart and others (1972b) and Blakey and Gubitosa (1983a, b).

Chinle nomenclature is regional in extent, although distribution of members yields a varying stratigraphic column from place to place (Figures 2 and 3). Faunal zones are not established, so correlation is based on cyclic events as seen in vertical sections. These broad cyclic events can be correlated regionally (Blakey and Gubitosa, 1983a).

Facies

Bentonitic sandy mudstone is the most voluminous facies in the Chinle Formation. Thick sequences are generally color banded and may contain thin, less than 1 m-thick sandstone or thicker channel-shaped sandstone bodies 10 m or more thick. Associated sandstone contains mostly trough cross-stratification; epsilon cross-stratification is less common. Broad sheet, locally conglomeratic, sandstone forms the Shinarump and Moss Back Members. Planar-tabular and trough cross-stratification and planar bedding are the dominant sedimentary structures. Limestone occurs as bedded micrite, micritic and oncolitic nodules, calcarenite, and calcirudite (Gubitosa, 1981) and forms significant parts of the Owl Rock (bedded micrite) and Moss Back (nodules, calcarenite, calcirudite) Members. Chert-, quartzite-, and quartz-pebble conglomerate are distributed within the Shinarump and Moss Back Members. In most locations it is accessory to cross-stratified or planar-bedded sandstone.

Paleogeography and Provenance

The Chinle Formation was unconformably deposited on the Moenkopi Formation within a broad nonmarine basin that lay on the edge of the craton in a backarc setting (Blakey and Gubitosa, 1983a). The Shinarump and Moss Back Members were deposited in broad braided stream complexes. The Monitor Butte and Petrified Forest Members formed in swamps, flood plains, and channels associated with widespread meandering river systems. The Owl Rock Member, characterized by intercalated micritic limestone and mudstone was deposited within a series of broad, shallow, possibly ephemeral lakes. In and around Canyonlands National Park, the Moss Back Member intertongues with the Petrified Forest and Owl Rock Members. Gubitosa (1981) identified complex fan-delta, fluvial and lacustrine depositional systems. The Church Rock Member was deposited during a period of increasing aridity as indicated by increasing red color, considerable sheet-flood deposits, and eolian deposits (Blakey and Gubitosa, 1983a). Broad ephemeral streams, scattered dune fields, and rare lakes formed on a flat fluvial plain.

Chinle streams flow to the north and west from the Mogollon Highlands and Ancestral Rockies (Stewart and others, 1972b). Volcanic ash was blown into the basin from the Cordilleran arc terrane which lay to the southeast. Whether the streams flowed to the distant Pacific Ocean or dried up in desert sands in western Utah and Nevada is unknown.

MOENAVE FORMATION

Regional Stratigraphy

The Early Jurassic Moenave Formation overlies and intertongues with the Wingate Sandstone in northeastern Arizona and southeastern Utah and disconformably overlies the Chinle Formation in northwestern Arizona and southwestern Utah (Averitt and others, 1955). The Moenave averages 100 m in thickness in the east and over 175 m near Cedar City and Zion, Utah (Figure 2). The formation is dominated by red, brown and variegated fine-grained siliciclastics and minor carbonates. Although facies changes are pronounced and occur over short distances, in general the Moenave becomes finer grained toward the west.

As defined by Harshbarger and others (1957) and Wilson (1958) the formation includes three members: the Dinosaur Canyon, Whitmore Point, and Springdale Sandstone (Figure 2). The Whitmore Point Member intertongues with the basal part of the Springdale Sandstone Member and is not recognized east of Johnson Canyon, Utah (Figure 2). In the eastern outcrop areas the Springdale Sandstone overlies the Dinosaur Canyon Member. The slope-forming Dinosaur Canyon Member comprises red to orange siltstone, very fine-grained sandstone, and minor claystone (Wilson, 1958). The Whitmore Point is composed of grey to red siltstone and claystone and minor carbonate and, where present, forms a prominent slope beneath the resistant Springdale Sandstone. The Springdale Sandstone consists of red, yellow, and grey fine- to medium-grained sandstone minor shale and conglomerate, the latter occurring at the base of the member and as scattered lenses throughout the unit.

Facies

Planar-tabular and trough cross-stratified lenticular sandstone and siltstone occur sporadically throughout the Dinosaur Canyon Member (Wilson, 1958). These facies are typically enclosed within siltstone and/or claystone. Mudcracked surfaces are common in the mudstones. Asymmetric ripple cross-laminated sandstone and siltstone are common. Horizontally laminated siltstones and claystones are interbedded with the rippled beds and comprise the bulk of this member. The Whitmore Point contains minor trough and planar-tabular cross-stratified sets of very fine-grained sandstone and abundant horizontally laminated and ripple-laminated siltstones and claystones (Wilson, 1958). Coarser grained facies increase to the east where this member intertongues with the overlying Springdale Sandstone. Facies within the Springdale Sandstone include: small- to large-scale trough cross-stratified sandstone, many of which consist of low-angle foresets, planar-tabular cross-bedded sandstone, horizontally-bedded sandstone, ripple cross-laminated sandstone and siltstone, mud pellet conglomerate, and minor mudstone.

Paleogeography and Provenance

The Moenave Formation was deposited in a west- to southwest-trending basin with a source area to the northeast in the Uncompahgre Upwarp (Harshbarger and others, 1957), and also numerous localized source areas consisting of older Mesozoic-Paleozoic, and Precambrian rocks (Johnson, 1969). Deposition occurred in a variety of fluvial systems ranging from small wadis to high-sinuosity meandering streams (Harshbarger and others, 1957; Johnson, 1969). A fluvial origin for much of the Moenave is indicated by the presence of fining-upward sequences of cross-stratified sandstone, lateral (point bar) accretion surfaces (Allen, 1965), and abundance of mudstones representing overbank sedimentation. Carbonates interbedded with the siliciclastics accumulated in small lakes on the floodplain. Fish and crocodile fragments occur in both the Dinosaur Canyon and Whitmore Point members suggesting a tropical to subtropical climate (Harshbarger and others, 1957; Wilson, 1958).

A subtidal to intertidal origin for parts of the Dinosaur Canyon Member has been proposed by Davis (1977). This interpretation is based primarily on bimodal-bipolar paleocurrent patterns and also the presence of flaser bedding and trace fossils with presumed tidal affinities.

Much of the Springdale Sandstone likewise is the product of fluvial sedimentation containing both braided and meandering deposits. Avulsive events were apparently frequent as evidenced by numerous overlapping channel deposits and variability of cross-bedding orientations (Wilson, 1958). These channels were broad extending across a wide floodplain of low relief. Harshbarger and others (1957) reported eolian sandstones within some of these fluvial sequences.

KAYENTA FORMATION

Regional Stratigraphy

The Lower Jurassic Kayenta Formation comprises a heterogeneous assemblage of coarse- and fine-grained siliciclastics and minor carbonates that crop out in western Colorado, southern Utah, and northern Arizona. Contact with the underlying Moenave Formation ranges from gradational to sharp and even to highly irregular (Wilson, 1958). The Kayenta is over 400 m thick at Cedar City, 230 m at Zion, Utah, and less than 200 m near Tuba City, Arizona (Figure 2). The variations in thickness are due to intertonguing with the overlying Navajo Sandstone.

Based on exposures in north-central and northeastern Arizona, Harshbarger and others(1957) divided the Kayenta into the two generalized facies. Their sandy facies is best developed near Kayenta, Arizona (Figure 1) where it consists of reddish brown, lenticular, lithic, and feldspathic sandstone interbedded with subordinate amounts of grey to red mudstone. Southwest and west of Kayenta the sandy facies grades laterally into a silt-dominated facies (Figure 2) consisting of reddish-brown siltstone and claystone with minor sandstone. Thin grey limestone beds are locally abundant.

The Kayenta and overlying Navajo Sandstone intertongue throughout the area. Two tongues of each formation are recognized in southwestern Utah. The Shurtz Sandstone tongue of the Navajo lies between 125 and 250 m below the base of the main body of the Navajo (Averitt and others, 1955). The Lamb Point tongue of the Navajo (Wilson, 1958) occurs within the upper 65 m of the Kayenta (Figure 2). Both tongues thin and eventually disappear into the main body of the Kayenta to the south and southwest. The Cedar City and the Tenney Canyon tongues of the Kayenta overlie, respectively, the Shurtz and Lamb Point tongues of the Navajo (Averitt and others, 1955; Wilson, 1958). Both Kayenta tongues thicken as the Navajo tongues thin. Numerous smaller scale intertonguing relationships occur between the Navajo and Kayenta Formations throughout the southern margin of the Colorado Plateau.

Facies

At least six facies occur within the main body of the Kayenta and in the intertonguing interval with the Navajo. These include: large-scale, planar-tabular and trough cross-bedded sandstone, ripple and horizontally laminated sandstone and siltstone, interbedded sandstone-mudstone, mudstone, and cryptalgal laminated and structureless carbonate. The large-scale cross-bed sets occur in the upper part of the formation. Deformation structures are common within these sets. The smaller scale cross- and ripple-bedded sequences occur throughout the Kayenta. The heterolithic facies of sandstone-mudstone and mudstone facies decrease in abundance upwards and the carbonate facies become more abundant. Sandstone geometries are mainly lenticular although the large cross-bedded sets are persistent for considerable distances.

Paleogeography and Provenance

The Kayenta Formation was deposited on a broad tectonically stable alluvial plain. Streams flowed toward the west and southwest and hence parallel the drainage systems of the Moenave. Mineralogy of the sandstones indicates both extrabasinal crystalline sources in the Uncompahgre Upwarp and intrabasinal sedimentary sources composed of Mesozoic and Paleozoic rocks.

Meandering stream systems are represented by lateral accretion sets composed of medium- grading into small-scale trough cross-stratification (Allen, 1965; Middleton and others, 1981). These sequences have a channeled base, contain fragments of underlying deposits, fine upwards, and generally are associated with thick mudstone sequences of overbank origin. In the upper part of the Kayenta, thin calcretes and root tubules indicate pedogenic processes (Leeder, 1975; Bown and Kraus, 1981). Braided stream deposits comprise cosets of medium-scale trough and planar-tabular cross-bedded and horizontally stratified sandstone. The braided sequences are similar to those reported from modern and ancient linguoid and transverse bars (Miall, 1977).

Large-scale cross-stratified sandstones occur within the intertonguing interval and contain smaller-scale features such as sandflow and grainfall laminae and climbing translatent strata that are indicative of eolian sedimentation (Hunter, 1981). These sets are the deposits of large dunes that migrated toward the south and southeast and reflect the onset of desert conditions associated with the overlying Navajo Sandstone. Interbedded with the dune deposits are wadi and interdune sediments composed of horizontally bedded fine siliciclastics, algal-laminated carbonates and mudcracked surfaces (Middleton and others, 1981). Such sequences are well documented from modern interdunal areas periodically subjected to flooding (Glennie, 1970).

The reason for the increasing aridity associated with the Kayenta-Navajo transition is unclear. Studies by Stanley and others (1971) suggest that no major orographic barrier existed to the west to promote this change. Other possibilities that must be considered to explain the change from dominantly fluvial to eolian styles of sedimentation are base level and regional climate change.

EARLY MESOZOIC EOLIAN TRANSITION FROM CRATONAL MARGIN TO OROGENIC-VOLCANIC ARC

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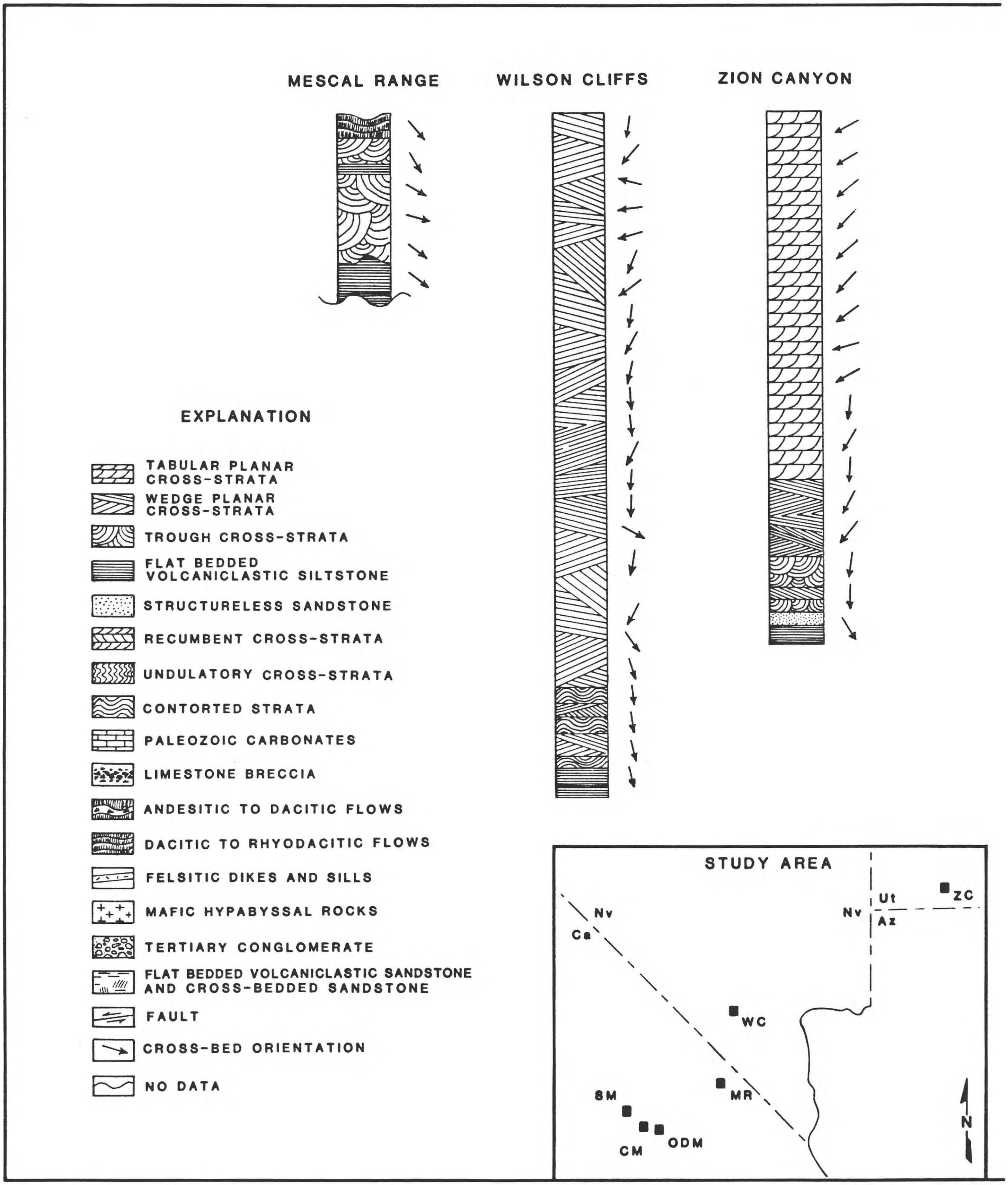
INTRODUCTION

The Navajo and Aztec Sandstones form a wedge of predominantly cross-bedded sandstone that thickens westward from a zero edge in western Colorado to over 650 m in western exposures along the eastern edge of the foreland fold and thrust belt in southern Nevada. Westward into the Mojave Desert of California, it becomes increasingly structurally dismembered or absent by erosion. Coinciding with this change in physiographic and structural expression, the Aztec Sandstone and underlying and overlying rocks change character as well. In the


MESCAL RANGE
WILSON CLIFFS
ZION CANYON
EXPLANATION
TABULAR PLANAR CROSS-STRATA
WEDGE PLANAR CROSS-STRATA
TROUGH CROSS-STRATA
FLAT BEDDED VOLCANICLASTIC SILTSTONE
STRUCTURELESS SANDSTONE
RECUMBENT CROSS-STRATA
UNDULATORY CROSS-STRATA
CONTORTED STRATA
PALEOZOIC CARBONATES
LIMESTONE BRECCIA
ANDESITIC TO DACITIC FLOWS
DACITIC TO RHYODACITIC FLOWS
FELSITIC DIKES AND SILLS
MAFIC HYPABYSSAL ROCKS
TERTIARY CONGLOMERATE
FLAT BEDDED VOLCANICLASTIC SANDSTONE AND CROSS-BEDDED SANDSTONE
FAULT
CROSS-BED ORIENTATION
NO DATA
STUDY AREA
Nv
Ca
Nv
Ut
Az
ZC
WC
MR
SM
CM
ODM
N

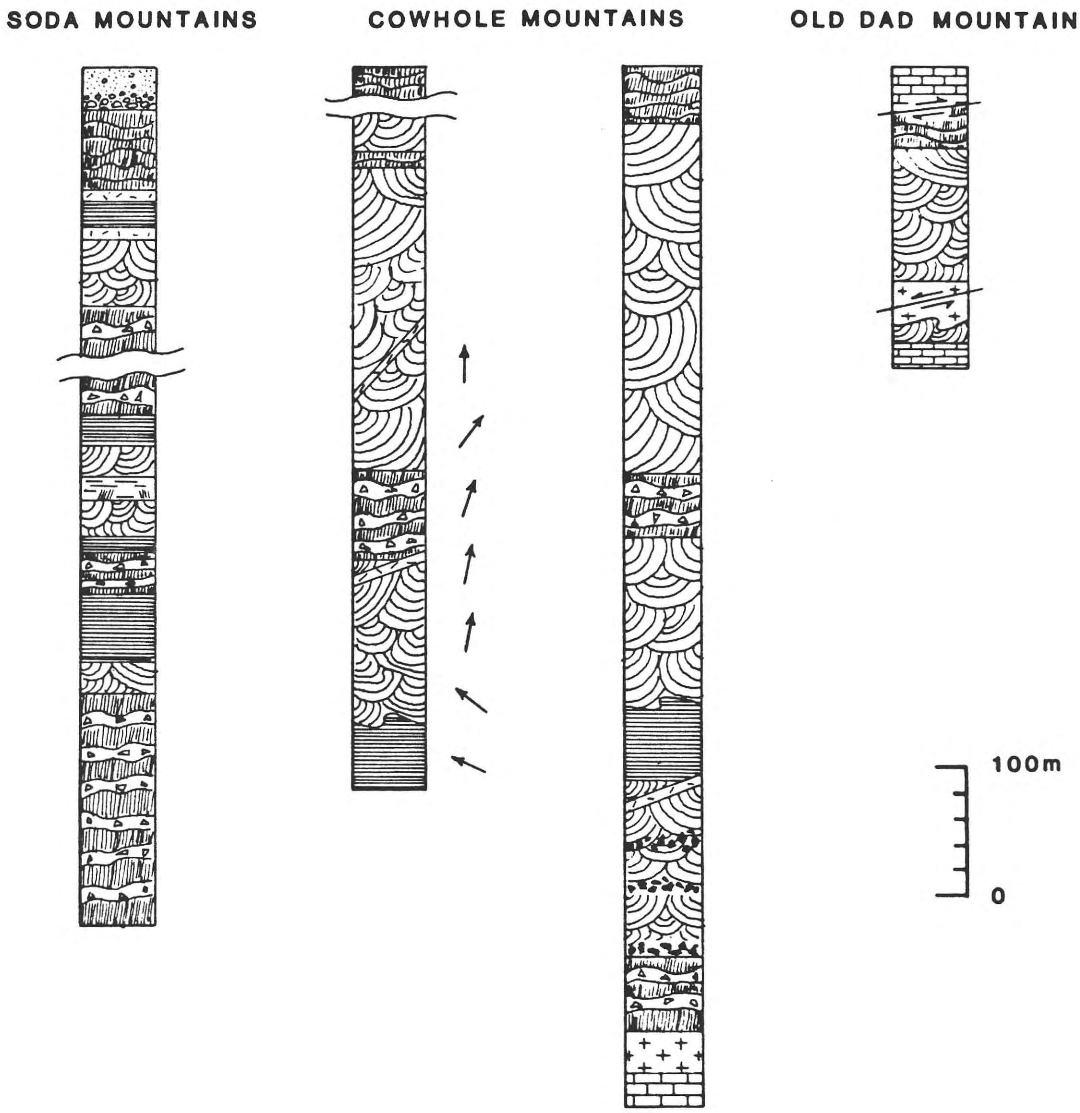


Figure 4. Measured sections of Navajo and Aztec Sandstones showing distribution of sedimentary structures and orientation of crossbeds in 30 m vertical increments. Arrows indicate vector-resultant dip directions. North at top of the page. Location of sections in inset map.

following discussion, these two regions are treated separately. Localities referred to in the text are illustrated in Figure 4.

SOUTHWESTERN UTAH AND SOUTHERN NEVADA

In southwestern Utah, the Navajo Sandstone is separated from the overlying Temple Cap Sandstone and marine Carmel Formation by regional unconformities (Peterson and Pipiringos, 1979). Where the contact is stratigraphic in southern Nevada, the Aztec Sandstone is unconformably overlain by continental strata of Cretaceous and Tertiary age. Elsewhere, it is in thrust contact beneath rocks of early Paleozoic age.

In Utah, the Navajo is the uppermost of three formations included in the Glen Canyon Group. In southwestern Utah, in ascending order the underlying formations are the Moenave and Kayenta. The Navajo Sandstone interfingers southwestward with the Kayenta Formation. Westward into southern Nevada, the Moenave and Kayenta become indistinguishable but form a distinct statigraphic unit unconformably overlying the Chinle Formation as far west as the Wilson Cliffs (Wilson and Stewart, 1967). In the Valley of Fire and the Wilson Cliffs, unnamed tongues of basal Aztec Sandstone interfinger with the Kayenta-Moenave equivalent.

Wilson and Stewart (1967) identified a conglomerate which, from place to place, overlies the Chinle Formation at the base of the Moenave-Kayenta equivalent. The conglomerate reaches its maximum thickness in the Wilson Cliffs. It is composed of pebbles and granules of limestone, clastic sedimentary rocks, chert, and volcanics (Figure 5). At this locality, it appears likely the conglomerate has a nearby western Precambrian and Paleozoic sedimentary and an early Mesozoic volcanic provenance.

Within the Navajo and Aztec Sandstones of southern Utah and southern Nevada, large-scale trough, wedge-planar, and tabular-planar crossbedding are the the most conspicuous sedimentary structures. Although less conspicuous, other sedimentary features are important to the interpretation of the depositional environment of the Navajo and Aztec Sandstones. These features include: Limestone lenses; desiccation-cracked, horizontally stratified sandstone and horizontal erosion surfaces; and contorted stratification and structureless sandstone (Figure 4). These features are indicative of a shallow groundwater table during deposition of that part of the Navajo and Aztec Sandstones which contains them (Marzolf, 1978; 1982a; b). In southeastern Utah, features of groundwater origin are distributed throughout the Navajo Sandstone but, in southwestern Utah and southern Nevada, they are restricted to the lower 100 to 85 m, respectively.

A change in wind direction from northwesterly to northeasterly during deposition of the Navajo and Aztec Sandstones is indicated by change in crossbed orientation (Figure 4). Crossbed-dip direction rotates clockwise from 152° at the base of the Navajo in Zion Canyon to 228° at the top and, in the Wilson Cliffs, from 169° at the base to 226° at the top of the Aztec. The similarity of orientations between these two localities also implies little or no tectonic rotation of the Wilson Cliffs relative to the Colorado Plateau. This conclusion is supported by preliminary paleomagnetic measurements on the sub-Aztec and sub-Navajo redbeds (E. M. Shoemaker, oral communication, 1982).

Maximum and minimum ages of the Navajo and Aztec Sandstones are based on data from widely separated localities. In southeastern Utah, the evidence of terrestrial vertebrate fossils in both the Kayenta Formation and the lower part of the Navajo Sandstone is controversial (Lewis and others, 1961; Galton, 1971; Olsen and Galton, 1977). Late Triassic and Early Jurassic ages are tenable. In southwestern Utah, palynomorphs in the Whitmore Point Member of the Moenave Formation indicate an Early Jurassic age - probably Sinemurian to Pliensbachian (Cornet *in* Peterson and Pipiringos, 1979). An upper age limit of late middle Bajocian for the Navajo Sandstone in southern Utah is based on marine invertebrates in the limestone member of the Carmel Formation. The maximum and minimum ages at these localities serve as points in space and time between and beyond which the Navajo and Aztec Sandstones can be placed with some latitude.

THE MOJAVE DESERT

Hewett (1956) extended the Aztec Sandstone and underlying Chinle Formation to the Mescal Range in the Mojave Desert and Grose (1959) suggested that 7,000 ft. (2,121 m) of interbedded quartz sandstone and volcanics in the Soda Mountains represented a western facies of the Aztec Sandstone. Within the last five years, cross-bedded quartz sandstone and interbedded volcanics of demonstrable Triassic or Jurassic age have been recognized further and further west in the Mojave Desert of California (Dunne, 1977; Novitski-Evans, 1978; Miller, E. L., 1978; Miller and Carr, 1978; Cameron and others, 1979) and further south in the

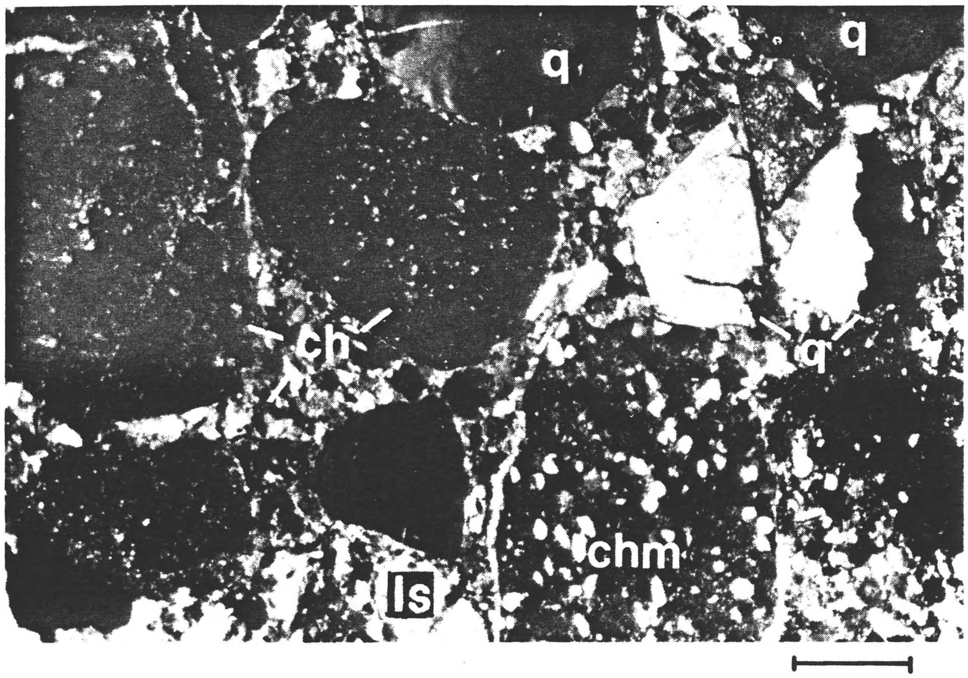


Figure 5. Photomicrograph of pebble conglomerate at the base of the Kayenta-Moenave equivalent in Wilson Cliffs: ch, chert; chm, cherty mudstone; ls, limestone; and q, quartz. Bar scale equals .5 mm. (Photo by Don Zaengle.)

eastern Mojave and south-central Arizona (Bilodeau and Keith, 1979; Haxel and others, 1980; Hamilton, 1982; Miller and others, 1982). These rocks are presumed to represent further extensions of the Aztec and Navajo Sandstones into an early Mesozoic orogenic-volcanic arc.

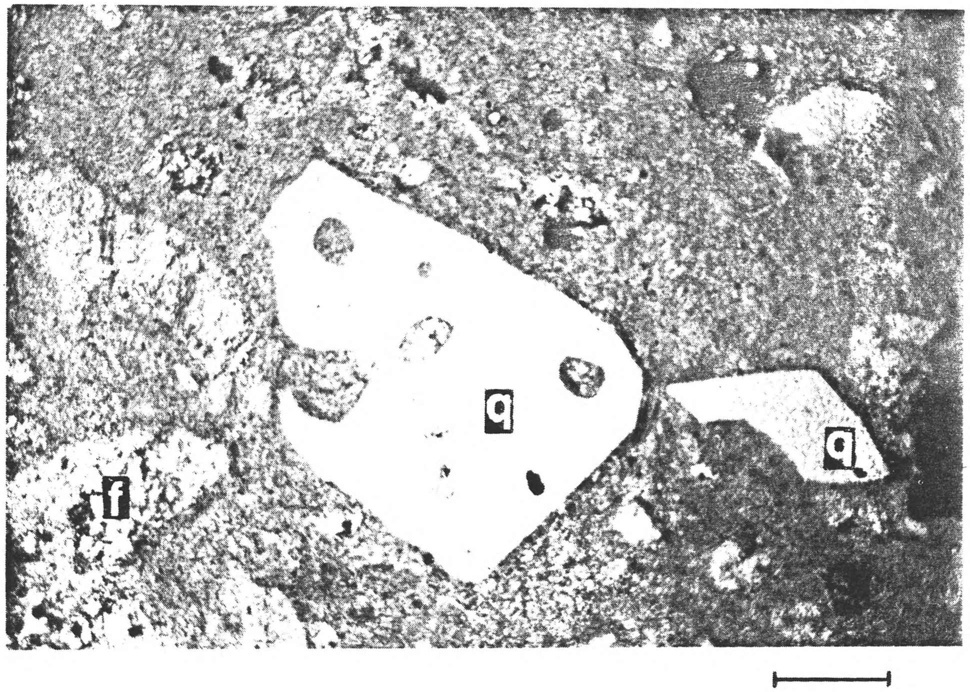
In the eastern Mojave Desert, in the areas to be visited on this field trip, the Aztec Sandstone interfingers with intermediate to silicic volcanics and is overlain by predominantly silicic volcanics. It overlies rocks of varying lithology and age.

In the Mescal Range, cross-bedded Aztec Sandstone overlies redbeds assigned by Hewett (1956) to the Chinle Formation and is overlain by the Delfonte volcanics (Burchfiel and Davis, 1971). Within the cross-bedded sequence, a single thin bed of siltstone containing volcanic clasts is the only hint of the volcanics interbedded with the Aztec Sandstone to the west. Dinosaur tracks within the Aztec Sandstone at this locality are the only dinosaur tracks known in California (Reynolds, in this book).

The redbeds consist of quartz-rich siltstone containing pockets and lenses of coarse sand-size clasts of porphyritic volcanics. Marzolf (1980) suggested these strata correlated with a redbed facies in the lower part of the Aztec Sandstone in the Cowhole Mountains. They may equally well correlate with the Kayenta-Moenave equivalent of the Wilson Cliffs. Whichever of these two interpretations is correct, they probably should not be assigned to the Chinle Formation. The anomalously thin Aztec Sandstone in the Mescal Range is believed to be the result of post-Aztec, pre-Delfonte erosion (Marzolf, 1982a). In the southeast corner of the Mescal Range, fluvial cratonal rocks appear to be in contact with volcanics of the orogenic-volcanic arc.

Farther to the west at Old Dad Mountain and in the Cowhole Mountains, the Aztec Sandstone rests unconformably on deformed Paleozoic carbonates or Triassic volcanics. The basal part of the section contains one or more horizons of limestone breccia consisting of angular blocks of Birdspring limestone in a sand matrix. At the northern end of the Cowhole Mountains the limestone breccia is overlain by redbeds of quartz siltstone containing coarse sand-size volcanic clasts. Farther to the south the redbeds rest on cross-bedded quartz sandstone which, in turn, overlies and contains lenses of limestone breccia. The basal breccia rests on volcanics which have been cut by numerous dikes (Figure 4).

Throughout the Cowhole Mountains, the redbeds are overlain by large-scale cross-bedded sandstone interbedded with and overlain by volcanic



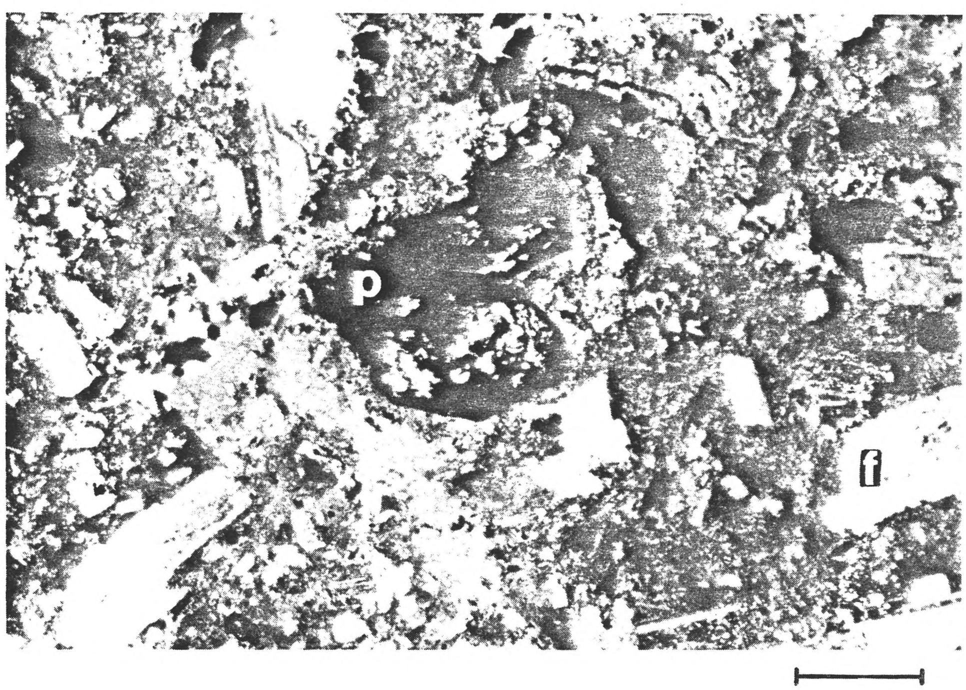


Figure 6. Photomicrographic of flow rocks interbedded with the Aztec Sandstone in the Cowhole Mountains. Top) q, volcanic quartz and f, calcite-replaced feldspar phenocrysts in altered glassy matrix. Quartz latite from flows overlying Aztec Sandstone; Bottom) Phenocrysts of iron oxide pseudomorphed after pyroxene (p) and plagioclase (f) in trachyandesite of lower flow unit. C is calcite. Bar scale equals .5 mm. (Photo by Don Zaengle.)

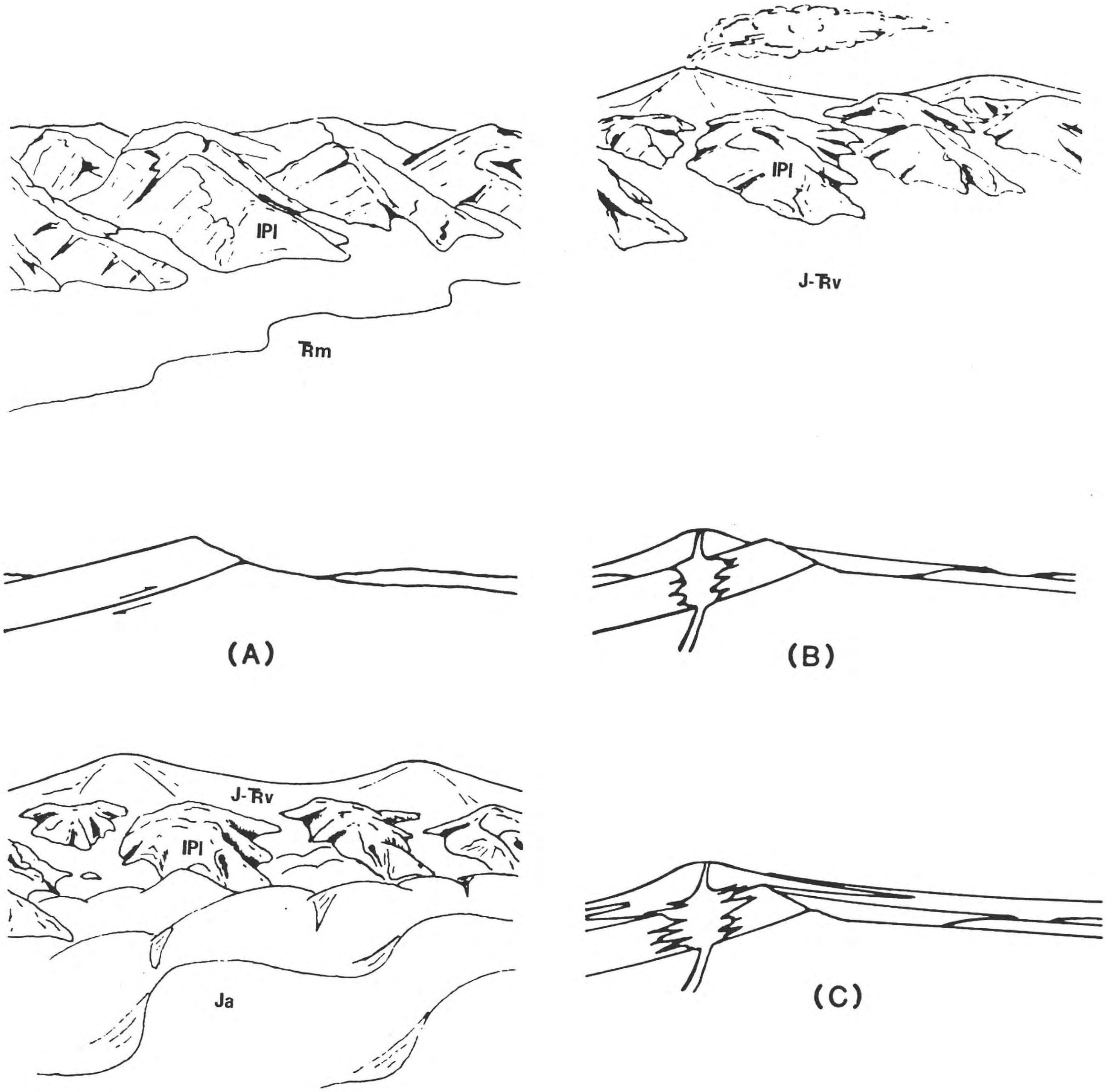


Figure 7. Postulated sequence of early Mesozoic events in the Cowhole and Soda Mountains area. A) Moenkopi Formation (TRm) has been eroded from deformed and uplifted Paleozoic limestone (IPl). Timing of thrusting can only be bracketed as post-Birdspring and pre-Aztec; B) Volcanics (J-TRv) fill topographic lows and bury Moenkopi Formation; C) Aztec Sandstone (Ja) covers earlier volcanics and Paleozoic limestone and interfingers with coeval volcanics.

flows (Figure 4, 6a-b). The lower flows are andesitic to dacitic and the upper and overlying flows are dacitic to latitic.

In the Soda Mountains, Grose (1959) described over 7,000 ft (2,120 m) of interbedded, cross-bedded quartz sandstone and andesitic to dacitic volcanics believed to be correlative with the Aztec Sandstone. In the southeastern corner of the eastern Soda Mountains, the Aztec Sandstone consists of light yellowish-brown cross-bedded, quartz sandstone and flat-bedded, grayish-purple quartz sandstone containing prophyritic volcanic clasts from coarse-sand to cobble size and is interbedded with prophyritic dacitic flows (Marzolf, 1981). The sequence is overlain conformably by porphyritic quartz latite with quartz and feldspar phenocrysts. At the base of the interbedded sandstone and volcanics, a volcaniclastic sedimentary breccia overlies a multiple-faulted sequence of andesitic to dacitic flows, flow breccias, and hypabyssal rocks of uncertain thickness. Near the bottom of this sequence, slivers of metamorphosed limestone and calc-silicates are enclosed in the hypabyssal rocks. At the base of the sequence, volcanic and hypabyssal rocks are in fault and intrusive contact with steeply-dipping Paleozoic carbonates and calc-silicates.

In the western Soda Mountains, the interbedded cross-bedded, quartz sandstone and volcanics overlie shallow-water marine limestones and siltstones assigned by Grose (1959) to the Lower Triassic Moenkopi Formation.

The relationships of Aztec Sandstone to interbedded volcanics and underlying rocks from the Mescal Range to the Soda Mountains indicate Aztec sands were being blown into a developing volcanic terrane. Relationships in the Cowhole Mountains indicate a surface of relief of at least 226 m had been developed on thrust-faulted Paleozoic carbonates prior to deposition of the Aztec Sandstone. Topographic lows were filled with volcanic flows upon which the windblown sand was deposited. Topographic highs protruded above the volcanics to be buried directly by the Aztec Sandstone. The dark reddish-brown siltstone appears to be of fluvial origin. The volcanic clasts apparently were washed from the flanks of the developing volcanoes, perhaps as the result of orographic precipitation. These relationships are illustrated in Figure 7a-c.

In the western facies attempts to determine the age of the Aztec Sandstone have been based on radiometric ages of interbedded or overlying volcanics. A minimum K/Ar date of 155 m.y. B.P. for the Delfonte Volcanics was obtained by Sutter (Novitsky-Evans, 1978). In southern Arizona, the oldest volcanics associated with possible Aztec Sandstone have been dated at 190 m.y. B.P. (Wright and others, 1981). Just across the border in northern Sonora, Mexico, Silver and Anderson dated similar volcanics at 180-185 m.y. B.P. (Anderson, oral communication, 1981).

The Navajo Sandstone and its lithologic equivalent, the Aztec Sandstone, are the youngest pre-Laramide formations that can be traced westward from the early Mesozoic cratonal margin of the present-day Colorado Plateau into the developing orogenic-volcanic arc. Deposited by northwesterly winds, sediment transport was independent of the newly developing paleoslope which limited the westward dispersal of Late Triassic and Early Jurassic (?) cratonally-derived fluvial sediments. The absence of eolian sand in the Late Triassic-Early Jurassic (?) sequences in the northern Mojave Desert and north of the Garlock Fault may well reflect the northwestward limit of eolian continental sedimentation to the east of the volcanic arc.

JURASSIC TRACKWAYS IN THE MESCAL RANGE, SAN BERNARDINO COUNTY, CALIFORNIA

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The only known dinosaur tracks in California occur in the Mescal Range of northeastern San Bernardino County. They are found in the Aztec Sandstone which overlies a maroon sandstone.

At this locality, the Aztec Formation may be divided into three mappable lithologies. The lowest unit is a red silty sandstone; the middle unit is a resistant white to variegated sandstone; the upper unit consists of alternating red and white sandstones. A thin dark volcanic bed is exposed 13.5 m below the contact of the uppermost unit with the overlying Delfonte Volcanics.

Trackways in the Aztec Sandstone are known only from the resistant middle unit. They consist of three types of quadruped tracks from 1.3 to 2.5 cm in size and three types of tracks left by bipeds (Figures 8a, 8b, and 9). The latter may be attributed to Coelurosaurs of a size similar to a small ostrich. Marzolf (1982a) suggests a possible range between early Pliensbachian and late middle Bajocian age for the Aztec and Navajo Sandstones. Work underway may provide further refinements of dates.

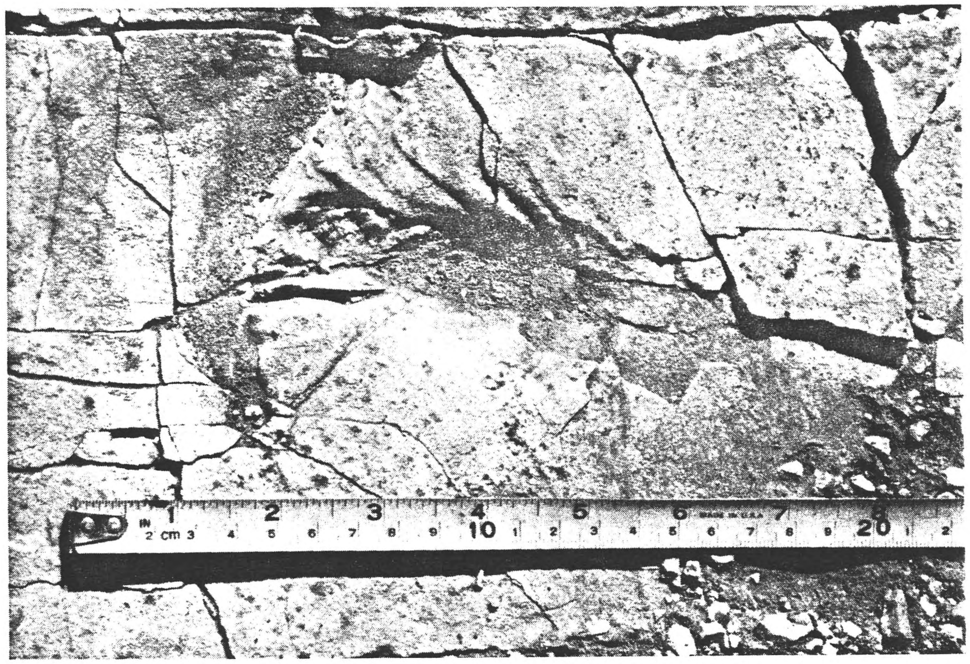

Figure 8a. Quadruped tracks in the Aztec Sandstone (2.5 cm diameter).

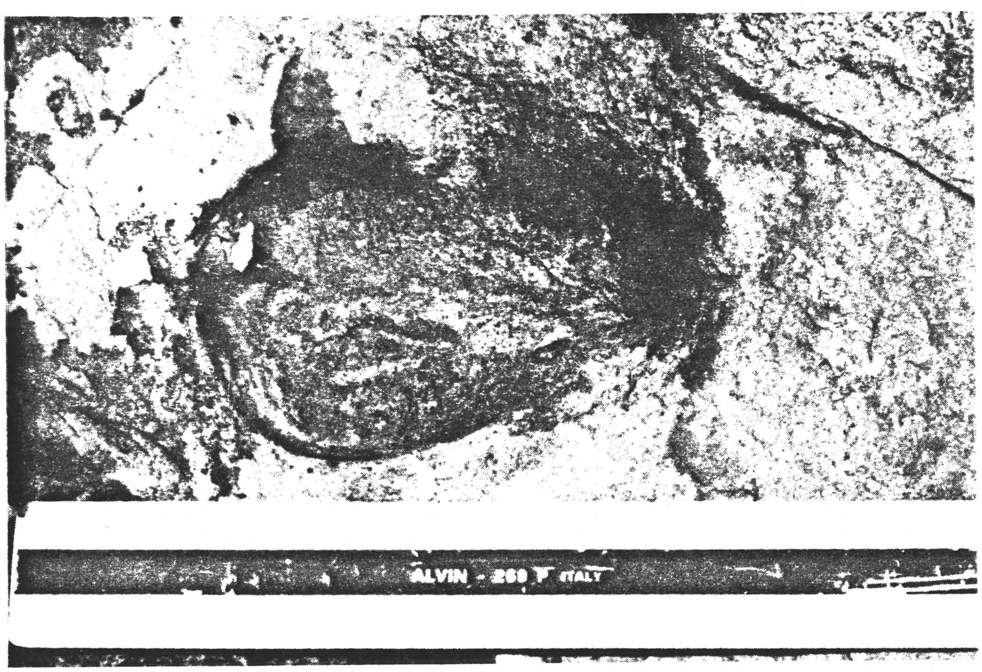

Figure 8b. Impression of three-toed Coelurosaur, hind foot (12.7 cm long).

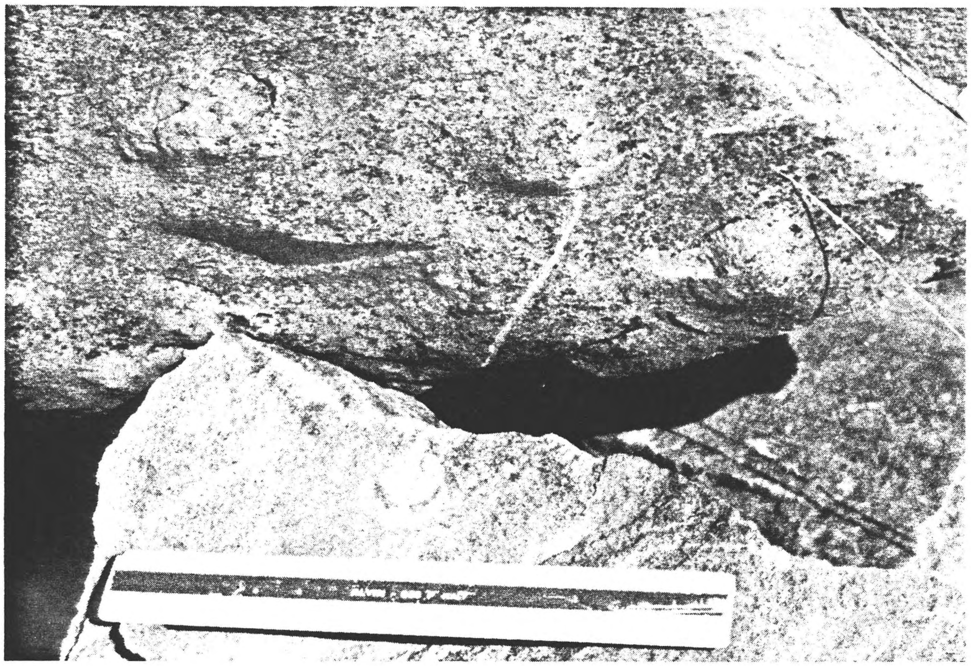

Figure 9. A narrow impression of a Coelurosaur, hind foot (8.9 cm long).

PERMOTRIASSIC ROCKS IN THE NORTHERN MOJAVE DESERT REGION AND THEIR TECTONIC SIGNIFICANCE

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INTRODUCTION

A complete section of Early Triassic Moenkopi Formation can be traced from the Colorado Plateau, through Frenchman Mountain, into the Spring and Clark Mountains in southern Nevada and eastern California (Figure 10; Hewett, 1956; Stewart and others, 1972a; Burchfiel and others, 1974). Rocks equivalent to the Moenkopi Formation can be traced westward to the Soda Mountains (Figure 10; Grose, 1959). A recently mapped sequence of calcsilicate metasediments, metaconglomerate, and limestone forms the base of Moenkopi equivalent rocks in the Soda Mountains (Burchfiel and others, 1980). Lithologically similar rocks occur at the base of probable Triassic sequences farther west in the Mojave Desert. These western occurrences constitute an overlap assemblage resting on a belt of metamorphosed and deformed Paleozoic miogeoclinal/transitional rocks, displaced eugeoclinal rocks, and Permo-Triassic plutonic rocks (Burchfiel and others, 1980). Deformation, plutonism, and juxtaposition of eugeoclinal and miogeoclinal rocks are probably the result of Permo-Triassic truncation of the southwestern Cordilleran orogen and the construction of an Andean arc along the truncated margin (Hamilton and Myers, 1966; Burchfiel and Davis, 1972, 1981; Miller, C. F., 1978; Miller and Cameron, 1982). Brief descriptions of probable Permo-Triassic sequences in the northern Mojave Desert are given below.

DESCRIPTIONS

A typical section of Moenkopi Formation is present in southern Nevada and eastern California. Recent mapping in the Clark Mountains reveals several lenses of conglomerate, containing clasts of Precambrian gneiss, granite, and quartzite as well as rare andesite, in the upper Moenkopi Formation (Figure 10; Walker and others, in press). This con-

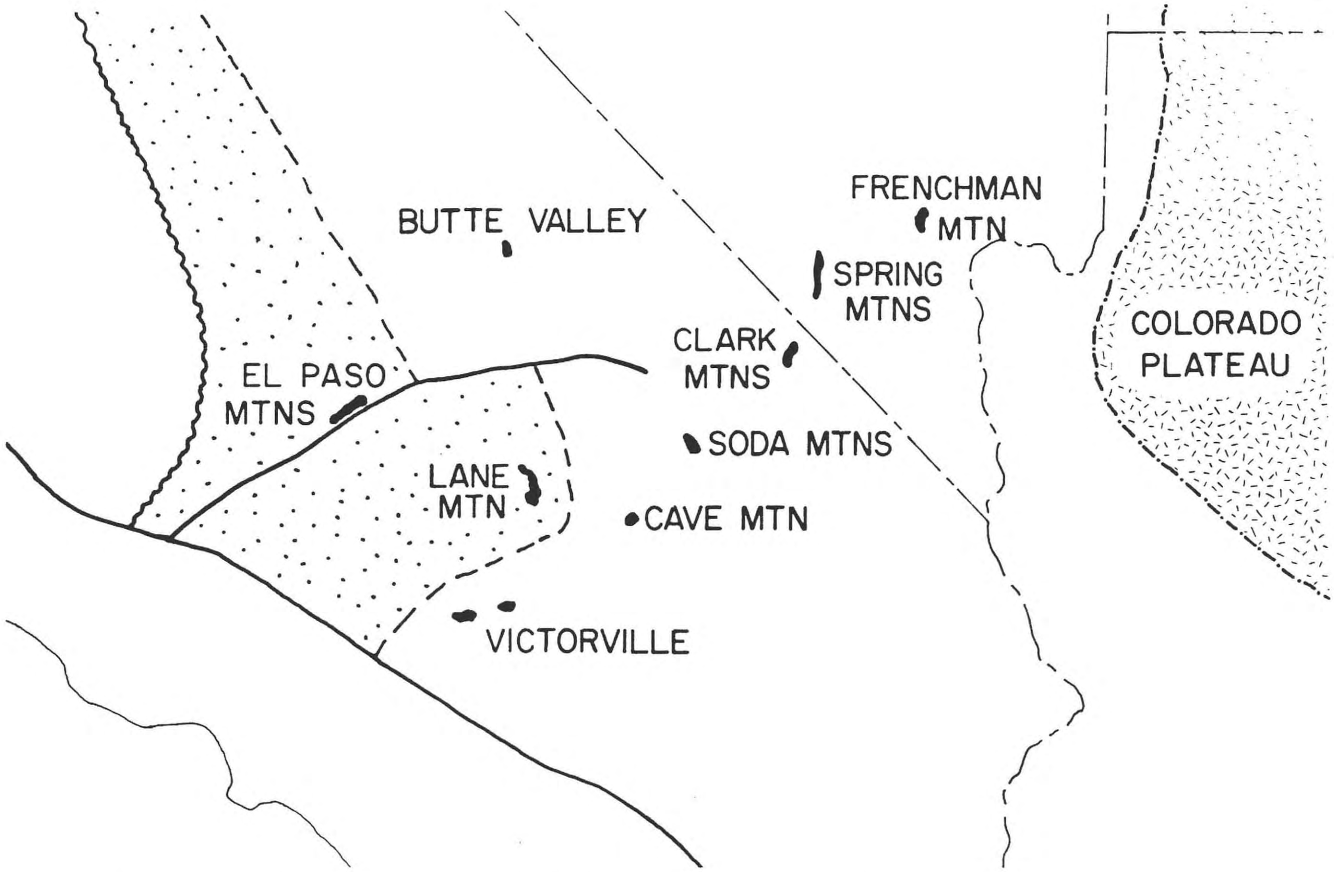

Figure 10. Location map for Permo-Triassic sections in the northern Mojave Desert. Colorado Plateau shown in granite pattern; area of displaced eugeoclinal rocks is stippled pattern. Dashed line shows approximate eastern boundary of the displaced assemblage. Wavy line is approximate line of Permo-Triassic truncation. Also shown are the San Andreas and Garlock Faults. State borders are in light dashed lines.

glomerate indicates Early Triassic deformation and igneous activity extends farther to the east than previously recognized.

Grose (1959) mapped a sequence of argillite, cherty limestone, and shale on Spectre Spur in the Soda Mountains which he considers equivalent to the Moenkopi Formation (Figure 1). The Spectre Spur rocks are lithologically similar to the Moenkopi Formation and contain gastropods dated as Early Triassic. Burchfiel and others (1980) report a previously unrecognized sequence of metasiltite, limestone, marble, and metaconglomerate that unconformably overlies the Lower Permian Bird Spring Formation on Spectre Spur, and apparently forms the base of Moenkopi equivalent rocks. Clasts in the metaconglomerate are mainly limestone, siltstone, chert; rare gneissic clasts are also present. The Moenkopi equivalent rocks are disconformably overlain by andesitic flows which contain lenses of Jura-Triassic Aztec Sandstone (Grose, 1959).

Early Mesozoic rocks in the Victorville area (Fairview Valley Formation) and at Cave Mountain unconformably overlie deformed and metamorphosed Paleozoic miogeoclinal/transitional carbonate rocks (Figure 10; Miller, 1981; Cameron and others, 1979). The Fairview Valley Formation comprises 1000 m of locally derived(?) conglomerate, silty limestone and calcareous siltstone in an interfingering alluvial fan and marine sequence. These rocks overlie a monzonite pluton dated at 233 ± 13 m.y. B.P. ($^{40}Ar/^{39}Ar$, Miller and Sutter, in press). Rocks at Cave Mountain consist of metasiltite, calcsilicate metasediments, marble, metaconglomerate, and arkosic metasandstone (Figure 10; Cameron and others, 1979). In both areas, the sedimentary sequences are overlain by volcanic and/or volcaniclastic sequences containing quartzite correlated with the Aztec Sandstone.

Similar rocks of probable Triassic age also occur in the El Paso Mountains (Bond Buyer sequence; Christiansen, 1961; Carr and others, 1981) and at

Lane Mountain (Noble Well Formation; McCulloh, 1952), and apparently unconformably overlie deformed and displaced Paleozoic eugeoclinal strata (Figure 10). The Bond Buyer sequence consists of calcsilicate metasediments, with interbedded quartzite pebble conglomerate, and it is intruded by latest Permian polutonic rocks (Carr and others, 1982). The Noble Well Formation comprises metasandstone, metaconglomerate, marble, and metasiltite, and has yielded Triassic fossils (McCulloh, unpublished data).

A sequence of Triassic age siltstone and limestone is exposed in Butte Valley in the southern part of the Panamint Range (Figure 10; Butte Valley Formation; Johnson, 1957). These rocks have been correlated with Moenkopi equivalent rocks in the Soda Mountains based on lithologic similarities and the presence of Triassic gastropods in each sequence (Grose, 1959; Burchfiel and others, 1980). The Butte Valley Formation is unconformably overlain by a volcaniclastic and limestone pebble conglomerate that grades upward into andesitic flows and breccias of unknown age. The base of this formation is not exposed. The Butte Valley Formation ties orogenic Triassic sequences in the Mojave to miogeoclinal(?) rocks to the north.

Thus, early Mesozoic rocks in the northern Mojave Desert comprise a Triassic overlap sequence, probably in part correlative to the Moenkopi Formation, which overlies deformed Paleozoic miogeoclinal and eugeoclinal rocks to the west and undeformed Palezoic sequences to the east. A transition from sections in the Soda Mountains into similar rocks in the southern Panamint Range is present.

LOWER TRIASSIC MARINE SEDIMENTARY ROCKS IN EAST-CENTRAL CALIFORNIA

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INTRODUCTION

Lower Triassic marine sedimentary rocks crop out in east-central California north of the Garlock fault in a relatively narrow belt that extends discontinuously from Union Wash to Butte Valley in the Panamint Range (Figure 11). Lower Triassic sedimentary rocks in the south-central and southern Inyo Mountains, east of Darwin, and in the eastern Argus Range are included in this study (Figures 11 and 12).

UNION WASH

The only complete section of Lower Triassic marine rocks known in east-central California is at Union Wash in the south-central Inyo Mountains. On the north side of Union Wash, the Lower Triassic rocks overlie shallow water Upper Permian rocks, apparently paraconformably. On the south side of the Wash, where the complete section is about 530 m thick, they rest upon Lower Permian rocks. The Triassic sequence begins with a bioturbated, sparsely pebbly, silty limestone of probable marginal marine origin. These rocks grade upward into an argillite and very fine-grained limestone sequence which compose most of the remainder of the unit. Two major ammonoid bearing limestones are present in this sequence (Figure 12, Table 1), one about 100 m above the base, which is early Smithian in age, and the other almost at the top (about 475 m above the base of the section), which is late Spathian in age (Silberling and Tozer, 1968). The associated conodont faunas confirm these age assignments (Figure 12, Table 1). This marine sequence probably was deposited in relatively deep water far from shore. The Lower Triassic Marine rocks are overlain by a nonmarine conglomerate of uncertain age that has yielded a pebble of limestone consisting of coated grains. This suggests that originally a very shallow top to the Lower Triassic marine unit had been present, that these upper beds subsequently had been eroded and that clasts from them were incorporated into the overlying nonmarine conglomerate. The nonmarine unit is overlain by a very thick volcanic unit, also of uncertain age.

CERRO GORDO

The lower marine sequence also crops out along the Cerro Gordo road in the southern Inyo Mountains. Here, the section is highly faulted and folded so that no complete stratigraphic sequence

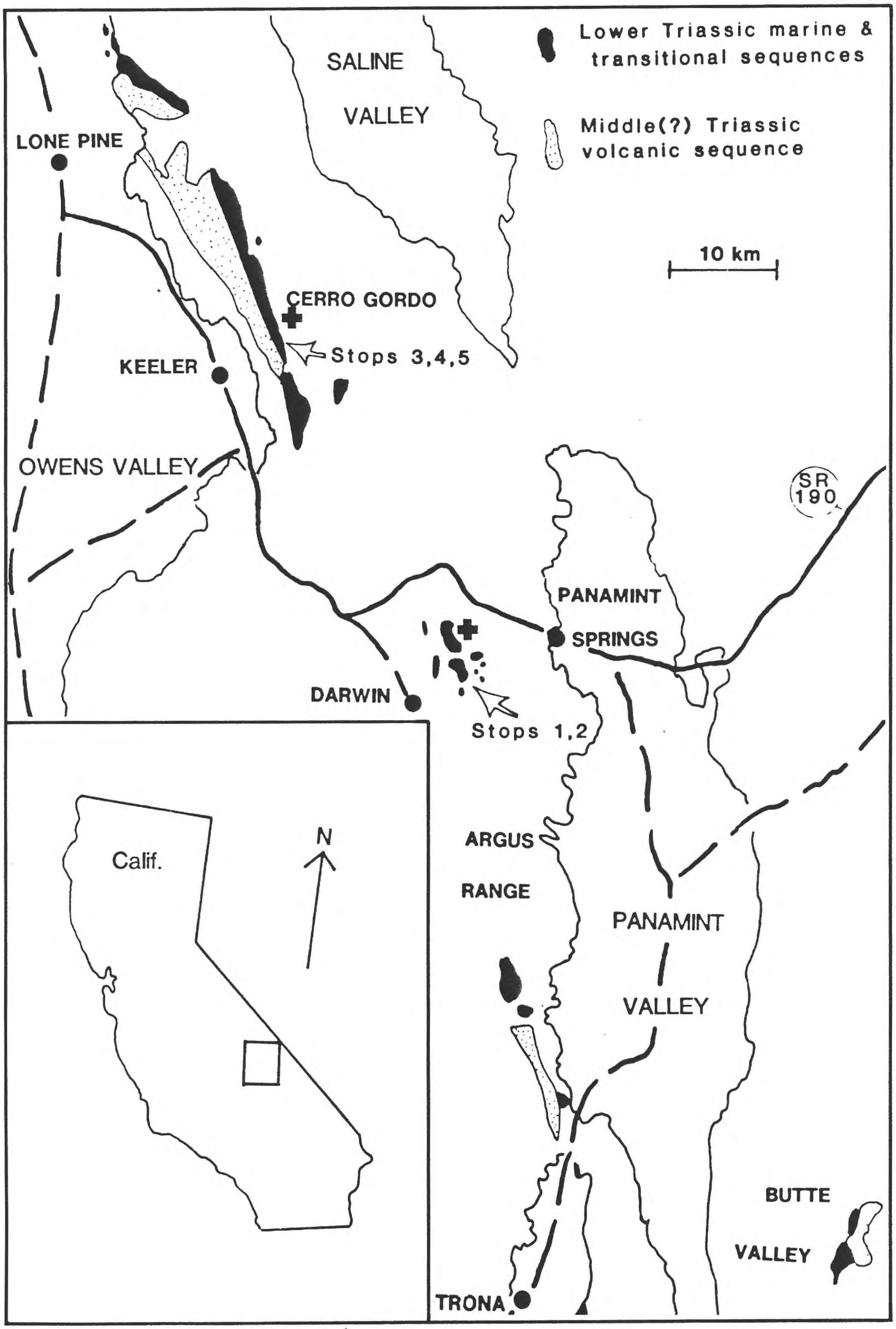

Figure 11. Location of Lower Triassic sedimentary rocks north of the Garlock fault, east-central California (positions of outcrops in the Argus Range and southern Inyo Mountains are from unpublished maps of Paul Stone).

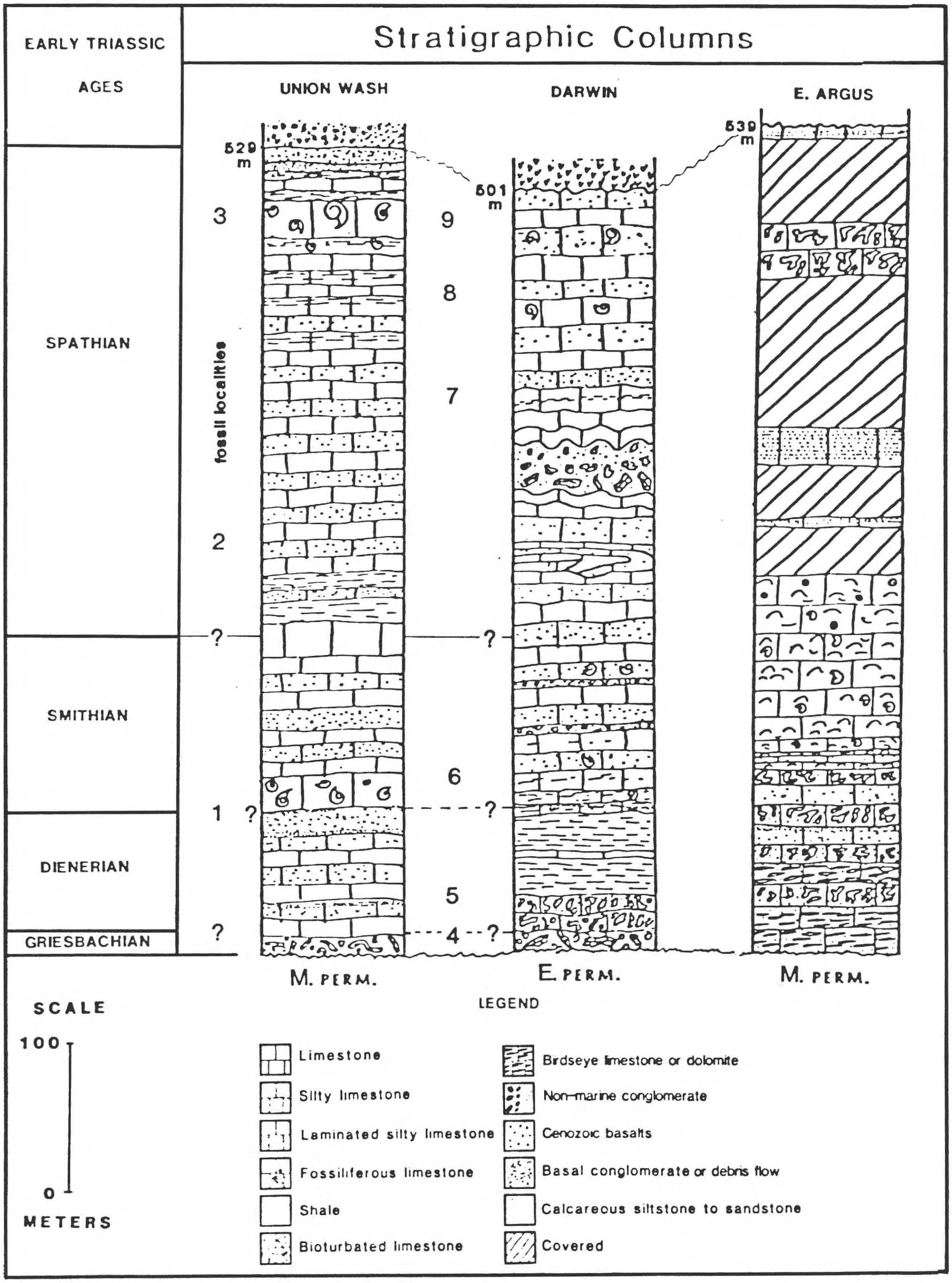

Figure 12. Stratigraphic sections of Lower Triassic rocks in east-central California (numbers adjacent to columns refer to numbered fossil lists in Table 1).

TABLE 1.

Locality No.	FAUNA	AGE (Stage)
Union Wash		
1.	*Meekoceras gracilitatis* * *Neospathodus conservativus Neospathodus* sp. *Ellisonia triassica Hadrodontina* sp.	Smithian
2.	*Neospathodus homeri Ellisonia triassica Ellisonia gradata Neospathodus* sp.	Spathian
3.	*Neopopanoceras haugi**	Late Spathian
Darwin Wash		
4.	*Claraia stachei***	Griesbachian
5.	*Neogondolella carinata Neospathodus dienéri Xaniognathodus curvatus Ellisonia triassica Ellisonia* sp.	Dienerian
6.	*Meekoceras* ? *	Smithian?
7.	*Neospathodus* sp. *Neogondolella* sp. *Ellisonia triassica*	Spathian
8.	*Neopopanoceras haugi**	Late Spathian
9.	*Neospathodus timorensis Neospathodus* sp. *Ellisonia triassica*	Late Spathian

*Ammonoid, ** Bivalve, all others Conodonts
Age determination based on work by Silberling and Tozer (1968), Collison and Hasenmueller (1978), and Clark and others (1979).

can be pieced together. However, the base of the Lower Triassic section is well exposed southeast of Cerro Gordo road where it can be seen to rest with angularity upon the Permo-Pennsylvanian Keeler Formation. The basal unit appears to be highly bioturbated and it contains many small gastropods suggesting deposition in a marginal marine setting. The four prominent dark blue limestone outcrops along the northwest side of the road, which have been called the upper reefy limestone by Merriam (1963), are thought to be repetitions of a single limestone sequence that presumably marks the top of the Lower Triassic marine unit. The lower beds of the upper reefy limestone, one of which is more than 2 m thick, appear conglomeratic, but are difficult to interpret. These rocks are overlain by shallow water, thin, fine-grained limestones that bear coated grains, ooliths, and numerous convex upward pelecypod shells with well formed geopetal structures. These beds are overlain by a very fine-grained carbonate with very fine stromatolite-like bedding and possible birdseye structures. This carbonate probably was deposited in a shallow subtidal to intertidal environment. Nonmarine, terrigenous beds followed by a thick volcanic sequence lie above the Lower Triassic marine rocks.

DARWIN CANYON

East of Darwin, an incomplete Lower Triassic sequence about 540 m thick is exposed. It rests with angularity upon Lower Permian turbidites and debris flows. The top of the Lower Triassic sequence has been eroded and covered. The sediment types, sedimentary structures, and fossils allow a reasonably close interpretation of the sequence of depositional environments represented. Early Triassic deposition began on an almost flat surface cut on gently folded Lower Permian rocks. In places, shallow channels filled with large clasts of Lower Permian rock are present. This intermittently developed conglomerate is overlain by a highly bioturbated, silty carbonate bearing tiny gastropods. This unit probably was deposited in a marginal marine environment. Water depth apparently then increased rapidly, so that by about 85 m above the base of the section, laminated radiolarian-rich, dark gray, cherty argillites and limestones were deposited. Interbedded with these presumed deepwater rocks are several limestone gravity-flow deposits, one containing slabs of limestone many meters in length. These rocks have been determined as Dienerian in age on the basis of conodonts (Figure 12, Table 1). Higher still in the stratigraphic section, a thick sequence of dark-colored, uniformly thin-bedded limestones followed by apparently disturbed limestones with obscure bedding are present. In the upper part of the section, the return of shallow shelf conditions is suggested by the presence of numerous pelecypods in a few of the beds. Conodonts show that these beds are Spathian in age (Figure 12, Table 1).

EASTERN ARGUS RANGE

The section on the east side of the Argus Range rests apparently paraconformably upon a supposed Upper Permian conglomerate. The top of the section is intruded and metamorphosed. This section apparently contains the shallowest water, most nearshore Lower Triassic rocks studied. The 140 m thick basal section of unfossiliferous carbonate with birdseye structure probably was deposited on the highest parts of a tidal flat. This unit is overlain by about 113 m of pelecypod-rich limestone probably deposited on a shallow marine shelf. The environmental significance of these rocks is unknown.

SUMMARY

Overall, the depositional pattern of the Lower Triassic marine rocks seems quite clear. Deposition apparently began everywhere in shallow, marginal marine environments. The water depth then increased, slowly in the Argus Range, but rapidly elsewhere to the northwest. Very fine-grained rocks deposited throughout the remainder of the Early Triassic gradually filled the known parts of the basin. The sections available for study, unfortunately are situated linearly preventing determination of the orientation of this Early Triassic depositional basin. This study, however, shows that (1) the section in the Argus Range contains a thick, shallow water sequence, (2) at Darwin Canyon, to the northwest, this part of the section consists of cherty, radiolarian-rich beds and limestone gravity-flow deposits representing much deeper water, and (3) still farther west, at Union Wash, the section lacks coarse-grained sediment gravity flows, suggesting conditions still farther offshore and in deeper water. A model depicting these environments of deposition is shown in Figure 13. No sediment is known to have been derived from the west, and there is no evidence bearing on the nature of Early Triassic environments of Union Wash.

STRATIGRAPHIC ANALYSIS OF MIDDLE (?) TRIASSIC MARINE-TO-CONTINENTAL ROCKS, SOUTHERN INYO MOUNTAINS, EAST-CENTRAL CALIFORNIA

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The transitional sequence of strata separating the Lower Triassic marine assemblage of rocks from overlying, in part, terrestrial volcanic-arc rocks are well exposed in the Inyo Mountains. Although the rocks are strongly deformed, the sequence is interpreted to be substantially intact. Laterally interfingering lithosomes defined within this sequence are correlated on the basis of three marker horizons. These horizons are: 1) The base of the lowest, thick-bedded conglomerate unit, 2) the first appearance of volcanic clasts within the section, and 3) the base of the first volcanic flow rock.

Transition sequence lithosomes can be broadly grouped into shallow marine to intertidal, supratidal and/or sabkha, and alluvial fan depositional environments. Fine-grained clastic rocks and carbonate rocks in the lower portion of the sequence, lithosomes Hm and Bls (Figures 14 and 15) are interpreted to be shallow marine deposits formed in a quiet water, tidally dominated coastal area that was subject to the passage of storm waves and to mild wave-current influence. Limonite bearing marble (Ym) with rip-up clasts (?), rarely preserved delicate wavy lamination, and possible calcitized anhydrite fabrics is present in southern exposures between the alluvial clastic deposits and the marine rocks. These strata are interpreted to be intratidal and/or sabkha deposits alluvial deposits succeeded and, in part, coexisted with, these marine and supratidal deposits. In general, these alluvial deposits coarsen upward and are therefore interpreted to become more proximal up section. Alluvial deposits of the Cerro Gordo area consist of a succession of lower and middle fan lithosomes interpreted to represent braided stream, stream flood, and sheetflood deposits (Rm, Rc); debris-flood, waning-stage flow, and braided stream deposits (Gc); abandoned channel fill deposits (Osh); and very shallow, braided channel, sheetflood, and low longitudinal bar deposits (Rss) (Figure 14). Reworked tuffaceous deposits or ignimbrite deposits (Wc) occur at various stratigraphic horizons within the fan sequence and are present as far north as the Burgess Mine area . Alluvial deposits (Rgss) are also recognized at exposures in the northern Swansea Wash and Union Wash areas. These alluvial deposits are interpreted to orig-

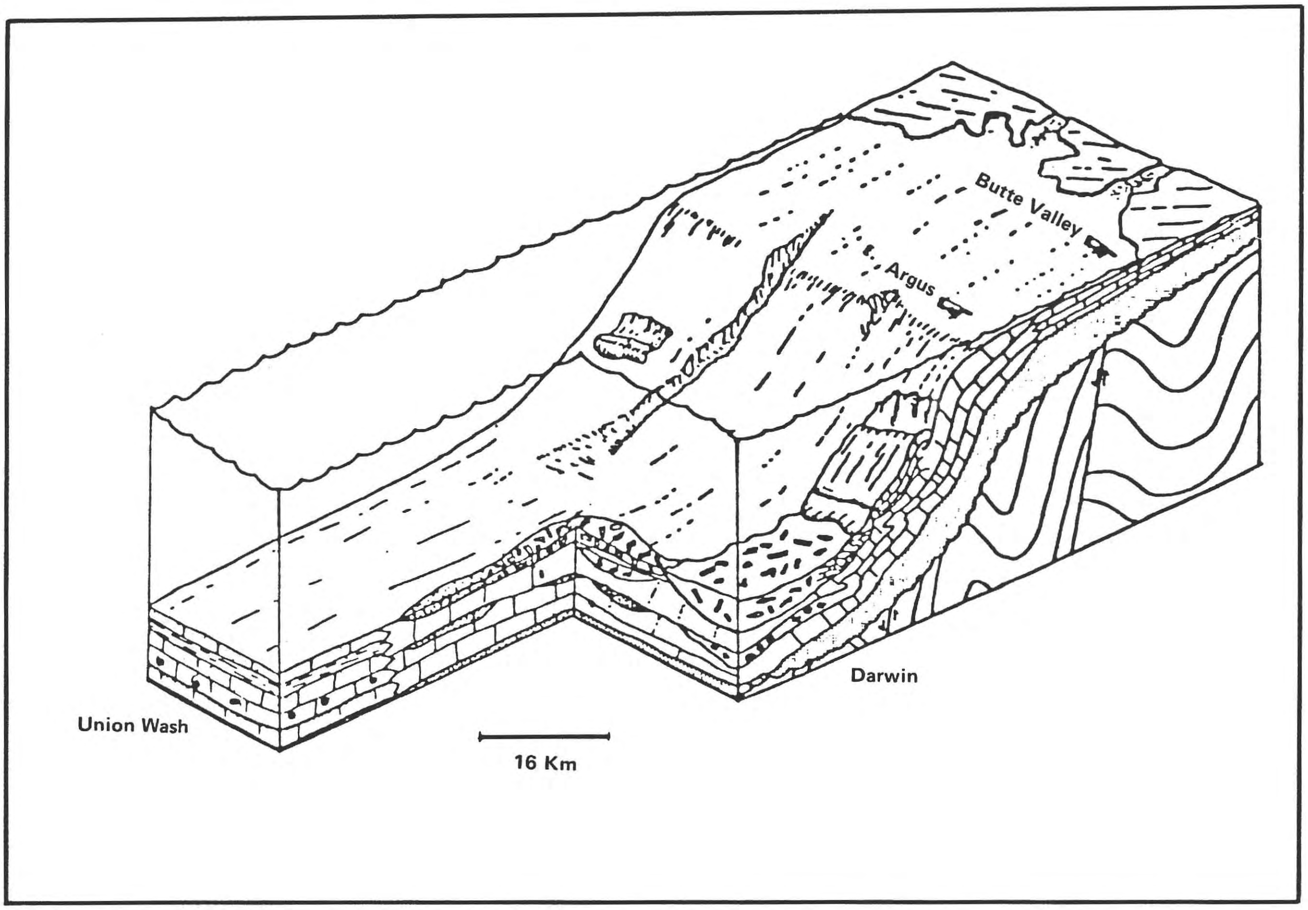


Figure 13. Depositional model for middle Lower Triassic rocks in east-central California (model adapted from Cook, 1977).

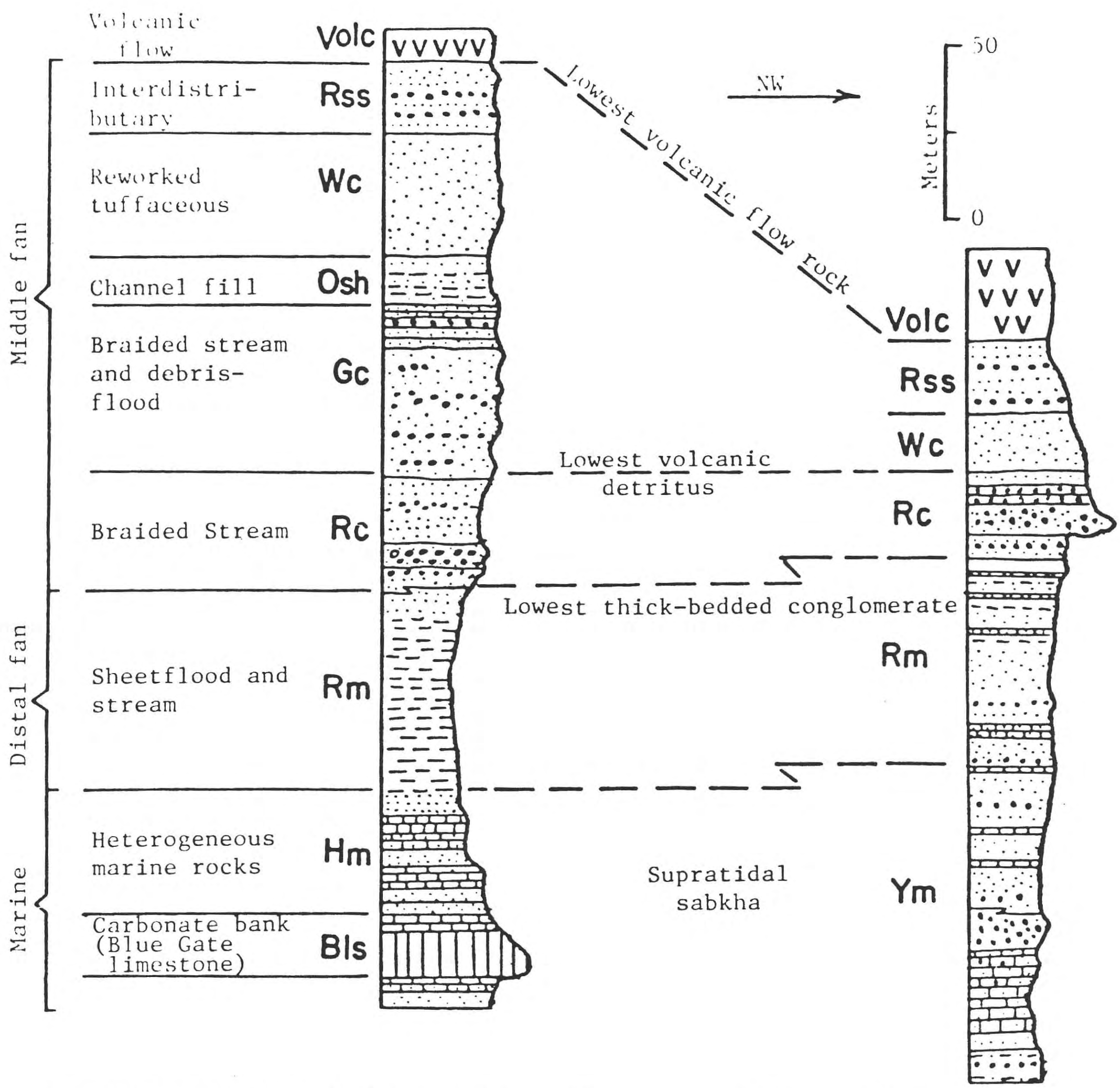
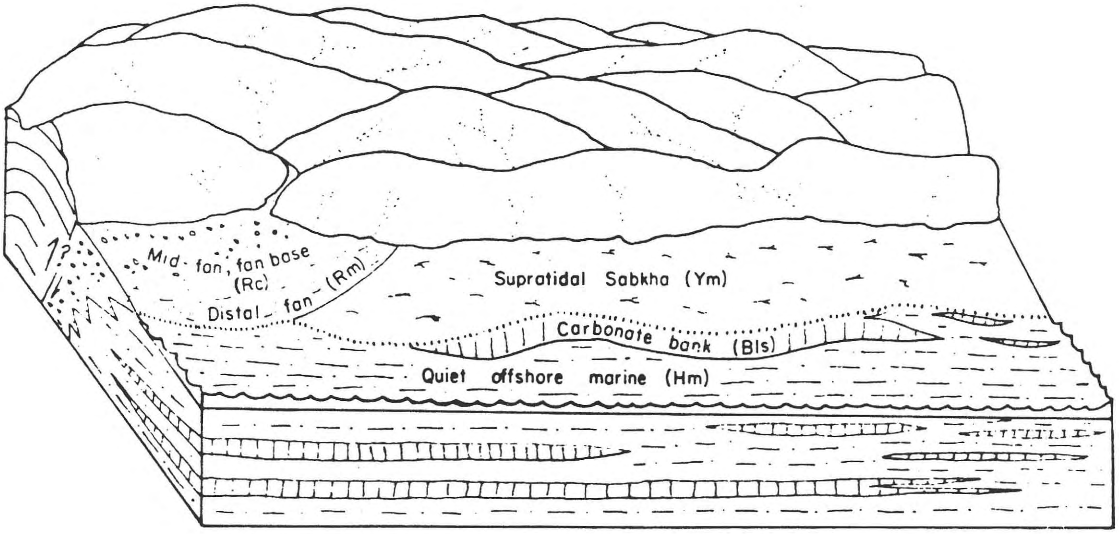
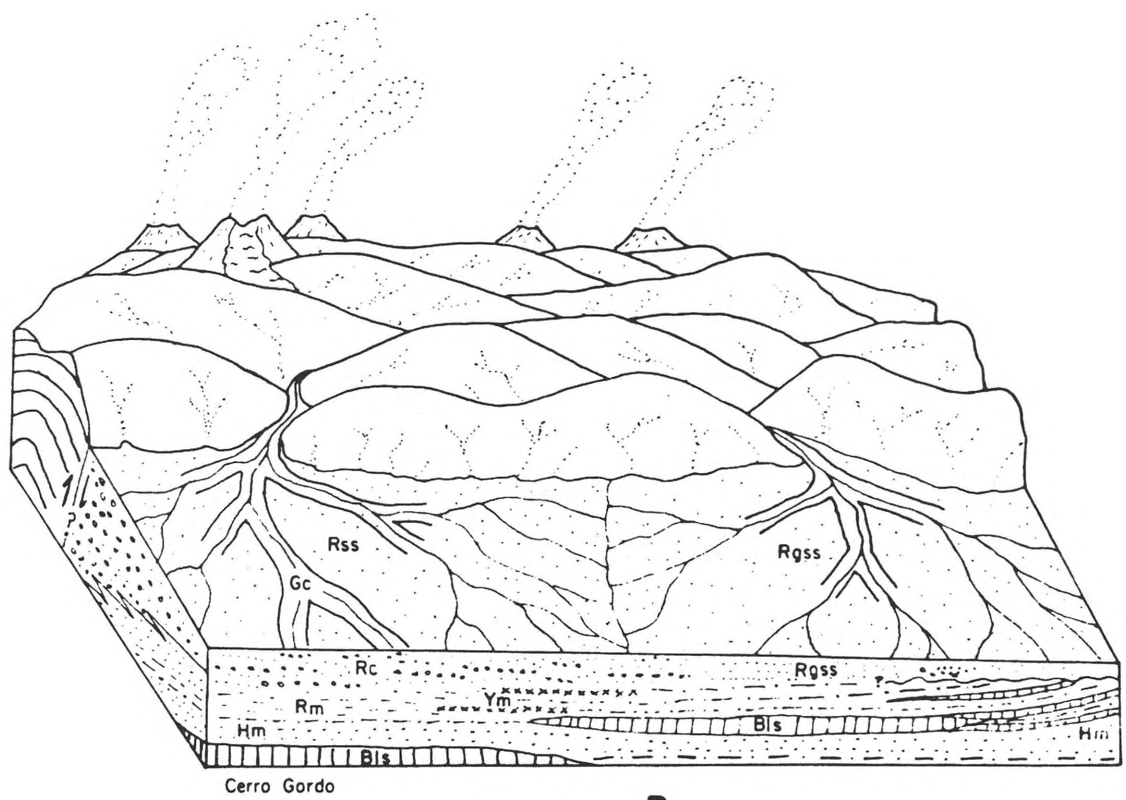


Figure 14. Stratigraphic columns of Triassic marine to terrestrial sequence, Cerro Gordo area, southern Inyo Mountains. Letter designations refer to informal lithosome divisions.

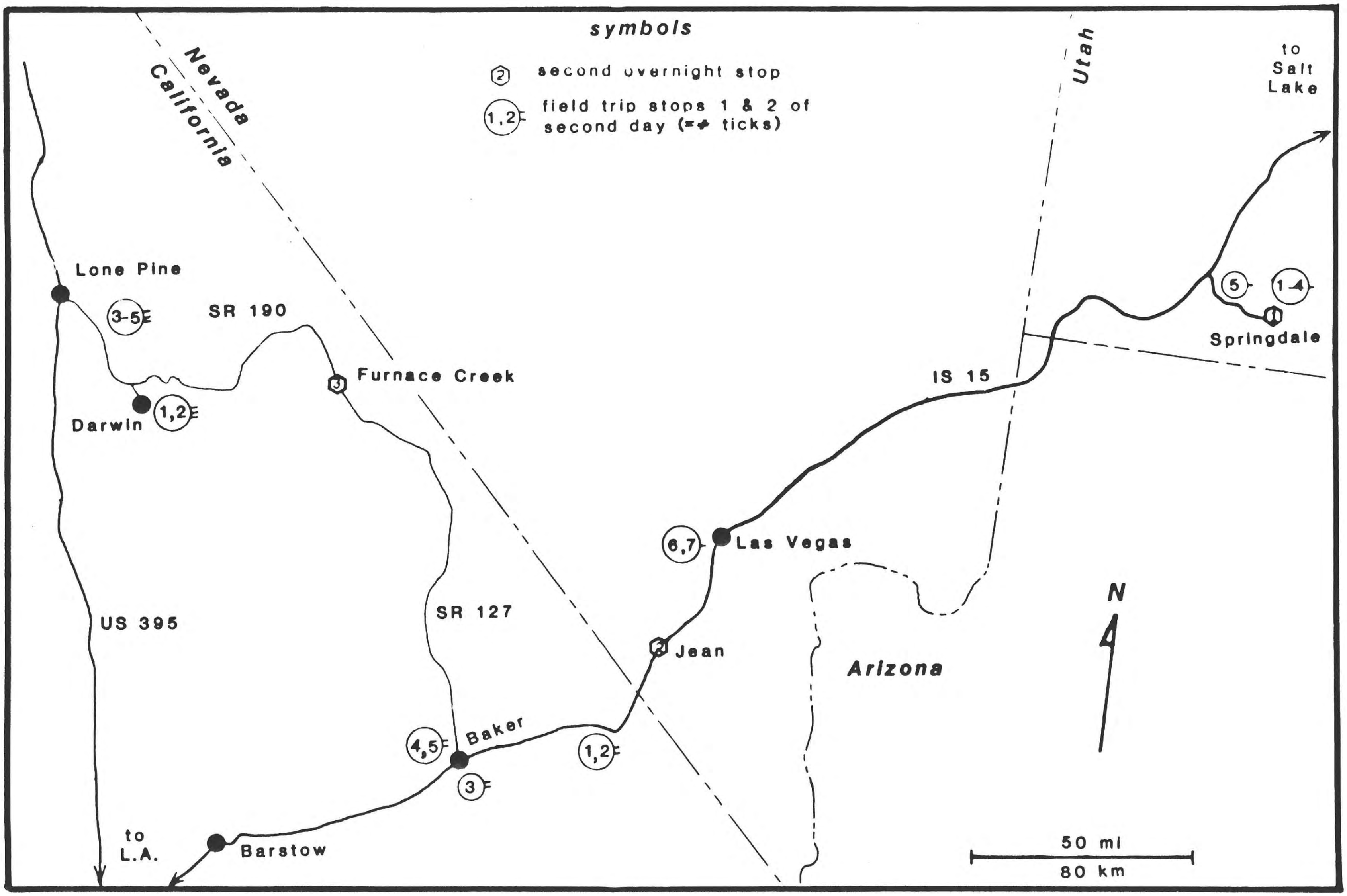


A



B

Figure 15. Paleogeographic reconstructions of the Early Triassic marine to terrestrial and volcanic transition, southern Inyo Mountains, California. Block A illustrates formation of a highland composed of sedimentary rocks and progradation of an alluvial fan across sabkha and nearshore marine environments. Block B illustrates a time following volcanic eruption. The alluvial fan shown in A has continued to prograde and a second fan (Rgss) has formed. Volcanic flow rock blankets the area shortly after this time.



Road Log Index Map

inate from a contemporaneous but less volcanic-rich provenance.

The paleogeographic framework of these lithosomes is illustrated in Figure 15. Correlation of the first appearance of volcanic clasts within the sequence, together with a coincident change in depositional character (as between Rc and Gc), leads to the interpretation that a period of uplift and erosion of sedimentary rocks preceded the onset of volcanism within the provenance. The upward shoaling of the marine strata and influx of clastic detritus and volcanic flow-rock observed in the Inyo Mountain transitional sequence documents initial uplift and igneous activity along this portion of the east margin of the Sierran, Andean-type arc in latest Early to early Middle (?) Triassic time.

FIELD TRIP ROAD LOG: FIRST DAY

INTRODUCTION

The purpose of this field trip is to examine spatial and temporal transitions in depositional environments, facies, and stratigraphic sequences within two fundamentally different early Mesozoic tectonic settings. Although the age limits are not well constrained, upper and lower limits of the stratigraphic sequences to be examined enclose approximately coeval lower Mesozoic rocks.

We will travel by van from Salt Lake City to Springdale, Utah, where we will have dinner and spend the night.

During the first day's stops, we will be examining the classical lower Mesozoic stratigraphy of the Colorado Plateau and similar strata in southern Nevada. In southwestern Utah, the lower Mesozoic section is predominantly cratonally derived, fluvial and eolian clastics. Only in the Lower Triassic section are marine strata represented. In southern Nevada, the marine part of the Lower Triassic section is greatly expanded. The Upper Triassic rocks are quite similar to those of the Colorado Plateau except for a slight hint of tectonic and volcanic events underway to the west.

In addition to research of contributers to this guidebook, the first day's stop description has drawn upon the work of Stewart and others (1972a, b) and Bessel (1970).

After breakfast the trip will begin at the Visitor's Center, Zion National Park.

Mileage Incremental	Mileage Cumulative	Description
0.0	0.0	**STOP 1**. Visitors center, Zion National Park. Gather on the patio in front of the Visitors Center. From this vantage point, we have a good overview of the upper part of the lower Mesozoic stratigraphic sequence in the eastern wall of Zion Canyon. Immediately to the northeast, the highest mesa, on top of which rests a layer of reddish-brown sandstone, is the East Temple of the Virgin River. The reddish-brown sandstone is the Temple Cap Sandstone. An unconformity at the top of the Temple Cap Sandstone separates it from the overlying marine Carmel Formation which contains invertebrate fossils of late middle Bajocian age (Peterson and Pipiringos, 1979). Approximately 60 km to the east, this unconformity truncates an unconformity separating the Temple Cap Sandstone from the Navajo Sandstone.

In the upper part of the Navajo Sandstone cliffs, crossbedding is truly large scale with individual sets having thicknesses greater than 40 m. In the middle and lower part of the sandstone, crossbed sets commonly are 3 to 7 m in thickness. In addition to crossbedding, the lower 100 m of Navajo Sandstone contains sedimentary features of groundwater origin. These include desiccation cracked horizontal erosion surfaces and limestone lenses. These can most readily be seen by taking the scenic drive along the floor of Zion Canyon. The erosion surfaces and limestone lenses crop out in the lower 100 m of the sandstone cliffs. The lower contact of the Navajo Sandstone is transitional with the underlying fluvial siltstone and sandstone of the Kayenta Formation.

The Kayenta Formation is composed of reddish-brown siltstone and light brown to reddish-orange cross-stratified sanddhstone. The prominent sandstone ledge within the Kayenta Formation is the Lamb Point Tongue of the Navajo Sandstone. As it thickens northward, intervening Kayenta strata pinch out.

The prominant sandstone cliff, at the base of the Kayenta Formation, is the uppermost of three members of the Moenave Formation, the Springdale Sandstone Member. The Springdale Sandstone consists of pale red to light brown fine-grained to very fine-grained sandstone containing tabular, planar and trough cross-stratification.

Successively beneath the Springdale Sandstone Member are the siltstones and claystones of the Whitmore Point and Dinosaur Canyon Members of the Moenave Formation. From our vantage point they are not readily distinguishable. The Whitmore Point Member tends toward greenish-grey colors; whereas, the Dinosaur Canyon Member is reddish-brown. The Whitmore Point Member is considered late Sinnemurian to early Pliensbachian (Early Jurassic) on the basis of palynomorphs (Peterson and Pipiringos, 1979).

1.4 1.4 Continue past Zion Canyon Scenic Drive through tunnel to turn out on right. Return to Utah Highway 9. Turn left.

5.9 7.3 **STOP 2**. At this stop, we will walk on bear rock exposures of multiple sets of eolian cross strata. Individual sets range in thickness from 3 to 7 m. Bounding surfaces of sets of cross-strata (Brookfield, 1977) are predominantly first and second order but third order surfaces (reactivation surfaces) can be found. Within individual sets of cross strata, grainfall and sandflow strata (Hunter, 1977) can be identified. The sand is fine to very fine and the coarser grains are well rounded. Typically, the grains are frosted but this feature is principally diagenetic rather than the result of eolian action.

Upward through the Navajo Sandstone the direction of dip of the cross-strata rotates clockwise from 152° at its base to 228° at the top (Figure 4). This indicates a change from northwesterly to northeasterly winds during deposition of Navajo sand.

3.5 10.8 Turn around and return through tunnel to turnout on lower switchbacks below tunnel. **STOP 3**. The Lamb Point Tongue of the Navajo Sandstone is clearly visible at this stop. It is eolian in origin as indicated by large sets of cross-strata and associated small scale structures. The Lamb Point Tongue is enclosed by a silt-dominated facies of the Kayenta. The siltstones are reddish-brown and either structureless, horizontally laminated, or cross-bedded. These were deposited on broad floodplains during periods of overbank flooding. Interbedded with the mudstones are channel-shaped sandstones up to 4 m thick and 30 m wide.

These represent in-channel deposition of both meandering and braided-stream systems.

5.5 16.3 Continue on Utah Highway 9 towards Springdale. Park entrance.

0.4 16.7 Continue to "Amphitheater" road on right.

0.6 17.3 Turn right and continue through gate in chainlink fence **STOP 4**. Although the base of the Chinle Formation is not exposed, this stop provides an excellent example of the variegated claystone and clayey siltstone and sandstone typical of the Petrified Forest Member of the Chinle Formation. The montmorillonitic clay resulted from alteration of pumice and volcanic ash which are believed to have been deposited in quiet water floodplain and deltaic environments.

The Petrified Forest Member of the Chinle Formation is unconformably overlain by the Dinosaur Canyon Member of the Moenave Formation. This member consists predominantly of horizontally laminated to thick-bedded reddish-brown siltstone. The upper part commonly is cross-stratified and ripple marked. These strata were deposited in stream channels and as overbank deposits. At many places in southwestern Utah and southern Nevada, this horizon is marked by conglomeratic sandstone containing granules to cobbles of quartzite, quartz, chert, siltstone, and limestone.

14.1 31.4 Return to Utah Highway 9. Turn right toward town of Virgin. Pass through Virgin to blacktop road to Hurricane Mesa.

3.5 34.9 Proceed to top of mesa. Turn around and return to base of Shinarump Member of Chinle Formation. **STOP 5.** The Shinarump Member of the Chinle Formation is a resistant, nearly flatlying sandstone and conglomeratic sandstone capping Hurricane and surrounding mesas. In the Belted Cliffs of Hurricane Mesa, the Shinarump and all but the base of the underlying Moenkopi are exceptionally well exposed.

From the lower part of the Shinarump Member, we will walk to the base of the Shinarump and down through as much of the Moenkopi Formation as time will permit.

The Shinarump Member of the Chinle Formation is typically yellowish-grey and pale yellowish-orange, fine- to coarse-grained friable sandstone. It is predominantly cross-stratified in tabular-planar and trough sets 0.2 to 0.6 m thick. From place to place, it contains lenses of granule and pebble conglomerate. Common clasts are quartz, quartzite, and chert. The top of the member is transitional upward to siltstones and claystones of the Petrified Forest Member. The base of the formation is unconformable on the Moenkopi Formation. Fossils indicate the entire Chinle Formation is continental and of Late Triassic age. It was deposited by a northwest-flowing braided fluvial system; whereas, the Petrified Forest Member represents deposits of meandering streams.

The Moenkopi formation is predominantly horizontally stratified siltstone and claystone. Gregory (1950, p. 115) subdivided it at this locality into six clearly recognizable members. In descending order, these are the: 1. upper red member 2. Shnabkaib Member 3. middle red member 4. Virgin Limestone Member 5. lower red member and 6. Timpoweap Member.

The Timpoweap Member is not well exposed at this locality. The upper, middle, and lower red members are predominantly reddish-grey, reddish-brown, and brown siltstone. The upper red member contains ripple laminated siltstone and cross-bedded sandstone. The middle and lower red members contain laminations, beds, and stringers of gypsum which in places are gradational into limestone or siltstone. These sediments represent coastal-plain and tidal-flat sedimentation. The Virgin Limestone and Timpoweap Members are predominantly gray limestone and siltstone and represent deposition in a very shallow, coastal marine environment. The Shnabkaib Member consists of grayish-red siltstone and white gypsum which is believed to be indicative of shallow marine sabkha conditions. In places, the red members contain abundant amphibian bone fragments and reptile tracks. The marine members contain a diverse marine fauna of shallow water affinities. The Timpoweap Member contains the *Meekoceras* fauna; the Virgin Limestone, the *Tirolites* fauna.

4.6 39.5 Return to Utah Highway 9. Turn right and continue to La Verkin intersection, Utah Highways 9 and 17.

11.8 51.3 Turn left on Utah Highway 9 west. Continue through town of Hurricane to I-15.

132.0 183.3 Head south on I-15 to Las Vegas.

5.0 188.3 In central Las Vegas, turn north toward Reno on U.S. Highway 95. Exit at Rainbow Blvd.

1.3 189.6 Turn left on Rainbow Blvd. to Charleston Blvd.

12.2 201.8 Turn right. Continue past Red Rock Recreation Area Scenic Drive.

0.8 202.6 Turn left on dirt road toward Blue Diamond Hill.

1.0 203.6 Excavations of Flintkote Corporation are visible on slopes to southeast. Continue about 1 mile. **STOP 6**. At Blue Diamond Hill, Permian and Lower Triassic strata strike north-south and dip gently west. At STOP 6, the east-facing cliff on the west is almost entirely composed of the Virgin Limestone Member of the Moenkopi Formation. The base of the Virgin Member rests on a thinned lower red member which unconformably overlies gypsum of the Harrisburg Member of the Permian Kaibab Formation. In the valley between Blue Diamond Hill and the Wilson Cliffs on the west, upper members of the Moenkopi are present but poorly exposed.

Here, the Virgin Limestone Member is approximately 365 m thick in comparison to only 35 m at Hurricane Mesa. The westward thickening is at the expense of the lower and middle red siltstone members so that the Shnabkaib Member rests directly on the Virgin Limestone Member.

As described by Bissell (1970), the Virgin Limestone

Member, is predominantly shallow-water, interbedded micritic limestone, fossiliferous limestone, pelecypod coquinite, algal limestone, and pelletal and oolitic limestone.

Blue Diamond Hill lies within Bissell's (1970) transition zone which is separated from the shelf by a northeast-southwest trending hinge line located in Las Vegas Valley. To the west and northwest, in the Spring Mountains Bissell assigned rocks of the Virgin Limestone Member to a depositional basin facies. Farther to the west in the northern Mojave Desert, lie the Moenkopi and Moenkopi like rocks of the overlap assemblage (Walker and Burchfiel, in this book).

2.4 206. Return to Red Bock Recreation Area Road (Charleston Blvd.). Turn left to Spring Mountain Ranch. Drive to parking lot. We will walk from the parking lot through the closed gate to the Wilson Cliffs behind the reservoir. This is a restricted area. Permission to enter must be obtained from Nevada State Parks and Recreation. **STOP 7.** At the base of the Wilson Cliffs, Upper Triassic and Lower Jurassic (?) rocks rest unconformably on the lower Triassic Moenkopi Formation. The imposing Wilson Cliffs are composed of Aztec Sandstone which here is approximately 755 m thick. The Triassic Jurassic sequence is autochthonous beneath the Keystone and related thrust faults which carried Cambrian Bonanza King and younger Paleozoic strata eastward. In addition to the bare cliffs of the Aztec Sandstone, the Shinarump Member of the Chinle Formation and the Moenave Kayenta equivalent are particularly well exposed.

The Shinarump Member is exposed in a series of channel-shaped lenses which crop out along the unconformity at the top of the Moenkopi Formation. The Shinarump consists of sandstone and conglomeratic sandstone. Stewart and others (1972b) recognized significant differences between these rocks and the Shinarump Member of more eastern localities. In southern Nevada, the Shinarump Member contains more abundant chert pebbles and, locally, sandstone contains conspicuous clay matrix. In addition, maximum pebble-size increases across southern Nevada from east to west (Stewart and others 1972b, p. 70). These data suggest the southern Nevada Shinarump may have had a western limestone and volcanic provenance.

The base of the Kayenta Moenave equivalent is marked by a dark brown to brownish-purple pebble conglomerate which is exposed at the base of the steep redbed slopes. Wilson and Stewart (1967) measured 5.5 m of conglomerate at this locality. Clasts of limestone, chert, volcanic, and clastic sedimentary rock fragments (Figure 5) suggests a western source in the Mojave Desert.

The overlying redbeds are more calcareous and contain a higher proportion of muddy siltstone and silty mudstone and less fine sandstone than the Kayenta or Moenave Formations farther east. At the top of the redbeds, a thick cross-bedded and wavey-bed, white sandstone is probably a tongue

		of the overlying Aztec Sandstone. The Aztec Sandstone is almost exclusively cross-bedded with only two or three zones of contorted stratification in the lower 85 m. As at Zion Canyon, crossbed dip direction changes from southeast at the bottom to southwest at the top.
5.8	211.8	Return to Red Rock Recreation Area Road. Turn right and continue to Pahrump Road.
10.3	222.1	Turn left and continue to I-15.
21.0	243.1	Turn right and continue to Jean exit. Turn left into Jean. We will have dinner and spend the night at Pop's Oasis.

FIELD TRIP ROAD LOG: SECOND DAY

The second day, we will be tracing the transition of the lower Mesozoic sedimentary rocks into the orogenic-volcanic terrane. This transition is best documented by the Aztec Sandstone as the pre-Aztec sediments were either never deposited or removed by erosion. In the western Soda Mountains we will examine Lower Triassic rocks of the overlap assemblage.

Mileage		Description
Incre-mental	Cumu-lative	
23.0	23.0	After breakfast, return to I-15, and turn left (west). Continue to Bailey Road Exit. Exit Bailey Road and turn left (south) over I-15.
3.1	26.1	Turn left at first road past on-ramp. Continue on dirt road. Kokoweaf road goes to left.
0.8	26.9	Continue straight on Piute Valley road to Delfonte Quarry road. Delfonte Quarry road is a track leading obliquely to right and then up a steep hill. Drive to top of hill and park. **STOP 1**. The first stop in the Mescal Range is near the Delfonte sandstone quarry. The resistant ridge of light-colored cross-bedded sandstone is overlain by the Delfonte Volcanics Burchfiel and Davis (1971). By comparison to the Aztec Sandstone in the Wilson Cliffs, here the Aztec is anomalously thin - only 137 m compared to 755 m. Otherwise the sandstone is quite similar in lithology and style of cross-bedding to that in the Wilson Cliffs. An additional feature at this locality is the presence of dinosaur tracks which can be seen on the surface of cross-strata (Reynolds, this guidebook). Approximately 13.5 m, below the top of the cross-bedded sandstone, a single bed of reddish-brown siltstone contains volcanic clasts. The consistent southeastward dip of cross-bedding (Figure 4) and the anomalously thin section suggests the upper part of the Aztec Sandstone was removed by errosion prior to being covered by volcanics. The underlying redbeds consist of well-rounded quartz siltstone in a clay matrix of probable devitrified ash. The siltstone is medium-bedded and cross-bedded in sets a few centimeters to a few tens of centimeters thick and contains numerous lenses and pockets of coarse sand-size to granules of volcanic rock-fragments. The base of the redbed section is covered. These rocks represent mixing of wind-transported, cratonally derived quartz with volcanic rock fragments and ash of the volcanic arc terrane.
0.9	27.8	Return to Piute Valley Road

and turn left. Return in direction of Kokoweaf Road. Turn left off road on to grassy flat and park. **STOP 2**. In the southeast corner of the Mescal Range, the Early Mesozoic section is faulted against Precambrian crystalline rocks. The Aztec Sandstone and underlying rocks strike northeast and dip to the northwest. Redbeds underlying Aztec Sandstone consist of reddish-brown quartz siltstone interbedded with sheered volcanics. Light-colored, cross-bedded sandstone within the redbed sequence may be slumped blocks of Aztec Sanstone but more probably are lenses of eolian sand in the predominantly fluvial redbeds. At the base of this sequence redbeds rest on volcanics which in the field appear to be debris flows or flow breccias. It is quite possible that this sequence is overturned but, if not, this locality represents the only place where pre-Aztec, cratonally derived fluvial rocks are in contact with volcanics of the arc terrane.

30.0	57.8	Return to I-15 and turn left (west) toward Baker to Central Baker exit.
1.0	58.8	Turn left (south) on Kelbaker Road. Continue to County Refuse Disposal Road.
1.0	59.8	Turn right. Road turns toward dump. Take road that bears to right.
1.5	61.3	Road forks back to left.
1.9	63.2	Continue straight to crossroad.
2.5	65.7	Continue straight. Cross edge of Soda Dry Lake to intersection at corral.
1.1	66.8	Turn left and continue to fork in road.
4.2	71.0	Bear to right and continue to Fox Hills Camp in Cowhole Mountains. **STOP 3**. At this stop in the Cowhole Mountains, the Aztec Sandstone overlies Paleozoic carbonates. Paleozoic rocks include the Pennsylvanian and Lower Permian Birdspring Formation in thrust contact with the Devonian Sultan Formation along the Cowhole Mountain thrust (Novitsky-Evans, 1978).

Basal redbeds of the Aztec Sandstone rest on a channel-shaped breccia containing blocks of Birdspring Limestone in a fine sand matrix. The overlying redbeds consist of quartz silt with dispersed volcanic clasts. Lenses of light-colored, cross-bedded sandstone are enclosed within the fluvial redbeds.

With exception of two volcanic units aggregating 81 m, the remaining 481 m of the Aztec section consists of light yellow-brown, quartz sandstone with very large scale cross-bedding. As at Zion Canyon and the Wilson Cliffs, vector resultants of cross-bed dip direction measured in vertical increments rotate clockwise from bottom to top of the section. Unlike the cross-bed orientation at Zion Canyon and the Wilson Cliffs, dip direction in the Cowhole Mountains changes from northwest at the bottom of the section to northeast at the top (Figure 4). This result suggests the Cowhole Mountains have been rotated tectonically by as much as 180°.

The Aztec Sandstone is overlain by a minimum of 575 m of predominantly porphyritic dacite, rhyodacite, and latite

(Figure 6a). In the lower part of the cross-bedded sandstone, 202 m above the basal breccia, a single volcanic unit, 69 m thick, consists of several altered and weathered dark greyish-purple flows and ash flows. The flow rocks are porphyritic and xenolithic and contain phenocrysts of oxidized pyroxenes (Figure 6b), biotite and plagioclase, and angular xenoliths of similar porphyritic volcanics a few centimeters in diameter. A much thinner, but less weathered porphyritic flow 43 m beneath the top of the Aztec Sandstone pinches out southward. The total thickness of the Aztec section is 542 m.

A short distance to the south, where the Aztec Sandstone appears to rest on older volcanics. The nature of the stratigraphic relationships at the base of the Aztec section are difficult to decipher because Paleozoic carbonates, pre-Aztec volcanics, and Aztec Sandstone are cut by at least two generations of dikes. Paleozoic carbonates are overlain by volcanics which in turn are overlain by the first of three horizons of limestone-block sedimentary breccia incorporated in the lower cross-bedded sandstone. The lower cross-bedded sandstone is overlain by a section virtually identical to the one where we are standing except for the absence of the upper, thin volcanic flow. The total Aztec section and underlying volcanics are 748 m thick. The cumulative thickness of volcanics and cross-bedded sandstone from the Paleozoic carbonates to the base of the reddish-brown siltstone is 226 m which presumably was the relief on the carbonate surface.

Similar relationships can be observed in a relatively small area of outcrop of Aztec Sandstone on the west side of Old Dad Mountain (Dunne, 1977). The Aztec contains a basal sedimentary breccia composed of Birdspring Limestone blocks in a sandstone matrix. Where the contact has not been intruded by fine-grained mafic hypabyssal rocks, the breccia rests on Paleozoic limestones. Yellowish-brown quartz sandstone containing large scale cross-bedding comprises 181 m of the section and is overlain by a porphyritic volcanic unit which is cut off by the Playground Thrust.

7.0 78.0 Return to central Baker for lunch. After lunch drive north from 4-way stop in central Baker, toward Death Valley. Continue to dirt road across Silver Dry Lake (talc piles by roadside).

2.2 80.2 Turn left and continue to intersection with powerline road.

5.1 85.3 Turn left. Continue past the end of Specter Spur. Turn off road to left and drive past power tower and park. We will walk up the fan to the Meonkopi outcrops south of the powerline road. **STOP 4**. At this stop we will examine early Mesozoic rocks in the Soda Mountains, which were first subdivided by Grose (1959). These consist of a lower unnamed unit of argillite, shale, and cherty limestone which is disconformably overlain by andesitic flows, breccias, and tuffs of the Soda Mountain Formation. Limestones in the lower unit have yielded Early Triassic gastropods. Grose

(1959) correlated this unit with the Moenkopi Formation based on age and lithologic similarities. The Soda Mountain Formation has not been dated directly but contains interbedded clean quartzite that Grose (1959) correlated with the Aztec Sandstone. It is cut by numerous faults which give the impression of a chaotic stratigraphy. Although not yet studied in detail, individual fault-bounded blocks display an interbedded sandstone and volcanic stratigraphy which, other than containing a larger number of flow units, is not unlike the sequence in the Cowhole Mountains.

Looking southeast our view is dominated by the Lower Permian Bird Spring Formation. The hills and ridges north of Spectre Spur are underlain by Moenkopi equivalent rocks and Soda Mountain Formation. Burchfiel and others (1980) mapped a sequence of conglomerate, calcsilicate metasediments, and limestone on the crest of Spectre Spur. These rocks probably form the base of Moenkopi equivalent rocks.

6.2 96.5 Return to Death Valley highway. Turn right towards Baker. Turn right at powerline road.

Continue until road approaches and turns along Highway Spur to the south. Park and walk up Canyon to top of Aztec Sandstone. **STOP 5**. At this stop on Highway Spur, the interbedded sandstone and volcanics are cut by numerous high-angle faults which are difficult to trace through the volcanics. At the base of the sequence, a volcaniclastic sedimentary breccia overlies approximately 2,100 m of andesitic to dacitic flows, flow breccias and hypabyssal rocks. Near the bottom of this sequence, slivers of metamorphosed limestone and calcsilicates are enclosed in the hypabyssal rocks. At the base of the sequence, volcanic and hypabyssal rocks are in fault and instrusive contact with steeply-dipping Paleozoic carbonates and calcsilicates. The volcanic and hypabyssal rocks in this part of the section probably have been repeated by faulting.

The Aztec Sandstone consists of light yellowish-brown cross-bedded quartz sandstone and flat-bedded grayish-purple quartz sandstone containing porphyritic volcanic clasts. Surfaces of beds in upper redbed units contain lags of feldspar phenocrysts. Within the cross-bedded sequence, cross-bedded sets of light-colored quartz arenite alternate with sets of purplish-gray volcaniclastic sandstone containing coarse sand-size volcanic clasts to volcanic cobbles at the base of the cross-bed sets. The alternation is believed to have resulted from changing wind direction. The volcaniclastic cross-bedded sandstones were formed from locally-derived, wind-transported volcanics.

From the basal breccia to the highest sandstone, the total thickness is approximately 357 m. This sequence is overlain conformably by porphyritic quartz latite with quartz and feldspar phenocrysts.

114.0 205.5 Return to Baker for dinner. After dinner, drive north on State Route 127 (toward Death Valley) to Furnace Creek Ranch where we will stay the night.

FIELD TRIP ROAD LOG: THIRD DAY

On the third day, we will examine western outcrops of a belt of Mesozoic strata present from Butte Valley in the southern Panamint Range to Union Wash in the southern Inyo Mountains (Figure 11). These strata rest with varying degrees and types of unconformity upon previously deformed Pennsylvanian and Permian rocks. Three upward gradational sequences are recognized within lower Mesozoic strata. These are: 1) a Lower Triassic marine sequence up to several hundred meters thick 2) a paraconformably (?) overlying transitional sequence about 200 m thick that was deposited in shoaling marine, then terrestrial environments and 3) conformably (?) overlying terrestrial to shallow marine (?) volcanic and volcaniclastic rock deposited sometime during the interval latest Early Triassic-Middle Jurassic.

Mileage Incremental	Mileage Cumulative	Description
72.0	72.0	After breakfast, depart Furnace Creek Ranch and continue driving west on State Route 190 (toward Lone Pine) to Darwin turnoff.
6.1	78.1	Proceed southeast to stop sign in heart of Darwin.
1.8	79.9	Turn left and follow narrow paved road over Darwin Hills and down toward Darwin Canyon. Follow dirt side road that branches left (Actually trending straight ahead) where paved road swings right.
2.3	82.2	Proceed on dirt road, bearing right at first (and only) fork with another dirt road. Turn vehicles around and park. **STOP 1.** Lower Triassic rocks crop out on both sides of the road. The Permo-Triassic contact is about 0.3 km to the west near the base of the Darwin Hills. The Lower Triassic rocks exposed on the east side of the Darwin Hills lie unconformably upon the deep-water Lower Permian Owens Valley Formation, here characterized by carbonate debris flows. Along the east side of the Darwin Hills, beds on either side of the contact between these two formations have similar strikes, but dip of the Permian beds is considerably greater than that of the Triassic section, indicating that pre-Early Triassic folds were refolded coaxially by post-Early Triassic deformation. An extensional vs. compressional origin for these folds will be discussed at this stop. The basal beds of the Triassic, commonly consisting of mottled, silty limestone, probably are of shallow, nearshore origin. The overlying siliceous shales with large limestone nodules contain ammonoids and radiolarians. A relatively deep water environment far from shore is suggested by the abundance of the pelagic organisms and the lack of benthic organisms and land derived sediment.
3.1	85.3	Return to paved road; turn left. Proceed down Darwin Canyon and park just short of a major right-hand bend in canyon. **STOP 2.** Exposed part way up the cliff straight ahead (north) of the vehicles is the most photogenic exposure of the angular unconformity between the Lower Permian Owens Valley Formation, here characterized by thin-bedded turbidites, and shallow water, Lower Triassic silty limestones. Very tiny gastropods are abundant in some of these Triassic beds. Shallow channels cut into the Owens Valley Formation and filled with clasts derived from that unit are exposed on the cliff.

This exposure of the unconformity can be traced only the length of the outcrop here because it is cut off by the left lateral Darwin shear zone a few meters behind the outcrop.

17.0 102.3 Turn around and return to State Route 190. Drive west on Route 190 to Cerro Gordo Road on the outskirts of town of Keeler.

4.0 106.3 Turn right onto Cerro Gordo Road and proceed uphill to a point where the road makes a broad 180° turn to the left, at the end of which it leaves the main canyon and traverses a narrow "rocked" cut across a low cliff. Park vehicles in uphill side canyon (access road to Estelle Tunnel tailing pile) just before entering narrow "rocked" roadcut. **STOP 3**.

Bold blue-gray limestone exposures just up the side canyon are the informally named Blue Gate limestone. Beds west (downhill) of the Blue Gate limestone represent an upward-shoaling environment transitional between the Lower Triassic marine strata (including Blue Gate) and overlying terrestrial volcanic rocks exposed west (downill) of the vehicles. Nonmarine clastic rocks of the transitional sequence at this locality are 220 m thick and are comprised in part of outer to middle fan deposits (Figure 14). At this stop, we will briefly examine Rc and Gc conglomeratic lithosomes. Rc deposits consist of parallel-bedded sandstone and sandy conglomerate. Bed thickness and clast size increase upward through the unit. The Rc lithosome is interpreted as longitudinal bar deposits that formed within a braided stream environment. Prominant lithologies of Rc lithosome are chert-arenite and calclithite.

The Gc lithosome, interpreted as predominantly debris flood and wanning stage flow deposits, is particularly well exposed at this locality. Sedimentary structures observed within this unit include channels, fining-upward beds of conglomerate to sandstone, and parallel bedding. Gc lithologies include calcitic submature calclithic and chert-bearing volcanic arenite and volcanic-bearing calclithites.

The contact between Rc and Gc units is sheared and locally intruded by a dike, and its original depositional nature is not clear. If the appearance of volcanic clasts above this contact indicates the beginning of volcanic eruption during time of deposition, then the consequent changes in provenance relief may have resulted in an intraformational unconformity represented by the Rc-Gc contact.

0.9 107.2 Return to vehicles and proceed up Cerro Gordo Road to point where Blue Gate limestone beds trend across the road. Park on shoulder of road. **STOP 4**.

We will examine the Blue Gate limestone at this stop. The several elongate bands of resistant Blue Gate limestone exposed here were repeated by faulting and tight folding. Several of the thick limestone beds are composed of tightly packed limestone clasts in a similar limestone matrix. Other beds contain cross sections of pelecypods and well-developed ooliths. The latest data concerning environment of deposition will be presented at this stop.

From vehicles walk northwest along bedding strike approximately 0.8 km.

Here we will make a westward traverse through nearly the entire transitional sequence from the top of the Blue Gate limestone to the first exposures of volcanic flow rock. This section of the transitional sequence contains deposits of sabkha through medial to inner alluvial fan facies (Figure 14). These exposures are significantly deformed by a pervasive cleavage, and the continuity of the stratigraphic sequence is uncertain. The traverse begins in the Ym unit (sabkha) which consists of tabular bedded yellow-orange, limonite-bearing very fine silty marble, and minor green matrix-supported conglomerate. Grayish-red and grayish-green sandstone, siltstone, claystone, and less abundant reddish conglomerate of Rm unit overlie and are in part interbedded with Ym marble beds. The abundant planar bedding and horizontal lamination within Rm deposits are interpreted to have formed at the outer fringe of an alluvial fan as sheetflood and braided stream deposits. Rm deposits become more sandy and conglomeratic up section, and in general there is a coarsening and thickening-upward transition from Rm through Rc lithosomes.

The Gc lithosome is not present between the Rc and overlying Wc units at this locality due to depositional discontinuity or structural truncation. Wc deposits are calcite and quartz rich and contain rare parallel bedding and concentrations of clasts resembling channel deposits. A tenable interpretation is that Wc deposits represent very altered reworked tuffaceous deposits or ignimbrite flow deposits. Volcanic clasts occur within Wc and overlying Rss lithosomes. Rss deposits consist of quartz arenite and submature, volcanic-bearing calclithic chert-arenite, and are interpreted to have formed in braided distributary channels on an alluvial fan. The traverse terminates at the contact with the first occurrence of volcanic rock that represents the spreading of flows across tha area.

Return to vehicles and drive to Las Vegas.

END OF FIELD TRIP.

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NOTES

Utah Geological and Mineral Survey Special Studies 60, 1983
Guidebook Part 2 — GSA Rocky Mountain and Cordilleran Sections Meeting
Salt Lake City, Utah — May 2 - 4, 1983

THE GEOLOGY IN AND NEAR CANYONLANDS AND ARCHES NATIONAL PARKS, UTAH

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INTRODUCTION

The region in and around Canyonlands and Arches National Parks is some of the most colorful on the Colorado Pleateau. Red rocks abound and the area is appropriately known as "red rock country." It is a stark, rugged land, carved unmercifully into myriad canyons by the Green and Colorado Rivers and numerous tributaries. From the canyons rise a multitude of crags, spires, buttes, mesas, all carved from red rocks.

Almost all geologic features seen at the surface resulted from the whimsical salt at depth which, in turn, was directed by basement structure. The younger Laramide Orogeny was an afterthought that had little effect in this particular part of the Colorado Plateau province. Diapiric salt formed the elongate, northwest-trending valleys, such as at Moab, and it was flowing salt that bulged up the smaller anticlines in the region. It was salt flowage or subsurface dissolution that resulted in the intricate fracture patterns common to the area. The culprit is the Paradox Formation, for this is the heart of the Paradox Basin, a mid-Pennsylvanian "Dead Sea."

TECTONIC HISTORY

The tectonic history dates back to about 1,700 m.y., when two continental-scale fracture zones broke up the basement. One of the lineaments trends northeast, extending from Lake Superior, through the Colorado Mineral Belt, and down the Colorado River to Grand Canyon. Warner (1978) named it the Colorado Lineament. The other, a northwesterly lineament, extends from northeastern Washington, through the Paradox Basin, and at least as far as the Wichita Mountains of Oklahoma. The latter is the more dominant of the pair and is known as the Olympic-Wichita Lineament. The lineaments intersect in the Moab region in swarms of basement wrench faults. The northeast set was offset in a left lateral sense, while the northwest faults show right lateral displacement (Figure 1). Thus, Sigma I of the stress field was oriented north-south, or 90° from the Laramide thrusting. The Precambrian stress field also produced north-south oriented normal faults that underlie the large monoclines across the Plateau, and east-trending anticlines of the ancestral Uinta Mountains (Baars and Stevenson, 1982). Once in place, the basement structural fabric dominated all subsequent tectonism on the Colorado Plateau and Southern Rockies.

The early Paleozoic was a time of relative structural quiescence, although minor rejuvenations of the basement faults occurred in Cambrian, Late Devonian, and Mississippian times. The vertical component of movement was sufficient to cause the various stratigraphic units to thin over the high structures, or in some cases shoaling of the seafloor caused variations in the depositional environments. The oil and gas fields producing from these strata produce from local paleotectonically controlled sand bars in the McCracken Sandstone Member of the Elbert Formation (Late Devonian) and crinoidal bioherms in the Mississippian Leadville Formation (Figure 2).

During Pennsylvanian time, the great east to west extension of the crust caused the Ancestral Rockies to bulge and the related basins, including the Paradox, to sag along the pre-existing basement faults. Almost immediately, the salt began to fill the nearly isolated Paradox Basin and arkosic clastics were shed from the adjacent Uncompahgre Uplift. Kluth and Coney (1981) blamed it all on an unfortunate collision of the South American — African plate with North America. By mid-Pennsylvanian time, the salt began flowing away from the arkosic

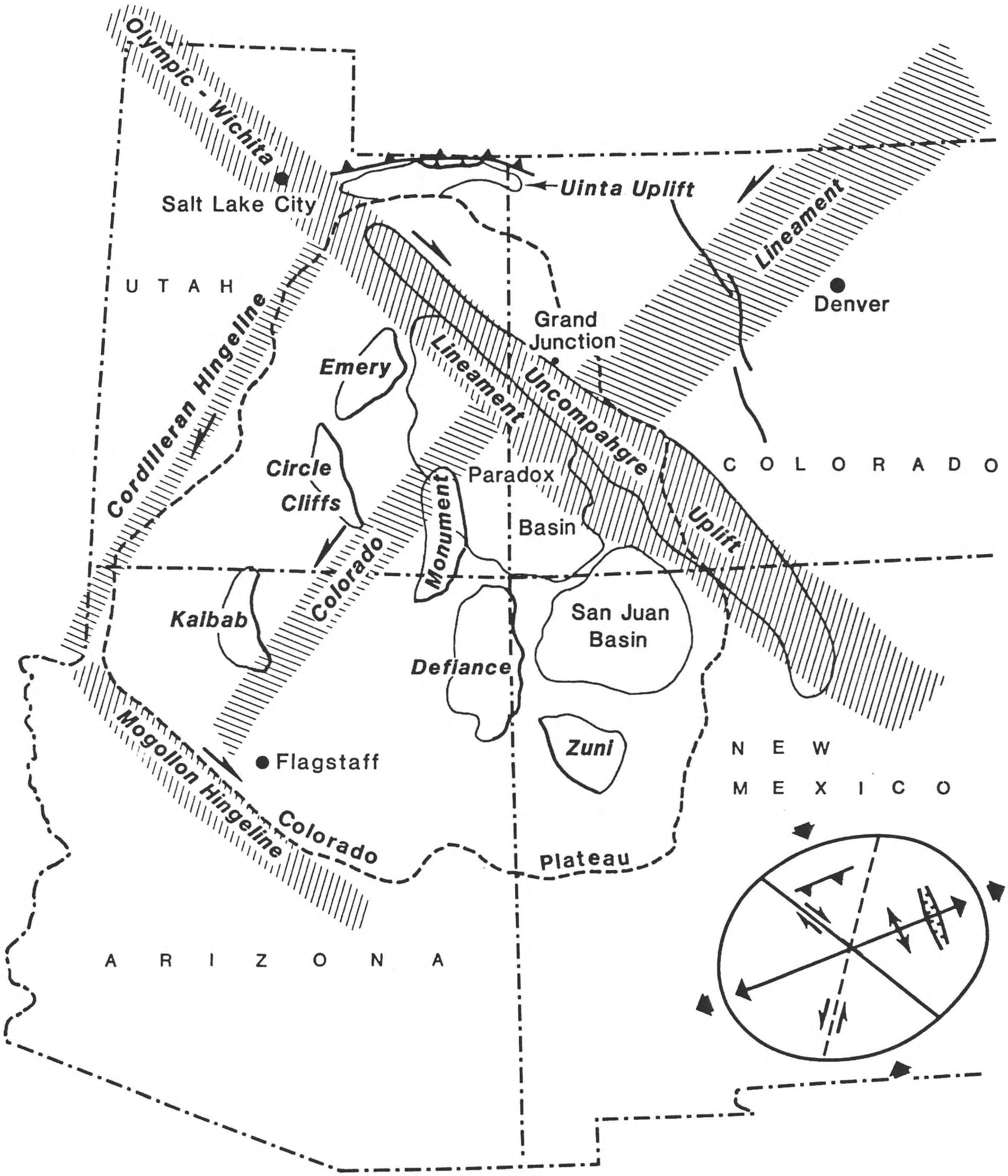

Figure 1. Map showing location of Colorado Plateau and relationship to major orthogonal set of basement lineaments. Northwesterly lineaments are right lateral, northeasterly lineaments are left lateral. Stress-strain ellipsoid oriented such that maximum compressive stress is directed from the north.

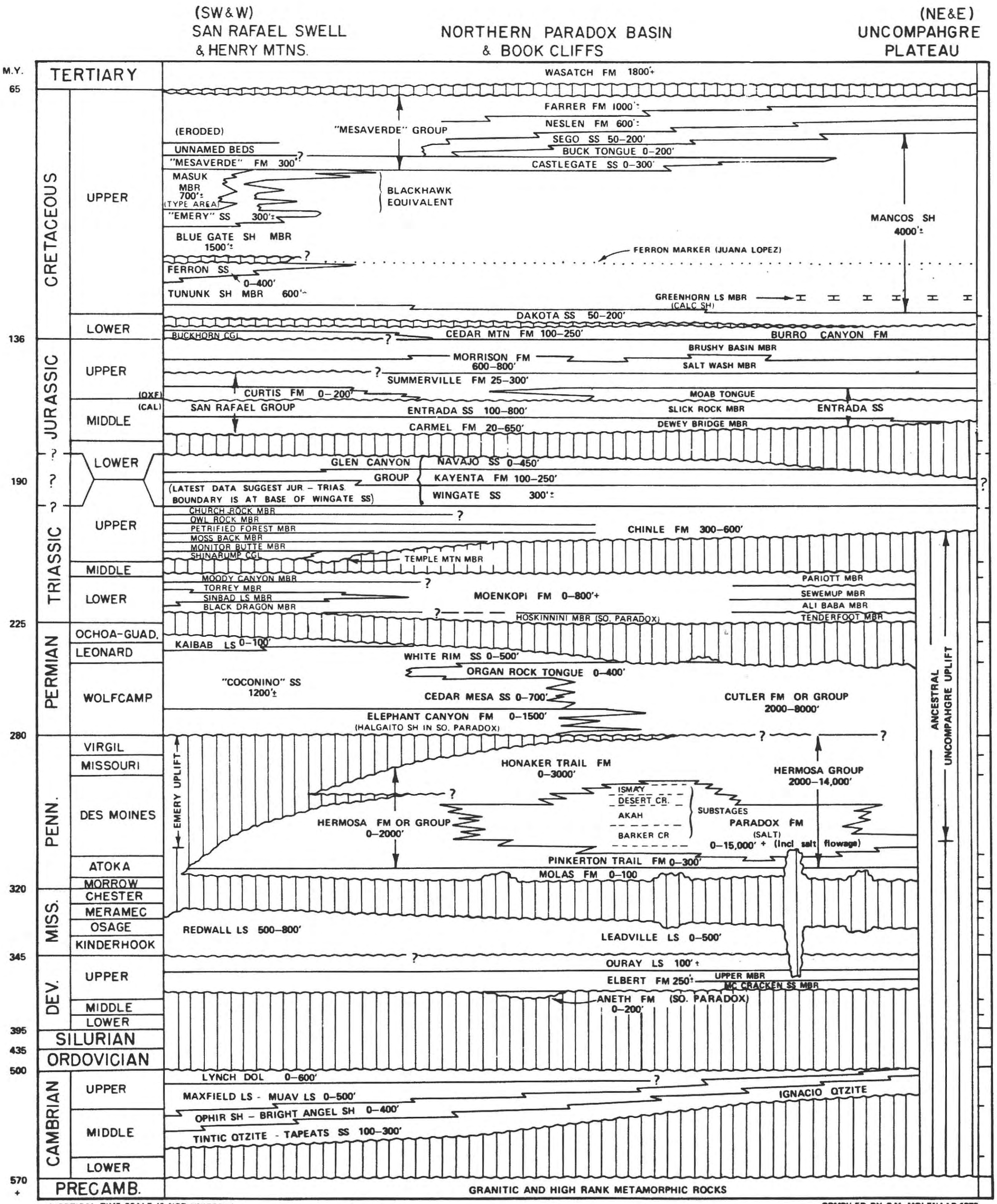

Figure 2. Correlation chart of northern Paradox Basin and Book Cliffs.

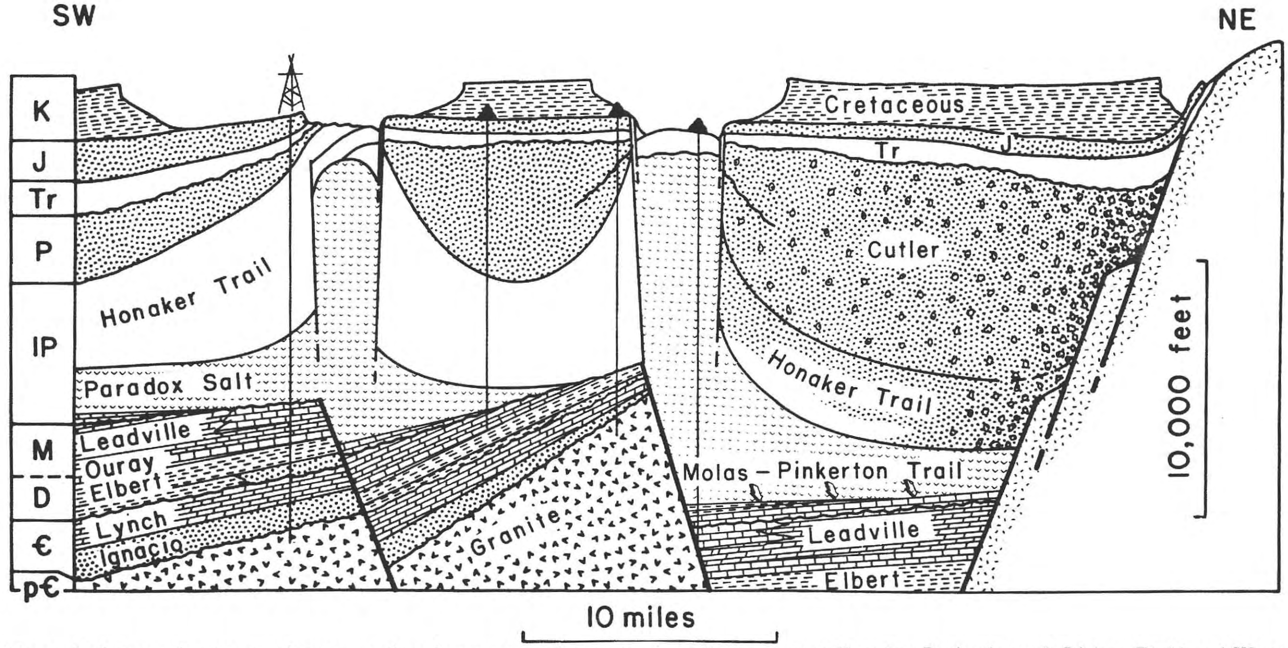

Figure 3. Schematic structural cross-section drawn normal to tectonic strike in eastern Paradox Basin through Lisbon Field and Wray Mesa region, showing relation between pre-salt basement faults and post-Desmoinesian salt-flowage diapiric structures (from Baars, 1966).

overload on the east, and was deflected upward along the now huge basement fault blocks (Figure 3). Nearly 5,000 m of salt were drilled in the Paradox Valley, adjacent to a deep-seated fault with nearly 2,000 m vertical displacement. Salt flowage continued through Jurassic time, when the supply of Paradox salt was finally depleted. The numerous salt anticlines were finally buried by Cretaceous sediments.

Even the smaller structures along the margins of the salt basin were in place during mid-Pennsylvanian salt time. The salt thins over and around the small structures of the northern Monument Upwarp, and the zero isolith is controlled in part by the present day anticlines (Figure 4). The overall Monument Uplift was still somewhat positive in Lower Permian time, as major facies changes were controlled by the margins of the huge feature. Virtually every structure at the surface today in the vicinity of the Paradox Basin was present and repeatedly rejuvenated throughout Paleozoic time.

There was little tectonic activity in the Paradox Basin during Triassic and Jurassic times. Salt was still flowing in the salt anticlines, but otherwise the red beds were burying the Paleozoic structures. Even the Uncompahgre Uplift was largely buried by Triassic sediments.

By the Late Cretaceous, the rolling waves of Laramide tectonism arrived in the Colorado Plateau from the west. Major structural features, such as the Monument Upwarp, the San Rafael Swell, and the Uinta Mountain arch, were again reactivated and greatly enhanced in structural magnitude. The Uncompahgre Uplift was re-elevated, but this time the added relief could be measured in a few hundred meters, a mere drop in the bucket to the relief it attained in the Pennsylvanian. Otherwise, there is little evidence of any Laramide effects on the salt anticlines. The massive salt bodies seem to have acted like the proverbial bowl of jello, and simply "shook off" the orogeny.

By Early Tertiary time, the Colorado Plateau was uplifted bodily toward the north. It all started in central Arizona, but involved the entire province. The initial stripping of the Cretaceous and older stratigraphic section began, dumping sedimentary debris into Lake Uinta that formed in a sag between the rigid high blocks of the Uncompahgre Uplift on the south and the Uinta Mountain arch on the north. In Middle Tertiary time the entire Rocky Mountain region was uplifted to approximately its present elevation. The "laccolithic" mountain ranges, such as the La Sal, Abajo, and Henry Mountains, were implaced during Oligocene time.

The runoff was gradually diverted toward the southwest, probably through a series of complex stream piracies, and the denudation of the Plateau country was rampant. Millions of cubic miles of

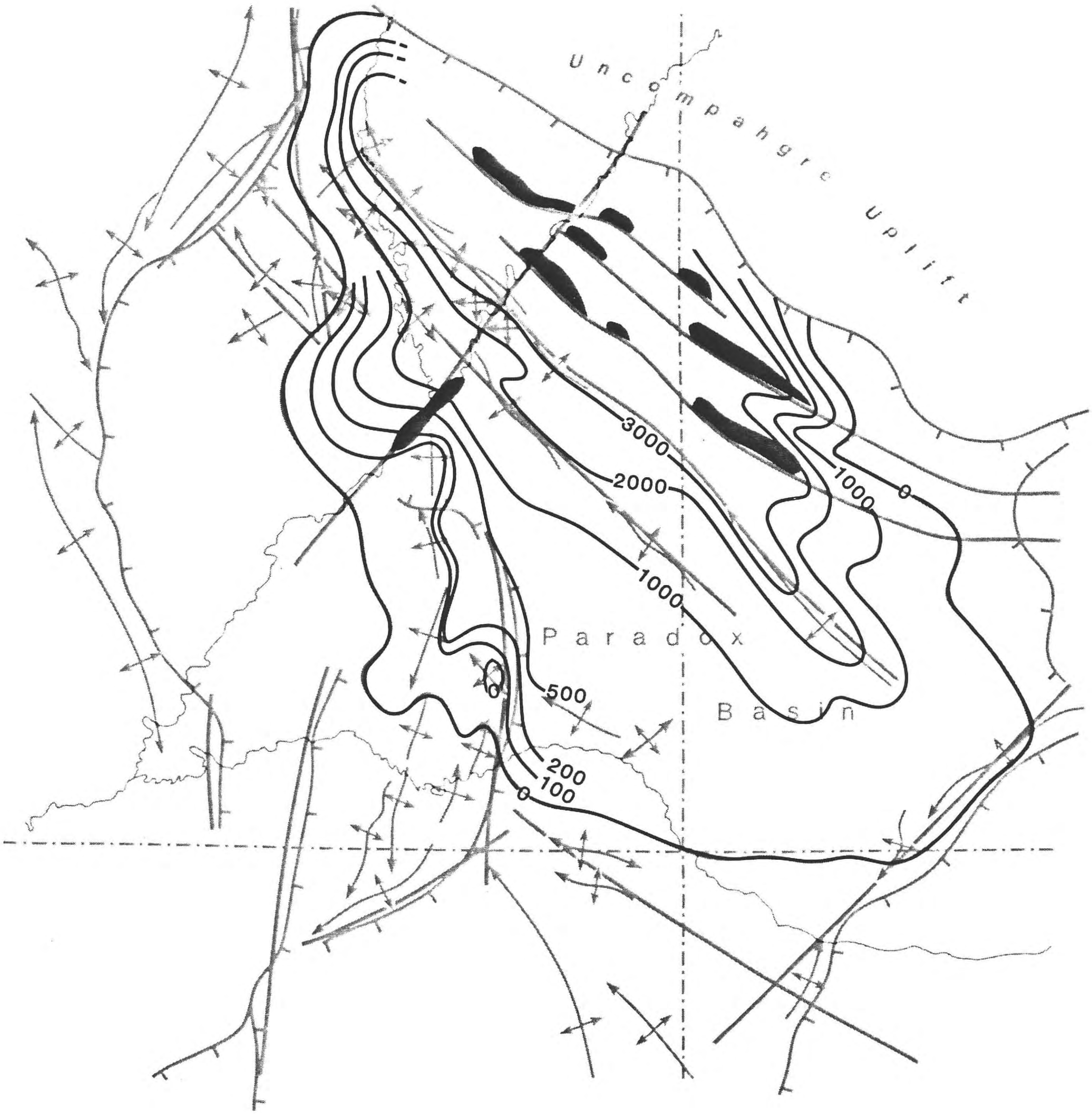

Figure 4. Index map of the Paradox Basin showing relationship of salt anticlines (solid black) to basement faults (thick gray lines). Contours are salt isoliths of the Paradox Formation; note relationship of net salt contours to structures along the shelf. Contour values in feet of net salt.

sediment were washed away by the Colorado River system, and redeposited in the Gulf of California where they belong. Now, instead, these river sediments are being deposited in Lake Powell and Lake Meade for future corn fields. The grandeur of Canyonlands and Arches National Parks is a result of the stripping away of vast amounts of the sedimentary rock cover that have been flexed by the Laramide, pierced and bulged by the salt, pulled apart by the Ancestral Rockies orogeny, all based on the basement tectonic fabric inherited from the Precambrian.

STRATIGRAPHY

Rocks of Middle Pennsylvanian through Jurassic age are well exposed in and near Canyonlands and Arches National Parks. Older rocks are present in the subsurface (Figure 2) and younger rocks are present to the north in the Book and Roan Cliffs. This discussion will concentrate on formations to be studied on this field trip.

The oldest formation exposed in the Moab region is the Paradox Formation. It is only seen in bedded form near the mouth of Gypsum Creek in the depths of Cataract Canyon, and no salt is exposed anywhere in the region. As the name implies, the Gypsum Creek exposures are mainly of gypsum, with some dark dolomite and black shale interbedded. Collapse breccias suggest, however, that salt was originally deposited there. Exposed are the Ismay, Desert Creek, and Akah stages. Other exposures of Paradox gypsum are inflowed and leached cap rocks of the various salt anticlines. The best place to see the torture these rocks have undergone is in the Onion Creek - Fisher Valley diapir east of Moab. Wells drilled into the salt structures encounter salt at about 800 to 1,200 ft depths. The only complete penetration of the diapirs was in Paradox Valley where Conoco drilled almost 15,000 ft of salt, topping the Mississippian Leadville Formation at 14,725 ft As previously discussed, the Paradox evaporite basin formed when there was extreme pull apart along basement rifts in Middle Pennsylvanian time. The basin was restricted by the Uncompahgre Uplift to the east, and tectonic shoaling along the other margins. The only truly documented entryway for sea water was from the San Juan Basin to the south. Other accessways to the open sea have been postulated, but not drilled to date. The evaporites are of a deep water type, being associated with black dolomitic shales and rare depauperate marine faunas, and are cyclic. There are 29 salt cycles in all (Hite, 1960), each separated by "black shales" and anhydrites. Potash salts occur in some cycles and are being solution mined at Potash, west of Moab. The evaporites interfinger with arkose along the source Uncompahgre Uplift on the east, and elsewhere grade into shallow shelf carbonate rocks at the basin margins. These shelf carbonates contain the algal bioherms that host prolific amounts of petroleum in the Four Corners area (Aneth Field, and numerous others).

The Honaker Trail Formation of the Hermosa Group constitutes the younger Pennsylvanian in the region. It consists of interbedded normal marine cyclic fossiliferous limestones, shales, and sandstones that comprise the gray, stairstep cliffs of the inner gorges of Cataract Canyon in Canyonlands, where it is about 330 m thick. The Honaker Trail is present throughout the region in the subsurface, and like the Paradox, interfingers with Cutler-like arkose eastward from Moab to the Uncompahgre frontal faults. A well drilled by Phillips Petroleum Company on the Onion Creek structure, penetrated 8,198 ft of the formation, which contained very little limestone or other recognizably marine rocks.

Sedimentation was apparently continuous across the Permo-Pennsylvanian temporal boundary from the vicinity of Moab eastward to the Uncompahgre Uplift. However, toward the west a disconformity appears and truncates downward into the Honaker Trail until finally Permian rocks rest directly on Mississippian rocks on the San Rafael Swell (Emery Uplift). Early Permian strata rest on Virgilian limestones at the Arches park entrance near Moab, on Missourian rocks at the confluence of the Green and Colorado Rivers, on Des Moinesian age beds in wells drilled just east of the San Rafael Swell, and finally on the Mississippian on the Emery Uplift (Baars, 1962).

Permian System

Paleoenvironments were rarely static in Canyonlands country in Wolfcampian and Leonardian time. Facies relationships are varied and intricate throughout the system. In brief, every part of the section interfingers with one or more other parts. It is a physical stratigrapher's delight, as everything is so excellently exposed to view.

The Uncompahgre Uplift was still shedding arkose into the basin, forming large fluvial fan systems of coarse sand and gravel along the mountain front, and more distal distributary channels flowed freely along and around the rising topography of the salt anticlines as they grey. There may be as much as 6 km of arkose adjacent to the source area, thinning and becoming finer in grain size toward the west as each rim syncline was filled. These Cutler arkoses thin to zero or near zero on the active salt structures, sometimes by piercement and in some cases by depositional onlap.

West of the salt structures, the arkosic red beds interfinger with sediments derived elsewhere, and the Cutler can readily be subdivided into several formations. The lower of these is the Elephant Canyon Formation (Baars, 1962), which is a marine limestone-sandstone-shale section that interfingers with Cutler red beds along the canyon of the Colorado River from Moab to the confluence with the Green. Above that, a clean white sandstone, called the Cedar Mesa Sandstone, interfingers with the red beds along a broad zone between the confluence and the Needles District of Canyonlands National Park. In the Needles, the banded spires and fracture controlled knobs are interbeds of Cutler and Cedar Mesa. Some of the Cedar Mesa is believed to be marine and derived from somewhere to the northwest, but it becomes more eolian toward the top (John A. Campbell, personal communication, 1982). The overlying Organ Rock Shale grades westward from a fine-grained sandstone to a red mudstone across Canyonlands, and is a distinctive formation west of the Colorado River. The uppermost unit, the White Rim Sandstone, is a prominent bench-forming layer that does not occur east of the Colorado River. The marine-eolian formation pinches out eastward onto the west flank and northern plunge of the Monument Upwarp. A large offshore bar in Elaterite Basin west of the Colorado River is saturated with tar.

Triassic System

The Moenkopi Shale disconformably overlies the Cutler and White Rim Sandstone. It is typically a brown, tidal flat mudstone, with the whole array of ripples, mud cracks, burrows, and rain drop impressions. The sedimentary structures are best displayed in the stone walkways at the Visitors Center at Dead Horse Point State Park.

The overlying Chinle Formation is a multicolored mudstone-siltstone unit, with the exception of its basal Mossback Member. The Mossback is a sporadic channel-fill sandstone and/or conglomerate that contains uranium deposits in some localities. Ubiquitous bulldozer scars throughout Canyonlands are obvious "index fossils" to the Mossback Member. Shafer and Flint trails were constructed for access to exploration of Mossback uranium in the early 1950s. Upper members of the Chinle, the Petrified Forest, Owl Rock and Church Rock Members, are largely fluvial and lacustrine mudstones with the purple, gray, red, or brown coloration, and make large, extensive slopes beneath the massive Wingate cliffs.

Jurassic System

It now seems probable that the Triassic-Jurassic boundary occurs at the top of the Chinle Formation on the basis of palynomorphs found elsewhere on the Colorado Plateau (Fred Peterson, personal communication, 1982). It has previously been placed at the base or top of the Navajo Sandstone. A sandstone threesome, known as the Glen Canyon Group, forms the awesome cliffs that border Canyonlands country and make access to the low country so difficult. The basal sandstone is the largely eolian Wingate Sandstone that is seen in massive, vertical cliffs and buttes throughout the region. It is usually reddish-brown in color, and hosts well-developed desert varnish at most exposures. Overlying the Wingate is the ledgy, fluvial sandstones of the Kayenta Formation. It is actually the Kayenta that caps the "Wingate" cliffs and buttes in the region. Another massive, highly cross-stratified unit, the Navajo Sandstone, tops the Glen Canyon Group. It is commonly white and forms rounded cliffs and knobs topping the reddish cliffs. Cross-bedding suggests that it is largely of an eolian origin. In some localities, the Navajo contains thin, dense limestone beds that suggest playa lakes or algal marshes dotted the otherwise Sahara-like desert.

The San Rafael Group disconformably overlies the Navajo Sandstone. The basal unit is a brown siltstone-mudstone unit that characteristically erodes back to form benches at the top of the Navajo. It was long known as the Carmel Formation, but has more recently been lumped with the overlying Entrada Sandstone for mapping convenience. It is now called the Dewey Bridge Member of the Entrada Sandstone, a term not popular with this author. In this area, it is usually composed of highly contorted bedding, suggesting down-dip slumping prior to total lithification. The contortions involve the overlying sandstone, suggesting that the intertidal muds were buried to some depth by sand when the slumping occurred. The sandstone, the old Entrada Sandstone, now known as the Slick Rock Member, is much like the

Navajo, but is usually a light color. It forms massive, rounded cliffs, and contains the many arches in Arches National Park. This highly cross-bedded sandstone is truncated by a very prominent surface that is probably a minor disconformity. Above the surface is another massive, cliff-forming sandstone called the Moab Tongue of the Entrada. It is more uniformly bedded, and may be a facies of the marine Curtis Formation of the San Rafael Swell. The so-called Summerville Formation, another brown, well-bedded siltstone-mudstone, caps the group. Some of these terms may not be entirely correctly used, as the Jurassic correlations are undergoing intense scrutiny at this time.

The Morrison Formation caps the Jurassic System. Fluvial sandstones dominate the lower Salt Wash Member, which contains considerable amounts of uranium in this region. Like the Moss Back, this ledgy member has been heavily burrowed and scarred by miners. Varicolored shale marks the upper Brushy Basin Member of the Morrison. It looks like the Morrison all over the southern Rockies, being a mottled purple, green, red, and/or gray color, and forming rounded "badland" topography. Dinosaur bones are common in the Morrison Formation, and are quarried in several localities.

Cretaceous System

Rocks of Cretaceous age have been generally stripped from the Canyonlands-Arches region. Some are found collapsed into the salt anticlines, such as the south ends of Moab and Salt Valleys. A basal conglomerate or conglomeratic sandstone, the Burro Canyon (Cedar Mountain) Formation, overlies the Morrison. It is a resistant cliff-forming unit where exposed. The Lower Cretaceous fluvial deposit is disconformably overlain by the less resistant Dakota Sandstone. The Dakota represents shoreline deposits of the first invasion of the Cretaceous sea, and is generally carbonaceous and contains some coal. It is Late Cretaceous in age, somewhat younger than the Dakota of the Great Plains. The Dakota grades upward into the infamous Mancos Shale, a black, dull shale that one does not drive on when wet. Its high shrink-swell factor makes houses sink, roads and runways buckle, and people miserable. Fortunately, there is little Mancos left in Canyonlands country. It is overlain by and interfingers with, the complexities of the Mesaverde Group, that form the Book Cliffs north of this report area. Because we will visit no Cretaceous outcrops on this field trip, the section will be discussed no further.

ROADLOGS

This field trip is designed to view the eastern Colorado Plateau at its best. The first day will be spent west of Moab, Utah to study the typical stratigraphic sequence of the region from the Jurassic down through the Permian systems. The scenery is not bad either. We will visit Grandview Point for an overview of Canyonlands, Dead Horse Point, the Upheaval "salt dome," and then have a spectacular tour down Shafer Trail for a close-up look at the stratigraphy.

The second day will be spent in the spectacular salt-intruded anticlines of the Paradox Basin east of Moab. Moab Valley is one of the largest of the salt structures, and some of the stratigraphic evidence for it's growth history will be seen. Then, after a beautiful drive up the Colorado River, we will travel through an exposed leached gypsum cap in the Onion Creek - Fisher Valley salt anticline. On the return to Arches National Park we will make a brief visit into the Castle Valley diapir and on to some real Arches and the Salt Valley salt anticline before returning to Salt Lake City.

The road logs were compiled from various previously published guidebooks by the Four Corners Geological Society and the New Mexico Geological Society. Authors of these logs include C. M. Molenaar, D. L. Baars, J. A. Campbell, W. L. Chenoweth, and L. L. Craig in the order of frequency of contributions. The logs result from the many field trips and guidebooks published over the past 30 years on the region, and the contributions of numerous pioneers in the field are gratefully acknowledged.

FIELD TRIP ROAD LOG: FIRST DAY

Moab, Utah Westward to Dead Horse Point, Grand View Point, Upheaval Dome, and Shafer Trail

Driving Distance: 125 miles

Mileage Incre-mental	Mileage Cumu-lative	Description
0.0	0.0	Start at south traffic light in Moab.
0.7	0.7	The house of Charlie Steen, the uranium millionaire, at 1:00.

2.2	2.9	State Highway 128 to right; go straight across Colorado River.
0.3	3.2	Wingate-Chinle (Triassic) contact in road-cut.
0.5	3.7	At 12:00, strata on hillside above railroad are Virgil portion of Honaker Trail Formation (Pennsylvanian). The uranium processing plant of Atlas Minerals on the left.
0.8	4.5	Road Junction. Stay right. Road to left follows the river to the Texasgulf, Inc. potash mine on the Cane Creek anticline.
0.1	4.6	At 12:00, are Navajo Sandstone (Triassic?) and Entrada Sandstone (Jurassic). Note contorted bedding in the lower red Dewey Bridge Member of the Entrada.
1.3	5.9	Crossing Moab fault, a high-angle normal fault caused by salt flowage. The Entrada Sandstone is in fault contact with the upper part of the Honaker Trail Formation. Stratigraphic throw is about 610 m.
0.2	6.1	The section exposed here is the base of the Cutler Formation (Permian) and top 400 ft of the Honaker Trail. Virgil fusulinids were collected in one of the upper limestone beds. The total Honaker Trail in this area is about 580 m thick. Note the cyclic nature of the Honaker Trail and the various sedimentary features. Note truncation of the Cutler by the Moenkopi (Triassic) on hillside across the road.
0.1	6.2	Highway parallels Moab fault. Well location at wide spot on right ahead is Delhi-Taylor No. 2 which is a potash strat test. This well went into salt at 713 m and bottomed near the base of salt at 2,872 m. Delhi-Taylor has drilled about ten potash tests in this area (known as the Seven Mile anticline). The other wells went only a short way into the salt section. This structure is presently classed as a potash reserve.
1.7	7.9	Moenkopi-Cutler contact on left is at color break between brighter red Cutler and red-brown Moenkopi.
2.3	10.2	At 1:00, green hillside in distance is green shale of the Brushy Basin Member of the Morrison Formation (Jurassic). At 11:30 is fault contact (Moab fault) between Chinle on left and Brushy Basin on right, a stratigraphic throw of about 610 m.
1.1	11.3	Bridge. At 3:00, exhumed sandstone channel in Salt Wash Member of the Morrison. Current direction was easterly.
0.3	11.6	Road junction. Dead Horse Point road; turn left
0.3	11.9	At 1:00 the basal sandstone of the Chinle Formation is exposed near the base of a green shale where a uranium road has been cut. Uranium is being mined in this area from lenses of carbonaceous sandstone and mudstone pebble conglomerate in the basal Chinle Formation. The Moss Back Member of the Chinle, which is the host rock for many deposits to the south is not recognized here. The Moenkopi-Cutler contact is at the top of bright red, massive sandstone bed.
1.3	13.2	At 8 to 10:00, note thinning of

		Chinle strata in uppermost part above prominent sandstone bed and below Wingate.
0.7	13.9	Wingate-Chinle contact. Note large scale eolian crossbeds in Wingate.
1.1	15.0	Kayenta-Wingate contact.
0.9	15.9	Navajo-Kayenta contact. View of section from Wingate Sandstone (Triassic) up to Salt Wash Member of Morrison Formation (Jurassic). Moab Tongue intertonguing with Summerville; to the east it merges with the Entrada and to the west it pinches out a short distance west of the Green River.
1.7	17.6	At 4:30, note lenticular sandstone bar(?) "build-up" in Dewey Bridge Member of the Entrada.
0.4	18.0	Note contorted bedding in Dewey Bridge Member in cliffs at right.
1.6	19.6	In distance to north are the Book Cliffs which are held up by Mesaverde Group (Upper Cretaceous) dipping north into Uinta Basin.
1.1	20.7	At 3:00 in far distance is the San Rafael Swell. Navajo Sandstone hogbacks can be seen on the steep, east flank of the uplift.
1.3	22.0	Pure Big Flat No. 5 (Bartlett Flat field) on left. This well was completed for 450 BO/D in the Cane Creek marker, a "clastic" break within the salt section. This unit averages 50 to 100 ft (15 to 30 m) in thickness and consists of interbedded black shale, dense dolomitic shale, and anhydrite. Where productive, it is probably a fractured reservoir. A subcommercial amount of oil was recovered from the Mississippian. A few other wells in this general area have encountered production in the Cane Creek marker but in most cases, production rates declined rapidly after high initial production rates. One exception is the Southern Natural Gas Company, Long Canyon No. 1 which has produced over 400,000 barrels of oil since 1962. Current production rate is still about 200 BO/D. A one-half mile offset was dry in the Cane Creek marker.
1.5	23.5	Road-cuts are in Kayenta.
0.6	24.1	At 2:00, in far distance are the Henry Mountains, Tertiary stocks and laccoliths.
1.2	25.3	Road junction. Turn left. Small hill on left is Navajo Sandstone.
0.7	26.0	La Sal Mountains at 12:00. Abajo Mountains at 2:00 and Monument Uplift at 2:30. Wingate Sandstone crops out on top of uplift. About a mile to the right is Big Flat field, a small Mississippian field. This field was discovered by Pure Oil Company in 1957. Three wells were productive but the casing collapsed in two of the wells several years ago and caused their suspension or abandonment. The third well produced marginally until recently and it is currently shut in. Total production from the field is 81,500 barrels of oil. At the surface, Big Flat is a broad dome with about 60 m of closure; however, at the Mississippian level, it is complexly faulted.

1.1 27.1 Road junction. Turn right

4.9 32.0 Dead Horse point Visitor Center.

1.5 33.5 End of road. **STOP 1.** Walk out to rim for spectacular view overlooking the Colorado River and Canyonlands National Park to the southwest. Vertical drop to river is about 610 m. The section exposed is from Kayenta-Wingate, on top, to Honaker Trail in the river gorge to the left. Prominent bench below is a Permian limestone bed locally known as the "Shafer Limestone," a tongue of Elephant Canyon Formation. It pinches out a short distance to the east.

Note pinchout of White Rim Sandstone at top of Permian section on canyon walls to right. This unit forms prominent rim at intermediate bench around mesa to southwest. It is a beach-dune complex(?) with common large-scale, festoon cross-bedding separated by zones with planar bedding. It is largely marine. It thickens to the west and merges with the upper part of the "Coconino Sandstone" of the San Rafael Swell. The White Rim Sandstone is stained or saturated with tarry oil over a large area west of the confluence of the Green and Colorado Rivers.

Shafer Dome, which is to the left front, is a small salt structure. The Texasgulf, Inc. potash mining operation constructed the obvious evaporation pans on the west flank of the Cane Creek anticline for the precipitation of salt from brine wells in the anticline.

Turn around.

6.4 39.9 Return to Long Canyon road and turn left to Canyonlands National Park entrance and Grandview Point turnoff. Road is in Kayenta and Navajo Sandstone for 29 km.

18.1 58.1 Turn left to Grandview Point.

4.9 63.0 Grandview Point overlook. **STOP 2.** Good view of Canyonlands, including Monument Basin directly below the overlook, the Needles and confluence of the Green and Colorado rivers in the middle distance, and the crest of the Monument Upwarp on the skyline. The strata capping the overlook are in the Kayenta Formation; the prominent white bench below is held up by the White Rim Sandstone; the rocks exposed at the confluence are in the lower Permian Elephant Canyon Formation.

5.0 68.0 Return to road junction and turn left to Upheaval Dome.

4.0 72.0 Upheaval Dome. **STOP 3.** Hike about one-half mile on trail to viewpoint. Return to turnoff of Shafer Trail, a distance of about 27.5 km.

14.0 86.0 Junction with Shafer Trail — turn right.

0.5 86.5 Dead Horse Point at 9:00 with the La Sal Mountains in the distance.

0.2 86.7 Road is on a ledge in the Kayenta Formation.

0.5 87.2 Re-enter Canyonlands National Park.

0.5 87.7 **STOP 4.** Good view to left of Triassic and Permian strata exposed in Shafer Canyon.

0.9	88.6	Road descends the Shafer Trail. This trail was constructed in 1951 by the U.S. Atomic Energy Commission to provide uranium prospectors an access to the many uranium occurrences that had been found in the area, and as an incentive for mine development. Steep grade. Use low gear. Road descends through the Triassic section.
0.5	89.1	Top of Wingate Sandstone.
0.4	89.5	Top of Chinle Formation.
0.4	89.9	Top of "Black Ledge Member" of Chinle Formation.
0.1	90.0	Base of "Black Ledge Member."
0.1	90.1	Top of Moss Back Member of the Chinle Formation. This lenticular fluvial sandstone is an important source of uranium in this region.
0.1	90.2	Top of Moenkopi Formation.
0.3	90.5	Exit Canyonlands National Park.
0.1	90.6	Note ripples and mudcracks in the Moenkopi redbeds.
0.4	91.0	Dead Horse Point at 1:00.
0.2	91.2	Junction — turn left toward Moab, a distance of 50 km. Road descends through the White Rim Sandstone, which is here eolian in the upper half, and the undifferentiated Cutler redbeds.
1.8	93.0	"Shafer Limestone" exposed in Shafer Wash on the right.
0.3	93.3	Shafer Dome, a salt-intruded anticline, is obvious at 1:00 to 3:00.
0.2	93.5	Colorado River visible below on right.
0.2	93.7	Axis of Shafer dome visible at 2:00 in undifferentiated Cutler Formation.
0.4	94.1	Dead Horse Point is high above at 11:00. Note that the White Rim Sandstone has pinched out.
1.5	95.6	Fence. Road descends along north flank of Shafer Dome. The La Sal Mountains are visible at 2:00.
0.8	96.4	Dead Horse Point visible at 8:00.
2.2	98.6	First evaporation pond of the Texasgulf, Inc. potash operation. Brine is pumped from the Paradox Formation beneath the Cane Creek anticline, which is visible beyond the pond.
0.2	98.8	Gate.
1.3	100.1	**STOP 5.** Good view of Texasgulf, Inc. evaporation ponds and the Cane Creek anticline.
0.4	101.5	Road descends through a red, eolian sandstone of the Cutler Formation.
0.4	101.9	Road crosses brine pipeline.
0.3	102.2	Road ascends west flank of Cane Creek anticline on a thin, marine limestone bed of Permian age.
1.0	103.2	Site of old oil wells, drilled mainly for promotional purposes by M. G. Mason. The wells produced small amounts of oil from the "Cane Creek" marker in the Paradox Formation.

0.3	103.5	Note the mine dump of the Section 32 uranium mine across the canyon ahead. Uranium is mined from a small northwest-trending channel in the Moss Back Member of Chinle.
0.2	103.7	Road descends east flank of the Cane Creek anticline. Note the numerous thin limestones in the Cutler. These represent far easterly tongues of the Elephant Canyon Formation, interfingering with the arkosic facies of the Cutler.
0.5	104.2	Boat launching ramp on right. Float trips through Cataract Canyon leave from this point.
0.2	104.4	Several well locations may be seen in this area; wells were drilled as water-injection wells for the Texasgulf, Inc. potash recovery operation.
1.1	105.5	Texasgulf Chemicals potash mine on left.
0.1	105.6	Pavement begins.
0.8	106.4	Top of Cutler on left.
1.0	107.4	The Jug Handle and top of Chinle Formation on left.
1.5	109.9	Approximate axis of King's Bottom syncline.
4.3	114.2	Dinosaur tracks in Kayenta Formation high on left.
0.9	115.1	Numerous petroglyphs in desert varnish on the Navajo Sandstone on left.
0.9	116.0	The abrupt west flank of the Moab salt-intruded anticline ahead. Road descends through the Glen Canyon Group.
1.1	117.1	Wingate-Chinle contact.
0.4	117.5	Outcrops of Honaker Trail and Paradox Formations low on the cliff at 4:00. Good view down Moab Valley to the La Sal Mountains.
0.7	118.2	Chinle-Hermosa unconformity on left. (Think that one over!)
1.0	119.2	Tailings pond on right for Atlas Minerals Corporation uranium mill.
0.7	119.9	Road crosses Moab fault.
0.2	120.1	Junction. Turn right to Moab.
1.3	121.4	Bridge over Colorado River.
2.0	123.4	Entering Moab, Utah.

END OF FIRST DAY OF FIELD TRIP.

FIELD TRIP ROAD LOG: SECOND DAY

Moab, Utah Eastward to Onion Creek, Castle Valley, and Arches National Park

Driving Distance 122.3 miles, return to Salt Lake City

Mileage Incremental	Mileage Cumulative	Description
0.0	0.0	Start at south traffic signal in Moab, which lies in the collapsed valley of a large salt diapir; salt is from the Pennsylvanian Paradox Formation.
1.0	1.0	House on hillside at right belonged to Charlie Steen who discovered the Mi Vida uranium mine in the Big Indian district.
0.7	1.7	At left in flat, the Suburban Gas Company has a L. P. gas storage reservoir in salt at about 610 m.
0.3	2.0	Note rollover at 11:00 at end of

		valley, where Navajo, Kayenta, and Wingate Formations are exposed. The Moenkopi Formation (Triassic) unconformably overlies the Honaker Trail Formation (Pennsylvanian) at 9:00.
0.9	2.9	Road junction, Utah State Highway 128. Turn right. Canyon walls are the Navajo Sandstone (Jurassic?), Kayenta Formation, and Wingate Sandstone (Triassic).
1.7	4.6	Axis of Courthouse syncline. Massive sandstone is Navajo.
1.5	6.1	Navajo-Kayenta contact at 10:00. Small Indian ruins are in caves at base of upper massive part of Navajo.
0.4	6.5	Road cuts in the Kayenta.
0.2	6.7	Top of massive sandstone is top of Wingate.
0.6	7.3	Wingate-Chinle (Triassic) contact is at break in slope at 9:00 across river.
1.3	8.6	Chinle-Moenkopi (Triassic) contact is across river at color break.
0.6	9.2	Chinle-Moenkopi contact is up gully at 3:00 at base of white marker bed.
1.3	10.5	Chinle Moenkopi contact is at white bed, halfway up slope at 11:00 to 12:00.
2.8	13.3	Note unconformity at base of the Chinle across river. This feature continues for next three miles and is believed to be a local unconformity due to salt tectonics during Triassic time. Here, the underlying Moenkopi(?) beds dip easterly away from an unexposed salt structure.
0.9	14.2	Salt Wash Creek coming in on left.
1.7	15.9	Unconformity at 9:00; underlying beds here dip westerly away from the Castle Valley salt structure.
0.6	16.5	Note thinning of prominent sandstone in Chinle at 9:00. Approaching northwest plunge of Castle Valley salt anticline.
0.3	16.8	Road is in Moenkopi Shale.
0.4	17.2	Castle Creek bridge. Plunging nose of Castle Valley structure ahead.
1.1	18.3	At 3:00 is small monoclinal flexure in Moenkopi associated with the Castle Valley salt structure.
0.5	18.8	Road junction. Castle Valley road. Go straight.
1.0	19.8	Down-faulted graben zone at left. This is on the Salt Valley-Cache Valley-Onion Creek trend and is due to solution-collapse in the underlying salt.
1.1	20.9	Moenkopi-Cutler (Permian) contact near base of hill at 9:00. Roadcuts ahead are in Moenkopi in graben zone.
0.9	21.8	Road junction. Stay left. La Sal Mountains at 3:00.
0.7	22.5	Cutler crops out in low hills along river.
1.1	23.6	Road junction. Turn right to Fisher Valley. After turn, Fisher Towers are at 11:00. Towers are mostly Cutler, but a little Moenkopi caps the large tower. The road is in fluvial

Cutler arkose and conglomerate for the next four miles.

0.8 24.4 At 3:00 the Wingate spire and butte are known as the Priest and Nuns.

0.8 25.2 **STOP 1.** Discussion of geologic setting at northwest plunge of Onion Creek-Fisher Valley salt structure. Both the Harry Hubbard, No. 1 Government-Campbell (12-T24S-R23E) and the Phillips, No. 2 Onion Creek (13-T24S-R23E) wells are located within a mile to the north. Both wells spudded several hundred feet below the top of the Cutler Formation. The Hubbard well drilled a few thin limestone stringers starting at about 1,220 m, but drilled dominantly arkose to TD (7,955). The Phillips well encountered thin sandy limestones from 1,319 to 3,767 m, but the vast majority of that interval was arkose. Neither well drilled any primary evaporites, presumably because of salt flowage into the Onion Creek diapir. The lithology log by AMSTRAT on the Phillips well indicates a faulted zone at 3,767 to 3,780 m, which may represent the flowed-salt section. If so, the well penetrated 3,767 m of mostly arkose above the Paradox evaporites. It bottomed in Cambrian quartzite at 4,362 m, after penetrating a normal pre-Pennsylvanian section. The Richfield, No. 1 Onion Creek well (31-T23S-R24E) was drilled farther down the northeast flank of the Onion Creek structure than the others and hit Honaker Trail(?) carbonate rocks at 1,444 m, and drilled about 1,700 ft (518 m) of salt below 11,555 ft (3,522 m), bottoming in Mississippian rocks at 13,922 ft (4,243 m). The section was again mostly arkose above the salt. The very thick Honaker Trail sections in all three wells suggest that much of the salt flowage occurred during Honaker Trail time (mid-Desmoinesian-Virgilian).

2.2 27.4 Bridge across Onion Creek.

0.3 27.7 **STOP 2.** Discussion of sedimentary structures in the Cutler.

0.2 27.9 Onion Creek diapir ahead.

0.3 28.2 Note collapse structure on right. Wingate blocks dropped down. Moenkopi-Cutler contact is high on hill at top of purplish beds.

0.9 29.1 Leached Paradox on-left. We will drive through this whole mess and stop at other end to look back on it.

1.9 31.0 Cutler-Paradox contact on left.

1.8 32.8 Entering Fisher Valley. Turn around past fence and **STOP 3.**

The flat valley surface of Fisher Valley is a late Pleistocene and Recent valley fill. It overlies leached caprock of the Fisher Valley salt cell. The prominent cliffs surrounding the valley are Wingate. About 600 ft (183 m) of Honaker Trail is exposed on the southwest flank of Fisher Valley. A fault and graben zone which trends northeasterly from here is probably directly or indirectly associated with the salt tectonics of Onion Creek-Fisher Valley structures. Although smaller than many of the breached salt anticlines, the Onion Creek structure is one of the most spectacular.

One of the conjectural aspects of the salt anticlines is their genesis. Perhaps the best hypothesis ascribes differential

		loading as the primary mechanism with pre-salt lineaments determining the present alignment. Recent deep drilling and seismic surveys have shown that the different structures had their major growth at different times; the ones closer to the Uncompahgre front being earlier. In fact, on most structures the salt flowed from the northeast flanking syncline before flowage occurred from the southwest flank. This evidence strongly supports the differential loading hypothesis.
1.3	34.1	Picture stop. Paradox-Cutler contact at 12:00. Fault contact due to leaching and collapse or due to diapiric salt flowage?
1.9	36.0	Contorted leached Paradox outcrops.
0.4	36.4	Fisher Tower at 12:00. Note Moenkopi-Cutler contact high on the spire.
1.7	38.1	Note fantastic shapes of Cutler outcrops.
3.9	42.0	Road junction, State Highway 128. Turn left.
0.4	42.4	Note truncation of Cutler by Moenkopi at 1:00.
4.2	46.6	Castle Valley road junction. Turn left. Road is on Moenkopi.
1.9	48.5	Entering Castle Valley. Stay on paved road.
0.5	49.0	**STOP 4.** The prominent thick white sandstone across Castle Valley is an eolian sand at the top of the Permian section. Note obvious truncation by the overlying Moenkopi up the valley. Around corner to right at plunge of anticline, this sandstone is faulted down and is not present on northeast flank of Castle Valley. Hill in center of valley to the southeast is Tertiary igneous stock with leached Paradox around the base. Wingate Sandstone forms the upper cliff on both sides of the valley. The deepest penetration in Castle Valley was a dry hole by Gold Bar Resources, located in the valley north-northwest of the intrusive. The well bottomed in salt at 6,502 ft (1,982 m).
1.0	50.0	At 11:00 the spire on right is Castle Rock. The group to the left is the "Priest and Nuns." Note that the massive white sandstone is not present at the top of the Cutler on this side of the valley. Turn around at intersection of road and continue back toward Moab.
3.2	53.2	Road intersection, State Highway 128. Turn left toward Moab.
15.8	69.0	Junction with U.S. Highway 169. Turn right.
0.1	69.1	Bridge over Colorado River. Collapse structures along the road ahead.
0.5	69.6	Atlas Minerals Corporation uranium-vanadium mill on the left. This mill is rated at 1,270 tons of ore per day. It has two circuites; an acid leach with solvent extraction for low lime/high vanadium ores of the Salt Wash; and a carbonate leach with caustic precipitation for the high lime/low vanadium ores of the Chinle Formation.
0.1	70.6	Road to left follows the river to the Texasgulf, Inc. potash mine at the Cane Creek anticline.

0.6	71.2	Entrance to Arches National National Park. Turn right.
0.3	71.5	Visitors center on the right.
0.3	71.8	Entrada, Dewey Bridge Member, and Navajo contact on the right.
1.0	72.8	Road cuts in the Navajo.
1.0	73.8	South Park Avenue turn-out in the Dewey Bridge.
0.4	74.2	La Sal Mountains viewpoint on right. Road continues near Dewey Bridge-Navajo contact. Arch on the skyline to the east (right).
1.0	75.2	Courthouse Tower viewpoint turnoff.
0.8	76.0	Courthouse Wash Bridge.
0.8	76.8	Dewey Bridge-Navajo contact well exposed along the road.
0.8	77.6	Petrified dunes viewpoint.
2.9	80.5	Balanced rock. Prepare to turn.
0.2	80.7	Road junction, turn right. Road to left goes to Cache Creek graben.
0.5	81.2	View to the left, 9:00, of Salt Valley salt anticline, another collapse structure.
1.9	83.1	Turnout at end of loop, **STOP 6.** Brief discussion of geology, but mainly a picture stop. Arches have formed in the Entrada Sandstone. Return to U.S. Highway 163 at entrance to park.
0.2	83.3	Double Arch and Parade of the Elephants on right.
2.3	85.6	Salt Valley salt anticline at 1:30.
0.1	85.7	Road junction, turn left to visitors center.
9.3	95.0	Visitors Center.
0.2	95.2	Road junction with U.S. Highway 163, turn right.
0.8	96.0	At about this point road crosses Moab fault, a high-angle normal fault caused by salt flowage. The Entrada is in contact with the upper part of Honaker Trail Formation, with a throw of about 610 m on the fault.
0.2	96.2	Section exposed here includes top 122 m of the Honaker Trail; 130 m of Cutler, then Moenkopi, Chinle, and Wingate cliff.
0.2	96.4	Highway parallels Moab fault and goes up Moab Valley. Well location at wide spot on right ahead is Delhi Taylor No. 2.
1.6	98.0	Moenkopi-Cutler contact on left is at color break between brighter red Cutler and red-brown Moenkopi.
2.3	100.3	Green hillside at 1:00 in distance is green shale of the Brushy Basin. At 11:30 is fault contact of Moab fault with Chinle on left and Brushy Basin on the right, a stratigraphic throw of 610 m (2,000 ft).
1.1	101.4	Bridge. At 3:00 sandstone channel in Salt Wash Member.
0.3	101.7	Road junction. Dead Horse Point road to the left. Stay on U.S. 163.
0.3	102.0	At 9:00 basal sandstone of the Chinle Formation is exposed near base of green shale where a uranium road has been cut.

Uranium was mined from lenses of carbonaceous sandstone and mudstone pebble conglomerate in the basal Chinle. The Moss Back Member of the Chinle is not recognized here. Dark-colored sandstone in upper part of Chinle is locally called the "black ledge."

0.1 102.1 Road for next 6 km is on westward-dipping Morrison Formation. Salt Valley salt anticline is to the northeast.

3.0 105.1 Road is in Brushy Basin Member of the Morrison Formation.

0.8 105.9 Approximate contact between Brushy Basin and Burro Canyon (Cedar Mountain). Road in the Burro Canyon.

1.1 107.0 Dakota-Burro Canyon (Cedar Mountain) contact about here.

0.2 107.2 Road is on Mancos Shale from here to Crescent Junction with west-dipping Dakota-Burro Canyon (Cedar Mountain) on the right.

1.6 108.8 Airport Road on the left.

1.2 110.0 Ferron Sandstone Member of Mancos at 9:00. Entrada dip slope at 3:00.

8.2 118.2 Crossing the Salt Valley graben which is the north end of the Salt Valley anticline. Looking down Salt Valley at 5:00.

4.1 122.3 Turn left onto I-70 at the Crescent Junction interchange. Return to Salt Lake City. Book Cliffs on the right are formed by the rocks of the Mesaverde Group which overlie the Mancos Shale. In this area, the Mesaverde is composed of the Blackhawk Formation, Castlegate Sandstone, Sego Sandstone, Nelsen Formation, Farrer Formation, and the Tuscher Formation. The Buck Tongue of the Mancos occurs between the Castlegate and Sego Sandstone. A sandstone of the upper Blackhawk and the Castlegate Sandstone form the lowest rim of the Book Cliffs.

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