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PROTEROZOIC TO RECENT STRATIGRAPHY, TECTONICS, AND VOLCANOLOGY, UTAH, NEVADA, SOUTHERN IDAHO AND CENTRAL MEXICO

Edited by

Paul Karl Link and Bart J. Kowallis

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CONTENTS

Triassic-Jurassic Tectonism and Magmatism in the Mesozoic Continental Arc of Nevada: Classic Relations and New Developments

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ABSTRACT

This field trip will focus on the early Mesozoic structural and magmatic evolution of the Black Rock Desert region of northwest Nevada, which forms part of the Mesozoic magmatic arc assemblages of the western U.S. Cordillera, and on deformed back-arc basinal strata of the Luning-Fencemaker fold-and-thrust belt, located east of the Black Rock Desert. The principal goals of this trip are as follows. (1) To demonstrate that the majority of early Mesozoic magmatic rocks of the Black Rock Desert region are of Late Triassic to Early Jurassic age—this contrasts with other arc assemblages of the Cordillera where the most voluminous early Mesozoic magmatism occurred in the Middle to Late Jurassic. (2) To demonstrate that Mesozoic shortening deformation in the Black Rock Desert is primarily of Early Jurassic age and that there is little evidence for shortening in the Middle to Late Jurassic—this contrasts with other arc assemblages of the Cordillera where major Mesozoic shortening occurred in the Middle to Late Jurassic. (3) To examine the nature of Jurassic deformation in the Black Rock Desert, which involved development of a large-scale ductile thrust zone at shallow mid-crustal levels (~8–10 km depth). This ductile thrust and its less deformed upper plate are well-exposed in the western Black Rock Desert. (4) To examine structures in the back-arc Luning-Fencemaker fold-and-thrust belt, and investigate the possibility that much of the deformation in these rocks may be coeval with Early Jurassic shortening in the Black Rock Desert arc province, rather than Middle to Late Jurassic as has been generally inferred in previous studies. (5) To explore the implications of these relations for the Mesozoic tectonic evolution of the western U.S. Cordillera.

INTRODUCTION

Early Mesozoic magmatic arc assemblages in the western U.S. Cordillera record events along an active plate margin that developed following accretion of Paleozoic arc terranes to the continent during the Permo-Triassic Sonoma orogeny. Traditionally, these arc assemblages can be divided into two groups (e.g., Schweickert, 1978; Saleeby and Busby-Spera, 1992). (1) Arc assemblages with basement or stratigraphic ties to the continent. These "continental arc assemblages" (Fig. 1) clearly developed on the North American plate above an east-dipping subduction zone. They vary from subaerial to marine and were built on continental crust to the south (southern California and Arizona) and accreted Paleozoic terranes to the north. (2) Arc assemblages without clear basement or stratigraphic ties to the continent. These "western arc assemblages" (Fig. 1) are entirely marine and are typically associated with subduction complex assemblages and ophiolite sequences, implying development in an oceanic environment. They may be the offshore continuation of the continental arc or they may be allochthonous. The continental arc assemblages, because of their undisputed origin on the continental margin, represent a crucial source of information about events affecting the margin during the early Mesozoic. Only when these continental margin events are fully understood can the evolution of the western arc assemblages be confidently linked to the continent.

Until recently, most of the work on the continental arc assemblages has been in the Klamath Mountains, Sierra Nevada, southeastern California and southwestern Arizona, while comparatively less was known about the Black Rock

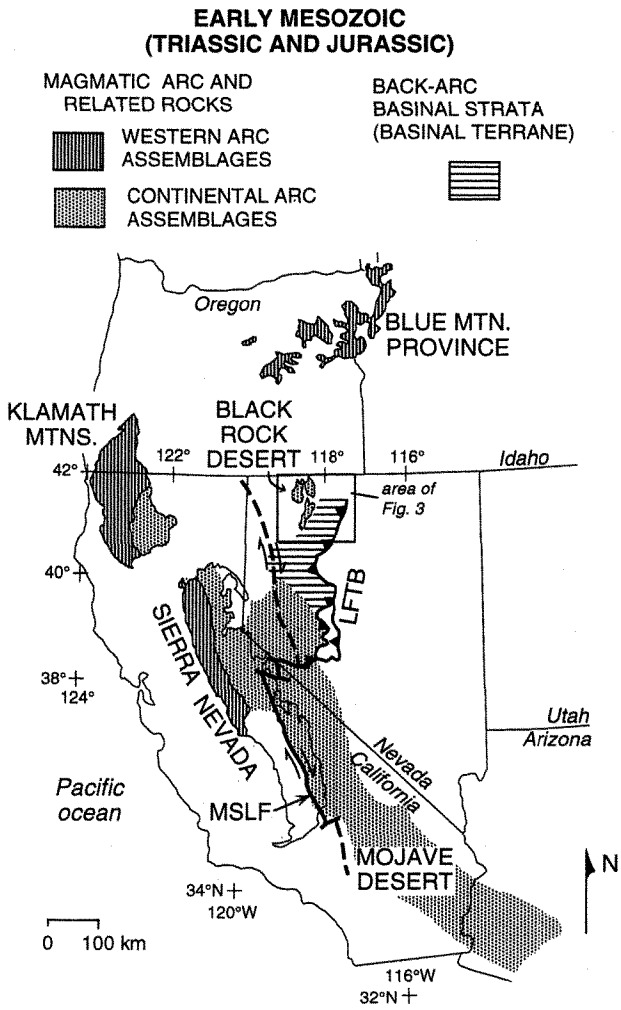


Figure 1. Map of western U.S. Cordillera showing distribution of early Mesozoic arc and back-arc basin assemblages, frontal thrusts of Luning-Fencemaker fold-and-thrust belt (LFTB; Oldow, 1984), and Mojave-Snow Lake dextral strike-slip fault (MSLF; solid line where well located, dashed line where inferred; Schweickert and Lahren, 1990).

Desert segment of the continental arc in northwest Nevada (Fig. 1). This area is an important source of information, however, about events affecting the active continental margin. First, it contains well-exposed and extensive sections of Mesozoic arc rocks and Paleozoic arc basement, collectively spanning over 200 m.y. of geologic history, from the mid-Paleozoic to the mid-Cretaceous (Ketner and Wardlaw, 1981; Russell, 1984; Maher, 1989; Wyld, 1990; Quinn, 1996). Second, it is directly adjacent to a back-arc marine province (see basinal terrane in Fig. 1) that was the site of major shortening deformation in the Jurassic (Luning-Fencemaker fold-and-thrust belt; Speed, 1978; Oldow, 1984). The evolution of the Black Rock Desert province may thus provide insight into the development of the back-arc fold-and-thrust belt. Finally, recent studies in the Black Rock Desert demonstrate that it experienced a substantially different early Mesozoic magmatic and structural history from other continental arc assemblages to the west and the south; specifically, early Mesozoic shortening deformation and voluminous magmatism occurred tens of millions of years earlier in the Black Rock Desert than in other arc provinces. This suggests a more complex tectonic evolution for the early Mesozoic continental margin than has previously been recognized, as explained further below.

On this field trip, we will examine stratigraphic, structural and magmatic relations in the Black Rock Desert and adjacent basinal terrane. Our goals are several. (1) To visit classic exposures of the early Mesozoic magmatic arc and its Paleozoic basement in the Black Rock Desert. (2) To examine the effects of Mesozoic shortening deformation on this arc assemblage, which included development of a large-scale ductile shear zone at shallow mid-crustal levels. (3) To demonstrate that the timing of early Mesozoic shortening and voluminous arc magmatism in the Black Rock Desert is different from that of the Klamath Mountains-to-Arizona portion of the continental arc, and to explore possible reasons for this disparity. (4) To examine part of the deformed basinal strata of the Luning-Fencemaker fold-and-thrust belt, and consider how deformation in these rocks may be linked with deformation in the Black Rock Desert.

REGIONAL GEOLOGIC FRAMEWORK

Studies in continental arc assemblages of the Klamath Mountains, Sierra Nevada, west-central Nevada, southeast California and southwest Arizona support the following interpretations about intra-arc tectonism during early Mesozoic time. (1) Arc magmatism and sedimentation during the Triassic to Early Jurassic occurred in either a tectonically neutral (experiencing neither shortening or extension) or actively extensional environment (e.g., Busby-Spera, 1988). This conclusion is based on the preservation, within these areas, of continuous Triassic to Lower Jurassic stratigraphic sections with no evidence for coeval shortening deformation, and, locally, on direct structural or stratigraphic evidence for extension (Sanborn, 1960; Busby-Spera, 1988; Harwood, 1992; Saleeby and Busby-Spera, 1992; Wyld and Wright, 1993). (2) Intra-arc shortening deformation was widespread during the Middle to Late Jurassic, affecting all the areas listed above except southeasternmost California and southwest Arizona where extension continued (Harper and Wright, 1984; Oldow, 1984; Sharp, 1988; Wright and Fahan, 1988; Walker et al., 1990; Saleeby and Busby-Spera, 1992; Smith et al., 1993). This contractional deformation occurred, depending on specific locality, sometime during the interval 170–150 Ma and was accompanied by volumi-

nous arc magmatism. How this intra-arc deformation relates to events farther inland is somewhat less clear. Deformation of the early Mesozoic back-arc basin succession of central Nevada in the Luning-Fencemaker fold-and-thrust belt is interpreted to be coeval with Middle to Late Jurassic shortening to the west (Oldow, 1984; Speed et al., 1988); however, the timing of deformation in the fold-and-thrust belt is not tightly constrained and correlations with events to the west are thus uncertain.

New data from the Black Rock Desert (summarized here and discussed in detail in later sections) are at odds with relations farther west and south within the arc assemblages and require reinterpretation of events affecting the active Cordilleran margin in the early Mesozoic; these new data also suggest alternative interpretations of the deformational record within the back-arc region (Wyld et al., 1996). First, the principal Jurassic shortening deformation in the Black Rock Desert occurred in the Early Jurassic at ~200 Ma, and there is no clear evidence for any significant deformation in this area during the Middle to Late Jurassic (Quinn, 1996; Wyld, 1996; Wyld et al., 1996). Thus, the Black Rock Desert segment of the continental arc experienced major shortening deformation some 30–50 m.y. earlier than shortening in arc assemblages to the west and south. Furthermore, shortening deformation in the Black Rock Desert occurred at the same time that arc assemblages to the west and south were either extensional or tectonically neutral. Second, the principal period of voluminous Jurassic arc magmatism in the Black Rock Desert occurred in the Early Jurassic, whereas there is only limited evidence of Middle to Late Jurassic arc magmatism (Wyld, 1991a, 1996; Quinn et al., 1997).

These differences led Wyld et al. (1996) to conclude that the Black Rock Desert and other arc assemblages did not evolve together in the Jurassic in the same spatial arrangement in which they are now organized. Building on the work of Schweickert and Lahren (1990, 1993) who argued that a Cretaceous strike-slip fault with ~450 km of right-lateral displacement is located within eastern California (Mojave-Snow Lake fault) and that this fault or related structures may have continued northward between the Black Rock Desert and Klamath Mountains (Fig. 1), Wyld et al. (1996) concluded that the Black Rock Desert probably lay entirely north of the Klamath Mountains and more southerly arc assemblages during the Jurassic (Fig. 2). Differences in the Jurassic structural and magmatic evolution between the Black Rock Desert segment of the continental arc and the Klamath Mountains-to-Arizona segments can thus be viewed as reflecting north to south variations in events or processes affecting the active plate margin in the Jurassic. Two possible explanations were advanced by Wyld et al. (1996). (1) The Early Jurassic plate margin was curved. Given an appropriate relative-convergence vector between the North American plate and the subducting oceanic plate, this curvature could result in a large orthogonal component of convergence in the vicinity of the Black Rock Desert, which would promote shortening deformation in the upper plate, but a larger transverse component to the south, which would promote strike-slip faulting in the upper plate and could account for a record of neutral to extensional tectonism. (2) Restricted collision of an island arc or some other crustal mass occurred along the northern part of the plate boundary (in the vicinity of the Black Rock Desert) but did not affect the plate margin farther to the south. One intriguing aspect of this reconstruction is that the Luning-Fencemaker fold-and-thrust belt occupies a back-arc position relative to the Black Rock Desert segment of the continental arc, but lies mostly north of the other arc assemblages (Fig. 2). Because shortening deformation in the Black Rock Desert occurred in the Early Jurassic, this reconstructed spatial arrangement suggests that Luning-Fencemaker deformation may in part be Early Jurassic, rather than Middle to Late Jurassic as is generally inferred.

GEOLOGY OF NORTHWEST NEVADA AND THE BLACK ROCK DESERT

The geology of northwest Nevada is shown in Figure 3. This area is located near the western margin of the Basin and Range province. The Black Rock Desert area includes three mountain ranges: the Pine Forest Range, the Bilk Creek Mountains and the Jackson Mountains (Fig. 3). Paleozoic and Mesozoic rocks of magmatic arc affinity are exposed in all three ranges (Ketner and Wardlaw, 1981; Russell, 1984; Jones, 1990; Wyld, 1990; Quinn, 1996). Paleozoic strata, a combination of sedimentary and less common volcanic rocks, range in age from Late Devonian(?) or Mississippian to Permian. Mesozoic strata are mostly of latest Triassic to Early Jurassic age and are dominated by mafic to intermediate, subduction-related volcanogenic rocks; however, some older Triassic and some Lower Cretaceous strata, mostly sedimentary, are locally present. Numerous Mesozoic plutons, mostly of Jurassic and Cretaceous age but also including some that are Triassic, intrude the Paleozoic and Mesozoic strata (Fig. 3); these intrusions attest to the generally intra-arc position of the Black Rock Desert during the entire Mesozoic.

East and south of the Black Rock Desert are a number of ranges that are underlain in part by thick successions of Triassic sedimentary strata (Triassic basinal strata; Fig. 3). These strata, which were deposited in a deep marine back-arc basin, consist mostly of mudstone with lesser amounts of intercalated quartz sandstone and carbonate, and constitute part of the basinal terrane shown in Figures 1 and 2 (Speed, 1978; Lupe and Silberling, 1985). Basinal terrane

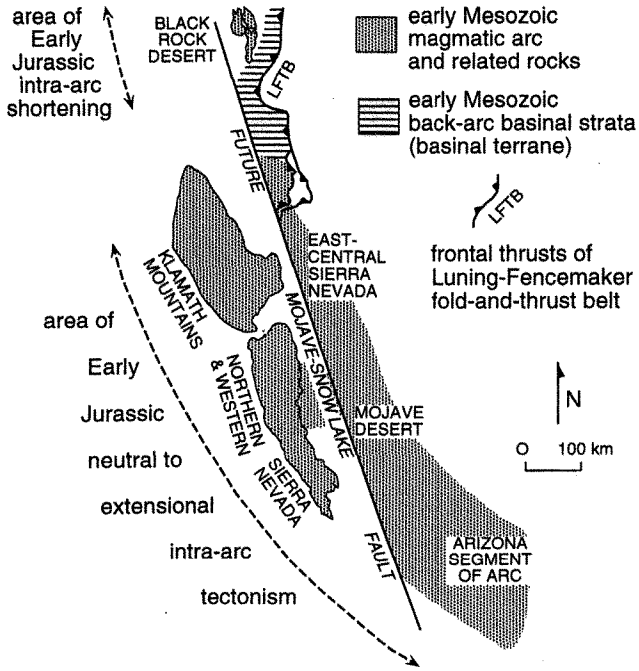


Figure 2. Locations of arc provinces, back-arc basinal terrane and Luning-Fencemaker fold-and-thrust belt reconstructed to show relative positions prior to inferred displacement along the Early Cretaceous Mojave-Snow Lake dextral strike-slip fault (based on Fig. 2 of Schweickert and Lahren, 1990).

strata were deformed in the mid-Mesozoic in the Luning-Fencemaker fold-and-thrust belt which encompasses the entire basinal terrane (Fig. 1; Oldow, 1984). In northwestern Nevada, the frontal thrust of the fold-and-thrust belt is called the Fencemaker thrust and it places Triassic basinal strata over coeval shelf deposits (Fig. 3). Paleozoic basement for the shelf sequence includes the Golconda and Roberts Mountains allochthons (Fig. 2; see ranges east of Winnemucca).

DAY 1: GEOLOGIC OVERVIEW; AND SOUTHERN PINE FOREST RANGE

Road Log, Winnemucca to the Pine Forest Range

0.0 Start mileage in downtown Winnemucca at the intersection of Winnemucca Blvd. and Melarkey St. (See Figure 3 for location of Winnemucca.) Proceed west on Melarkey, crossing the Humboldt River. This road turns north after crossing the river, at which point it becomes state Highway 95. As you drive north on Highway 95, the mountains to the east across a wide valley are the Hot Springs Range and Osgood Mountains, underlain primarily by rocks of the Golconda allochthon, Roberts Mountain allochthon and early Paleozoic miogeocline (Fig. 3). To the west is Winnemucca Mountain (closest to Winnemucca) and then the Bloody Run Hills (farther north), which are underlain mostly by Triassic strata of the basinal terrane. The prominent, light-colored outcrops in the Bloody Run Hills are Cretaceous plutons.

31.0 Turn left onto state Highway 140, heading toward Denio and Lakeview. Pull over into parking area at highway junction.

Stop 1.1 Geologic Overview

This stop is located on the Andorno Ranch 7.5 minute topographic map and is shown on Figure 3. From the vantage point at this stop, the geologic framework of northwest Nevada is readily observed. To the east and south are the Santa Rosa Range and Bloody Run Hills, both classic areas of exposure of rocks of the basinal terrane. These rocks consist mostly of weakly metamorphosed mudstone, which forms either dark outcrops or, more commonly, is poorly exposed on the grass-covered hillslopes; quartz-rich sandstone and carbonate beds are locally present but uncommon overall (Compton, 1960). Petrologic and facies studies indicate that the sediments of the basinal terrane were derived from the continental interior (Lupe and Silberling, 1985). Total structural thickness is estimated to be as much as 6 km, and biostratigraphic data indicate that the sediments were deposited mostly in the Norian, but also locally in the earliest Jurassic (Compton, 1960; Willden, 1964; Speed, 1978; Lupe and Silberling, 1985).

The most prominent structure formed in rocks of the basinal terrane during development of the Luning-Fencemaker fold-and-thrust belt is a slaty to phyllitic cleavage (Compton, 1960; Oldow, 1984; Elison and Speed, 1989). This cleavage is typically axial planar to folds that reflect transport to the southeast. Shortening associated with these structures is interpreted to have resulted in at least partial closure of the back-arc basin coupled with thrusting of basinal strata southeastward over coeval shelf strata. Younger folds, local cleavage, and thrusts also formed during development of the Luning-Fencemaker fold-and-thrust belt, but are less prominent and record less shortening (Oldow, 1984; Elison and Speed, 1989). The minimum age of deformation in the basinal terrane is provided by granitoid plutons that intrude across Luning-Fencemaker structures; some of these plutons have been dated as mid-Cretaceous (~104–70 Ma; K–Ar hornblende and biotite dates; Smith et al., 1971; Willden and Speed, 1974; Oldow, 1978, 1984). Several of these post-tectonic plutons are present in the Santa Rosa Range and Bloody Run Hills (Fig. 3; Compton, 1960); they form the well-exposed, cliffy, light-colored outcrops. The maximum age of deformation in the Luning-Fencemaker belt is late-Early Jurassic based on the youngest

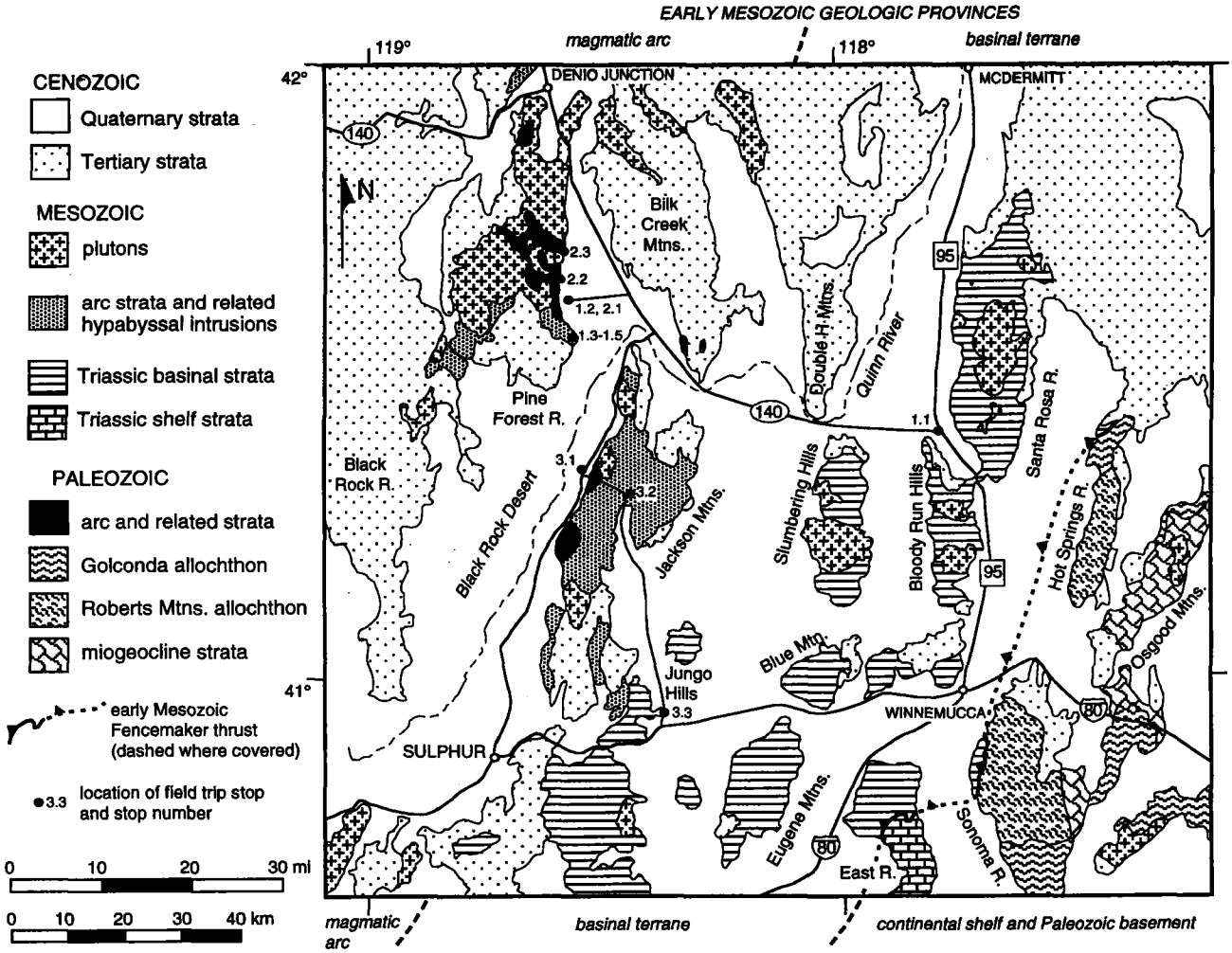


Figure 3. Location and summary geology map of northwest Nevada.

age (Toarcian) of basinal terrane rocks deformed in the fold-and-thrust belt (Speed, 1974, 1978).

To the west are low hills and ranges underlain by Tertiary strata (Double H Mountains) and more strata of the basinal terrane (Slumbering Hills). Farther west are the high ranges of the Black Rock Desert. Little work has been done on the Cenozoic extensional history of this part of the Basin and Range province, but it appears to be an area of relatively low-magnitude extension: Tertiary strata are still widely exposed in the ranges and some ranges have not been uplifted sufficiently to expose any older rocks (Fig. 3); multiple episodes of Cenozoic normal faulting and rotated normal faults have not been recognized; and metamorphic core complexes are not present.

Proceed west on Highway 140. At mile 57.6, the highway crosses the Quinn River. To the north are the Bilk Creek Mountains, at the southern tip of which is a very small but classic locality for upper Paleozoic and lower Mesozoic strata of magmatic arc affinity (Quinn River Crossing; Ketner and Wardlaw, 1981).

71.8 Turn left onto the Big Creek Ranch road (gravel); this junction is marked by a sign for a "Photo Gallery" on the west side of the road and several large mailboxes on the east side.

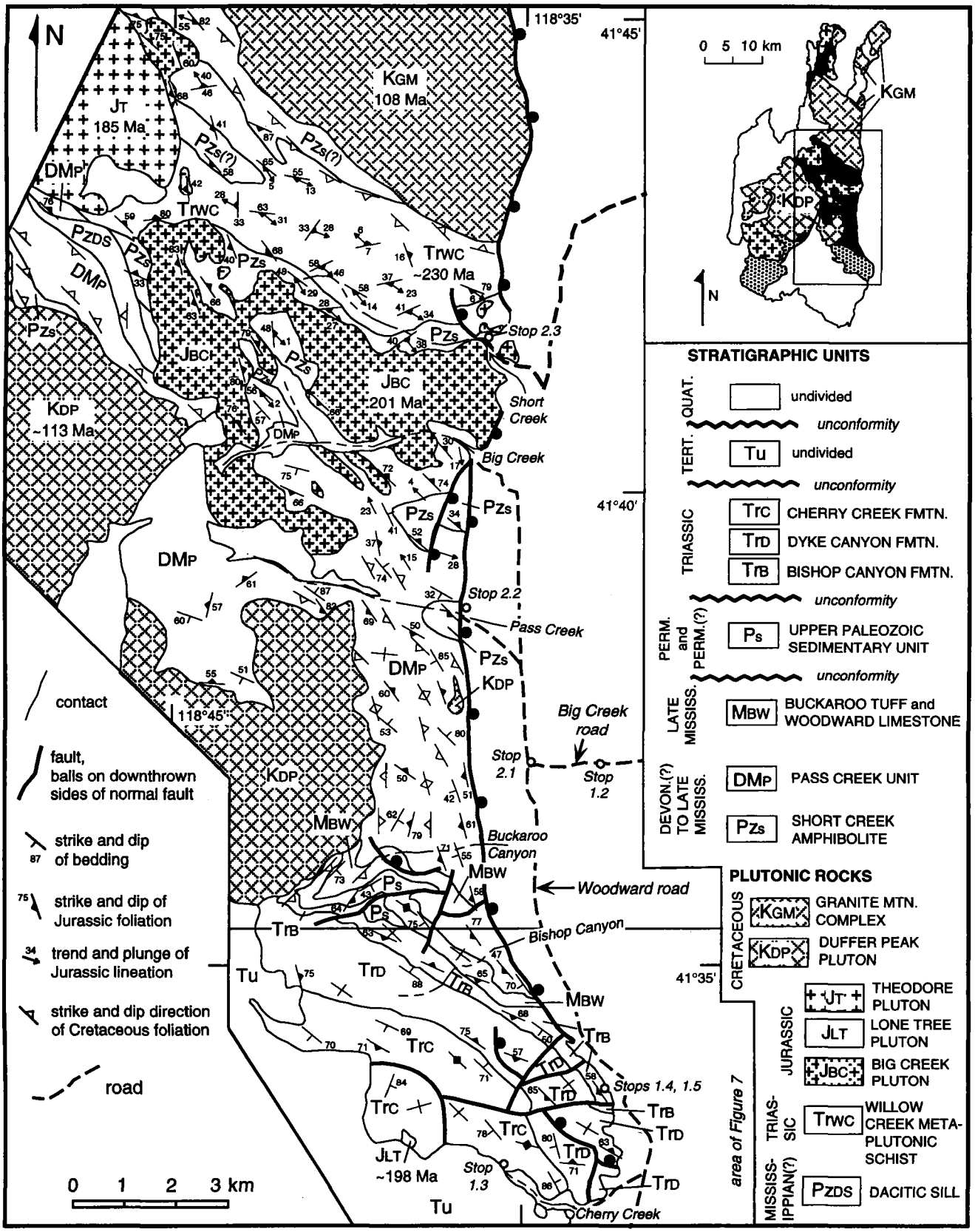
75.5 Stop at cattleguard.

Stop 1.2: Overview of Pine Forest Range

This stop is located on the Dyke Canyon 7.5 minute topographic map and shown in Figure 4.

Paleozoic and Mesozoic Stratigraphy

The most complete section of Paleozoic and Mesozoic strata in the Black Rock Desert is in the east-central Pine Forest Range (Fig. 4). Pre-Cenozoic strata in this area comprise a thick (~9 km) succession that ranges in age from


N
118°35'
41°45'
41°40'
41°35'
118°45'
KGM
108 Ma
JT
185 Ma
DMP
PZS(?)
TRWC
~230 Ma
PZS
JBC
201 Ma
KDP
~113 Ma
Stop 2.3
Short Creek
Big Creek
Stop 2.2
Pass Creek
Big Creek road
Stop 2.1
Stop 1.2
Buckaroo Canyon
MBW
Woodward road
Bishop Canyon
TRB
TRD
TRC
JLT
~198 Ma
Stop 1.3
Stops 1.4, 1.5
Cherry Creek
area of Figure 7
0 5 10 km
contact
fault, balls on downthrown sides of normal fault
strike and dip of bedding
strike and dip of Jurassic foliation
trend and plunge of Jurassic lineation
strike and dip direction of Cretaceous foliation
road
0 1 2 3 km
STRATIGRAPHIC UNITS
QUAT.
undivided
unconformity
TERT.
Tu
undivided
unconformity
TRIASSIC
TRC CHERRY CREEK FMTN.
TRD DYKE CANYON FMTN.
TRB BISHOP CANYON FMTN.
unconformity
PERM. and PERM.(?)
PS UPPER PALEOZOIC SEDIMENTARY UNIT
unconformity
LATE MISSISS.
MBW BUCKAROO TUFF and WOODWARD LIMESTONE
DEVON.(?) TO LATE MISSISS.
DMP PASS CREEK UNIT
PZS SHORT CREEK AMPHIBOLITE
PLUTONIC ROCKS
CRETACEOUS
KGM GRANITE MTN. COMPLEX
KDP DUFFER PEAK PLUTON
JURASSIC
JT THEODORE PLUTON
JLT LONE TREE PLUTON
JBC BIG CREEK PLUTON
TRIASSIC
TRWC WILLOW CREEK META-PLUTONIC SCHIST
MISSISSIPPIAN(?)
PZDS DACITIC SILL

Devonian(?) or Mississippian to latest Triassic (Fig. 5). In describing this stratigraphy, two features are important to emphasize. First, older units are progressively more deformed and metamorphosed than younger units (Wyld, 1996), and therefore older units are described as metamorphic rocks; the origin of this feature is explained below. Second, the section has been tilted southward and is now exposed in broadly cross-sectional view with units younging consistently toward the south (Fig. 4; Wyld, 1990). As will be explained below, this tilting most likely occurred in the Cretaceous and/or Cenozoic.

The Paleozoic and Triassic stratigraphy of the east-central Pine Forest Range has been described in detail in Wyld (1990, 1991a, 1991b) and is summarized here and in Figure 5. Paleozoic strata consist largely of sedimentary rocks, although two pulses of volcanism are represented by the mafic Short Creek amphibolite and the felsic Buckaroo tuff. Deep marine conditions of deposition characterize the older part of the section, but shallow marine conditions predominated in the Late Mississippian and Permian. Apparently all of the Pennsylvanian and possibly the early part of the Permian are missing across a low-angle unconformity (Fig. 5). Another low-angle unconformity separates Permian and Middle(?) to Late Triassic strata (Fig. 5). Triassic strata record subsidence of the arc to relatively deep marine depths and the onset of early Mesozoic arc magmatism (Wyld, 1990, 1991a). Older Triassic rocks (Ladinian or Carnian to early Norian) include abundant carbonates and siliciclastic strata (Bishop Canyon formation), and these were succeeded during the remainder of the Norian by basaltic to andesitic lavas and volcaniclastic strata (Dyke Canyon and Cherry Creek formations) which accumulated to a thickness of about 3 km (Fig. 5).

The ages of the upper Mississippian to uppermost Triassic units described above are well defined by either paleontologic or radiometric age data (Fig. 5; Wyld, 1990, 1991a). The ages of the two oldest units, the Pass Creek unit and the Short Creek amphibolite, are defined in part on less direct evidence, listed as follows. (1) Fossils from the top of the Pass Creek unit are Late Mississippian (Wyld, 1990). (2) Late Devonian (resedimented) radiolaria are present in chert clasts in strata of the Bilk Creek Mountains (Fig. 2) that are lithologically identical to the Pass Creek unit (Jones, 1990; Wyld, 1990). (3) In situ Late Devonian to Early Carboniferous radiolaria have been found in strata of the Jackson Mountains that are lithologically identical to the Pass Creek unit (Fig. 2; Russell, 1981; Quinn, 1996). Based on these three relations, the Pass Creek unit is interpreted to be of Late Devonian(?) to Late Mississippian age. (4) The Short Creek amphibolite is in gradational contact beneath the Pass Creek unit. (5) The Short Creek amphibolite is intruded by a dacitic dike (Mississippian[?] dacitic sill unit in Fig. 4) that is very similar in composition to the distinctive Buckaroo tuff which has yielded a ~327 Ma (Late Mississippian) U–Pb zircon age (Wyld, 1991a). Collectively, these relations indicate that the Short Creek amphibolite is no younger than Late Mississippian and suggest that it is probably Devonian.

Wyld (1990) suggested that intensely deformed quartzo-feldspathic and mafic schists located north of the Short Creek amphibolite were derived from arkosic sedimentary and mafic volcanogenic protoliths that were possibly broadly correlative with lower Paleozoic rocks in the Sierra Nevada and Klamath Mountains. These rocks, which were given the names Paleozoic(?) quartzo-feldspathic schist by Wyld (1990) and Willow Creek schist by Wyld (1996), have thus previously been interpreted to constitute the oldest exposed unit within the stratigraphic succession. Recent work has shown, however, that at least part of this map unit is a deformed and metamorphosed Middle Triassic pluton because a sample of typical quartzo-feldspathic schist yielded a morphologically homogeneous population of euhedral zircons, 11 of which gave Middle Triassic (~230 Ma) single-crystal U–Pb dates (B. Darby and G. Gehrels, personal communication, 1997). How much of this unit is plutonic remains uncertain at present; it appears likely, however, that a mixture of highly strained pluton and highly strained wall rock selvages are present. The Willow Creek schist is thus redefined here to include only those quartzo-feldspathic schists that are believed to be deformed plutonic rock, whereas areas underlain primarily by mafic schists are tentatively correlated with the Short Creek amphibolite (see map units labeled Pzs[?] in Fig. 4).

Figure 4. Geologic map of the east-central Pine Forest Range (outlined area in inset map) and locations of field trip stops 1.2–2.3. Inset map shows entire Pine Forest Range (see Fig. 3 for location of Pine Forest Range). Patterns in inset map as follows: dark shading is Paleozoic strata and Triassic plutons, light shading is Mesozoic strata, white is Tertiary strata; cross pattern is Jurassic plutons, diagonal line pattern is Cretaceous plutons. Ages of plutons are from U–Pb zircon analyses (see text).

Mesozoic Magmatism

Magmatic rocks of Triassic, Jurassic and Cretaceous age are widespread in the Pine Forest Range. Triassic magmatism in the Pine Forest Range apparently occurred in two pulses. The first is represented by plutons and stocks in the northern and east-central parts of the range that have been dated as Middle Triassic (~230–235 Ma; U–Pb zircon ages), including the Willow Creek schist (Wyld, 1991a; Wright and Wyld, unpublished data; B. Darby and G.

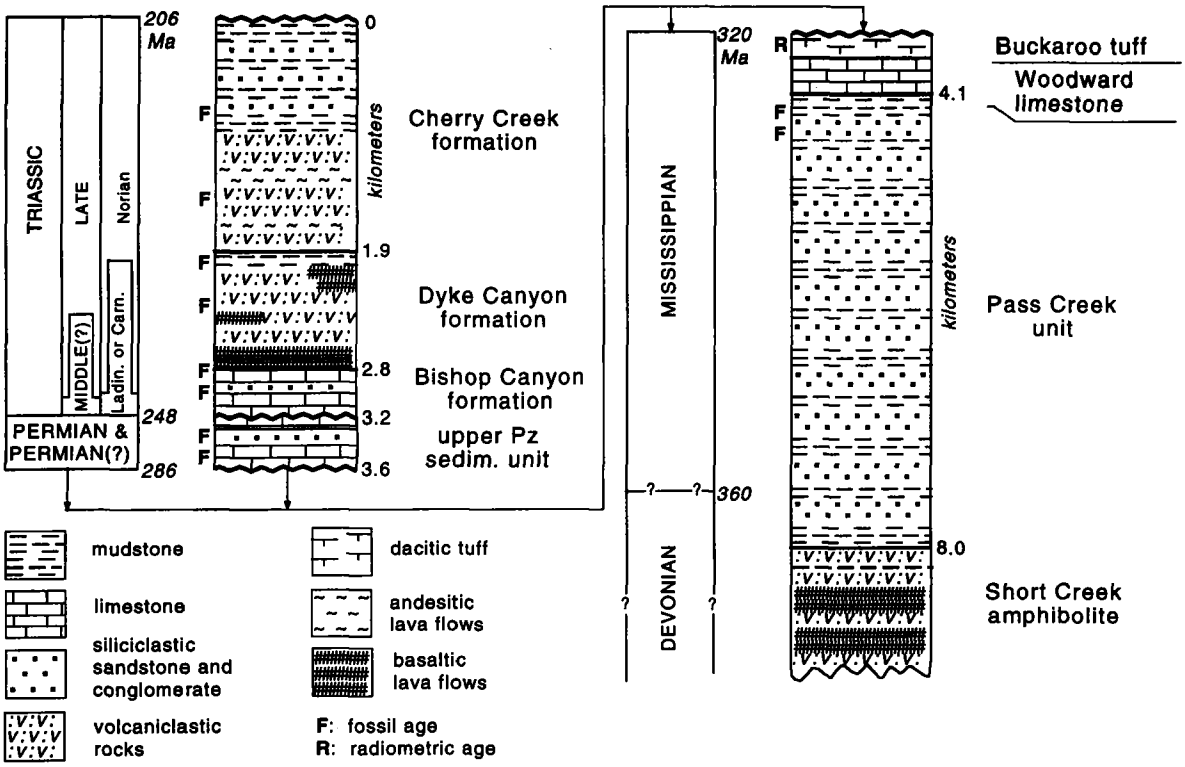


Figure 5. Paleozoic and Mesozoic stratigraphy of the east-central Pine Forest Range. All units are depicted in terms of protoliths regardless of degree of deformation or metamorphism. Thicknesses determined for different units are structural. Age data from Wyld (1990, 1991a).

Gehrels, personal communication, 1997). The second pulse of Triassic magmatism occurred in the latest Triassic (Norian) and is represented by the thick accumulations of volcanogenic Late Triassic strata (Fig. 5).

Widespread magmatism continued after this for the next ~15–25 m.y. into the Jurassic, and is represented primarily by plutonic rocks; no dated Jurassic volcanogenic strata are known, although volcanogenic rocks in the southwesternmost Pine Forest Range may be Jurassic based on stratigraphic relations (Wyld, 1991a). Three Jurassic plutons are present in the east-central Pine Forest Range (Fig. 4); the 201 Ma Big Creek pluton, the ~198 Ma Lone Tree pluton, only a small part of which protrudes from under a cover of Tertiary strata, and the 185 Ma Theodore pluton (U–Pb zircon ages; Wyld, 1996). Five other Jurassic plutons are present elsewhere in the Pine Forest Range. Of these we have preliminary U–Pb zircon data from four indicating crystallization of two at ~190 Ma and two between ~190–180 Ma (Wright, unpublished data). The fifth pluton is constrained by cross-cutting relations to be Early Jurassic, and pre-~190 Ma (Wyld, 1991a). Collectively, the data described above indicate that the continental arc in this area experienced a prolonged episode of magmatism from the latest Triassic to the Early and possibly early Middle Jurassic. Interestingly, however, there is no evidence for any late Middle or Late Jurassic magmatism in the Pine Forest Range.

Cretaceous magmatic rocks include two large plutonic bodies, the Duffer Peak pluton and the Granite Mountain plutonic complex (Fig. 4). These two plutonic bodies were emplaced in the mid-Cretaceous between ~114–108 Ma (preliminary U–Pb zircon data; Wyld, 1991a; Wright and Wyld, unpublished data).

Mesozoic Deformation

Two principal episodes of Mesozoic deformation and metamorphism have affected the east-central Pine Forest Range (Wyld, 1991a, 1996). The older episode was regional in extent and occurred during the Jurassic; it is discussed in detail below. The younger episode was more localized around the Lower Cretaceous Duffer Peak pluton and Granite Mountain complex (Fig. 4); it is described in more detail in the text for stop 2.1. This Cretaceous deformation is interpreted to primarily reflect strain related to pluton emplacement (Wyld, 1991a). The only other struc-

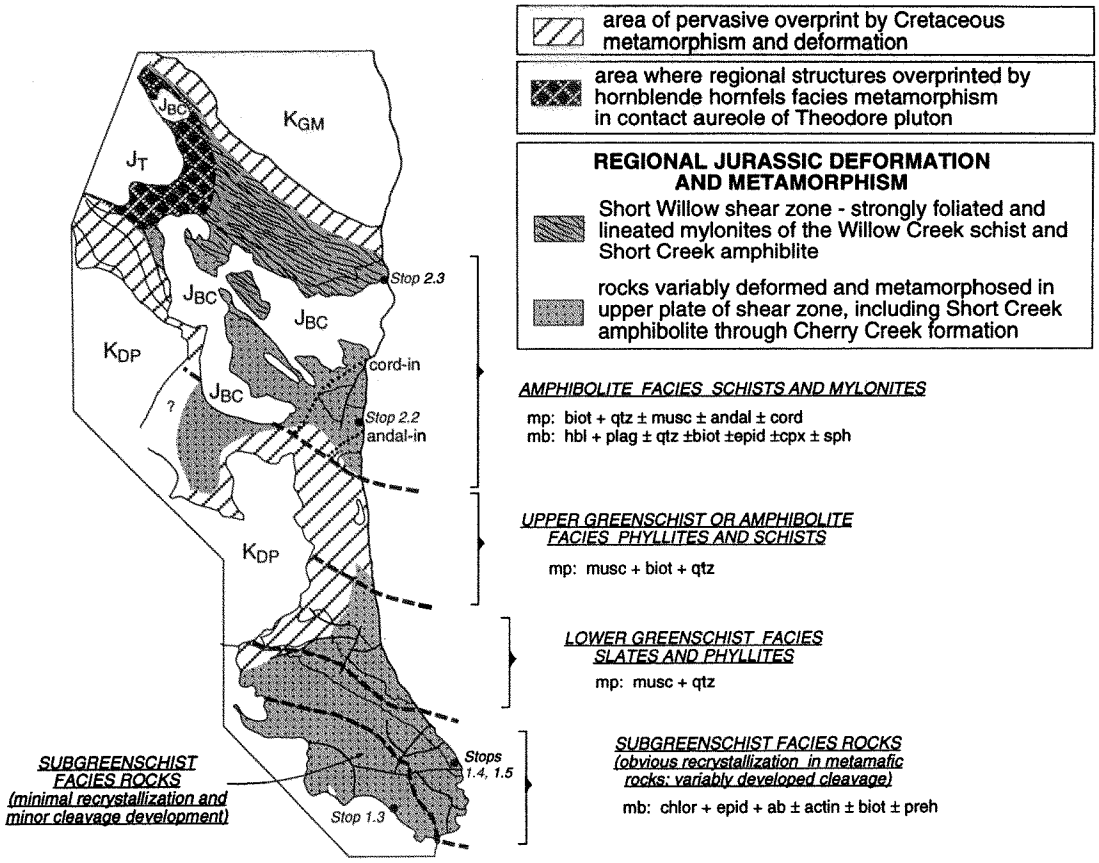


Figure 6. Simplified map of the east-central Pine Forest Range (same area as Figure 4) showing variation in regional Jurassic structures and metamorphic grade with stratigraphic depth, location of the Jurassic Short Willow shear zone, and locations of field trip stops 1.3–1.5 and 2.2–2.3. To aid in comparing locations between this figure and Figure 4, the outlines of Cretaceous and Jurassic plutons are shown (symbols same as in Figure 4), as are the contacts between the Paleozoic and Triassic units and the major faults (see thin lines running through the shaded and patterned areas). Abbreviations "mp" and "mb" refer to metamorphic assemblages developed in metapelitic and metabasaltic protoliths, respectively. Heavy dashed lines indicate boundaries between different metamorphic facies. Heavy dotted lines indicate mineral-in isograds. Other abbreviations are: ab, albite; actin, actinolite; andal, andalusite; biot, biotite; chlor, chlorite; cord, cordierite; cpx, clinopyroxene; epid, epidote; hbl, hornblende; musc, muscovite; plag, plagioclase; preh, prehnite; qtz, quartz; sph, sphene.

tures of significance in the east-central Pine Forest Range are high-angle brittle faults (Fig. 4), most of which are demonstrably related to Cenozoic extension, and all of which postdate Jurassic deformation (Wyld, 1991a). The general distribution and character of Mesozoic deformation and metamorphism in the east-central Pine Forest Range is shown in Figure 6, which shows the same area as Figure 4 but emphasizes key features of Mesozoic deformation and metamorphism. Regional Jurassic deformation produced structures whose character and metamorphic grade vary spatially in the east-central Pine Forest Range (Fig. 6). Important aspects of the Jurassic deformation in this area are as follows (summarized from Wyld, 1996).

1. Regional structures affect a coherent stratigraphic succession that is unbroken by any faults formed during or prior to regional deformation. This stratigraphic succession strikes NW-SE and is tilted so that strata young from northeast to southwest (Fig. 4).
2. Regional deformation produced structures and metamorphic assemblages that vary progressively with stratigraphic depth (Figs. 4, 6). At the highest stratigraphic levels (youngest Triassic strata), the principal structure is a pressure-solution cleavage developed only in rheologically incompetent rock types and accompanied by minimal metamorphic recrystallization. At deeper stratigraphic levels, within the Triassic to upper Paleozoic part of the section, the cleavage is more pronounced and pervasive and is accompanied by more complete metamorphic recrystallization to subgreenschist or lower greenschist facies assemblages. Proceeding downsection, regional structures

progressively change into a greenschist to amphibolite grade schistosity, and at the deepest exposed levels, where Devonian(?) strata are intruded by the Willow Creek meta-plutonic schist, the rocks are amphibolite grade mylonites with a pronounced foliation and stretching lineation. This zone of mylo- nites is called the Short Willow shear zone (Fig. 6). Nowhere within the sequence is there any break or abrupt change in the metamorphic and deformational gradient.

3. The regional foliation, where not significantly affected by Cretaceous deformation, is NW-striking and moderately to steeply dipping (Fig. 4). The stretching lineation present in the northern part of the area trends northwest or southeast and has a subhorizontal plunge.
4. There is only limited evidence for folding of stratigraphic layering during regional Jurassic deformation, and the regional foliation is typically at a low angle to bedding (Fig. 4). This is consistent with an interpretation that deformation occurred by broadly layer-parallel shear, probably coupled with layer-perpendicular shortening. It is not consistent with a model of deformation by layer-parallel shortening. The few folds observed are asymmetric and indicate that young rocks moved to the northwest relative to older rocks.
5. Shear sense indicators within higher strain schists and mylonites (described in detail in Wyld [1996]) also indicate that young rocks moved to the northwest relative to older rocks. Because the stretching lineation within the shear zone is subhorizontal, this appears to imply dextral strike-slip sense of motion; however, as explained below, this conclusion is misleading because it does not take into account tilting of the stratigraphic section.

The relations summarized above are interpreted as follows (Wyld, 1996). First, regional deformation in the east-central Pine Forest Range occurred when the Paleozoic-Mesozoic stratigraphic succession was still upright to account for the consistent and progressive increase in metamorphic grade with stratigraphic depth. Thus, tilting of the stratigraphic succession occurred after regional deformation. Second, regional deformation was related to development of the Short Willow shear zone, with the progressively lower strain rocks to the south representing a strain gradient above the shear zone. Because tilting of the stratigraphic section occurred after regional deformation, we can conclude that the shear zone developed originally as a low-angle (subhorizontal) ductile fault at the deepest exposed stratigraphic levels (~8–10 km depth) and that the lower strain rocks to the south are the upper plate of the shear zone, with both the shear zone and its upper plate now tilted to the south.

Subhorizontal ductile shear zones can develop during either regional shortening or regional extension and it is important to address which structural model is appropriate in the Pine Forest Range. The simplest way to approach this question would be by examining the lower plate of the shear zone. Unfortunately, this is not exposed because the shear zone is cut out to the north by the Granite Mountain plutonic complex (Figs. 4, 6). Preliminary work farther north in the Pine Forest Range, however, indicates the presence of metasedimentary rocks, most likely correlative with the Pass Creek unit, which were deformed and metamorphosed to amphibolite grade during the Late Triassic or Early Jurassic and which are a good candidate for lower plate rocks of the Short Willow shear zone (Wyld et al., 1996). If correct, this argues for older-on-younger relations and a thrust interpretation. The structural character of the upper plate of the Short Willow shear zone is also relevant. Numerous studies in areas where low-angle extensional shear zones are exposed indicate that their upper plate (1) is imbricated by brittle normal faults resulting in a structurally attenuated stratigraphic section that is tilted with respect to the ductile shear zone and (2) contains rocks that are at a notably lower metamorphic grade than rocks within the shear zone (e.g., Gans and Miller, 1983; Davis and Lister, 1988; McGrew, 1993). None of these features are present in the upper plate of the Short Willow shear zone. Instead, a coherent stratigraphic section in which metamorphic grade and strain increase steadily downsection is observed, and this is very similar to relations observed in the upper plates of ductile thrust faults (Ramsay, 1981; Sanderson, 1982; Miller et al., 1988; Smith et al., 1993). Finally, there is no evidence for Early Jurassic extensional tectonism elsewhere in the Black Rock Desert; however, there is a record of Early Jurassic thrust faulting in the Jackson Mountains (Fig. 3; Quinn and Wright, 1993; and see later section). The Short Willow shear zone is therefore interpreted to be a ductile thrust fault.

Structural relations summarized above indicate that, in present day coordinates, young rocks moved to the northwest relative to older rocks during shear zone development. The original sense of shear is more difficult to evaluate due to uncertainties about exactly how tilting of the stratigraphic section in the east-central Pine Forest Range was accomplished after shear zone development. Tilting is believed to be related in part to forcible intrusion of the Cretaceous plutons, particularly the massive Granite Mountain complex which extends from the central Pine Forest Range northeast into the Bilk Creek Mountains (Figs. 3, 4), but was probably also influenced by differential uplift during Cenozoic Basin and Range extension (Wyld, 1991a, 1996). Because tilting was likely accomplished as a multi-

stage process, and because the axes of tilting events are uncertain, it is not possible to precisely determine pre-tilting orientations of structures. One conclusion that can be made with some degree of certainty, however, is that no reasonable tilting history would reconstruct shear sense back to any direction other than top-to-the-west, top-to-the-northwest, or top-to-the-north. Original sense of shear is thus interpreted to be broadly top-to-the-northwest.

Regional deformation in the east-central Pine Forest Range occurred in the Early Jurassic based on the fact that rocks as young as latest Norian are involved in the deformation, evidence that the 201 Ma Big Creek pluton was intruded during deformation, as described below, and the presence of a metamorphic aureole around the 185 Ma Theodore pluton which overprints the regional fabric (Fig. 6; Wyld, 1996). Further evidence for an Early Jurassic age of regional deformation in the Pine Forest Range comes from the northernmost part of the range where regional structures deform rocks as young as ~230 Ma and are cross-cut by an ~180 Ma pluton (preliminary U–Pb zircon; Wright and Wyld, unpublished data). There is no evidence for any younger Jurassic deformation in the Pine Forest Range.

Proceed west along Big Creek road (Fig. 4).

78.6 Junction with Woodward Road (Fig. 4). Turn left and proceed south along the range front.

83.9 Junction with Cherry Creek road (unmarked dirt and gravel road). This road is shown on Figures 4 and 7. Turn west up; the road appears deeply rutted near the junction but improves in a short distance and is easily driveable. Remain on main road going west.

86.2 Junction with gravel mine road that goes uphill back to the east (see road junction at stop 1.3 position on Fig. 7b). Park on mine road near the junction.

Stop 1.3: Younger, lower strain Triassic strata

This stop is located on the Dyke Canyon 7.5 minute topographic map and is shown on Figures 4 and 7. The traverse at this stop (shown in detail on Fig. 7b) will take us northeast up the mine road about 800 m. Along the way, we will look at several units of the Cherry Creek formation, the youngest Triassic unit mapped in the east-central Pine Forest Range, focusing on the magmatic, stratigraphic and structural record.

The Cherry Creek formation has been divided into four members (Figs. 7, 8), the lower three of which (Trc1–3) are dominated by coarse clastic rocks rich in andesitic volcanic detritus but also contain some interlayered andesite lava flows, and the uppermost of which (Trc4) consists primarily of fine grained clastic strata rich in volcanic detritus. The late Norian age of the Cherry Creek formation is well defined by fossils (Fig. 8). Facies analysis indicates that this unit accumulated in a marine environment near the base of the slope, with members Trc1–3 reflecting nearby active volcanism and member Trc4 reflecting either distal volcanism or erosion of the volcanic carapace (Wyld, 1991a). On this traverse, we will see distal turbidites of member Trc4, lavas of member Trc3, and spectacular outcrops of volcaniclastic debris flows of member Trc2. This 1.9 km thick succession dominated by volcanogenic material (Fig. 8) is a manifestation of the flare-up of Late Triassic to Early Jurassic magmatism in the Black Rock Desert segment of the continental arc.

Within these youngest Triassic rocks there is very little evidence of internal strain or metamorphism (see Fig. 6 and text adjacent to stratigraphic column in Fig. 8); these relations can be seen in outcrops along the road. Sedimentary textures and structures are well preserved and there is only minor metamorphic recrystallization, even within volcanic material. In general, internal strain in these rocks is evident only within the lower two members of the formation, where it is manifested primarily by a low-grade cleavage visible only in shales or clastic rocks with a carbonate matrix. Low strain, low-grade conditions recorded by the upper Cherry Creek formation cannot be taken as evidence that these rocks did not suffer the same deformational history as older, higher strain, higher-grade rocks to the north because gradational contacts exist between all members of the Cherry Creek formation and between the Cherry Creek formation and the underlying Dyke Canyon formation. Instead, the Cherry Creek formation simply experienced less strain and lower P/T conditions than older rocks; this is interpreted to reflect location of the Cherry Creek formation at structurally high levels during regional metamorphism (Wyld, 1996).

Return to vehicles and retrace route back to Woodward Road.

88.4 Junction with Woodward Road. Turn north.

89.4 Junction with dirt and gravel road leading into Solo Canyon across the southern part of the Solo Canyon alluvial fan (see Figs. 7a and 7b for location). Turn west. Note that Solo Canyon is a local, not a formal, name for this drainage; its location is shown in Figure 7, but the canyon is not named on the Dyke Canyon 7.5 minute topographic map of this area.

89.8 Junction with road leading back down the northern part of the Solo Canyon alluvial fan (see Fig. 7). Pass this road and proceed west up Solo Canyon.

90.7 Small parking and turnaround spot by the side of the road.

Stop 1.4: Older, higher strain Triassic strata

This stop is located on the Dyke Canyon 7.5 minute topographic map and is shown on Figures 7a and 7b. At

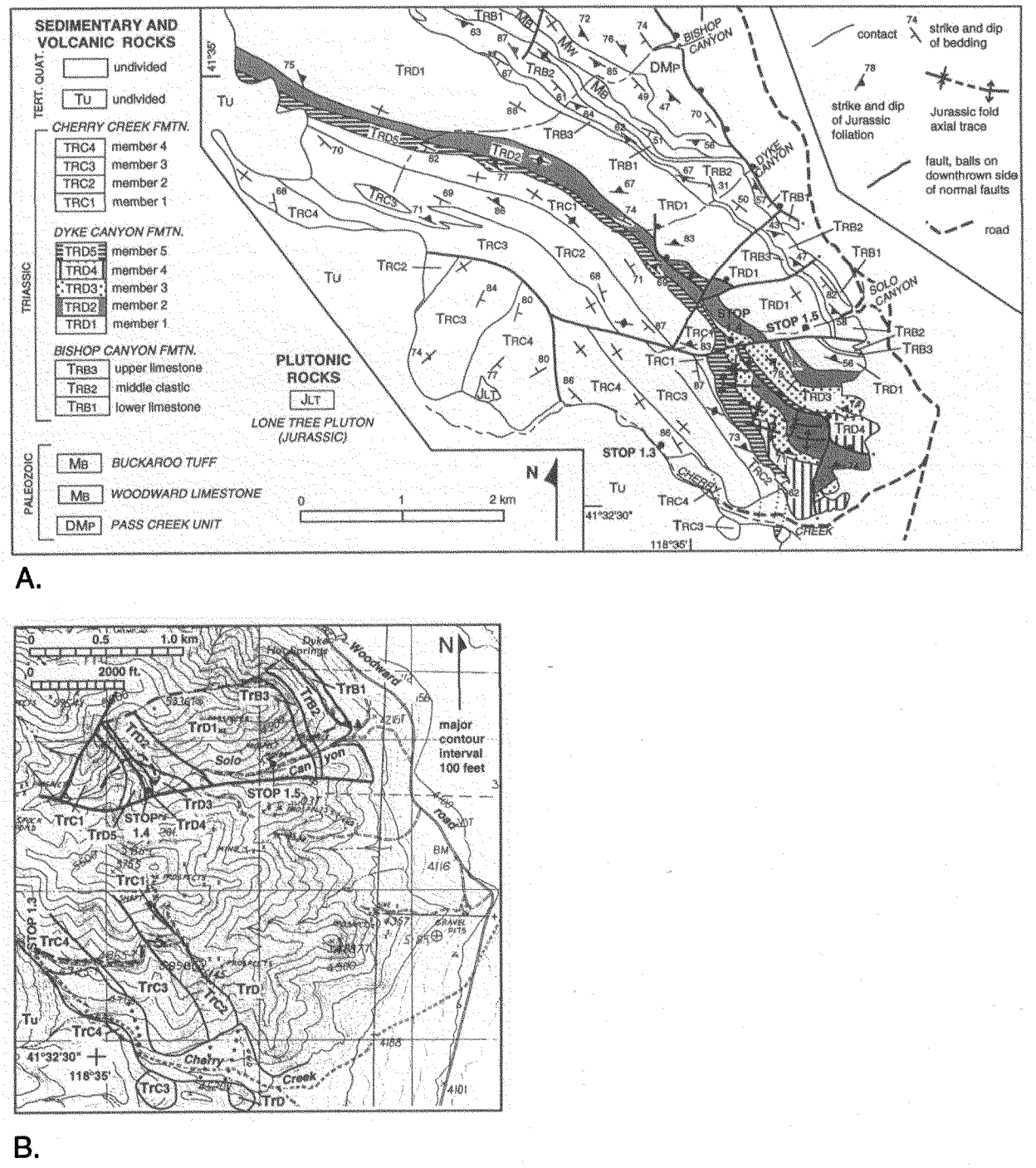


Figure 7. A. Detailed geologic map of the southeastern Pine Forest Range (area surrounded by box in Figure 4). B. Locations of stops 1.3, 1.4, 1.5 and respective traverses on topographic base; geologic contacts in vicinity of stops are shown. Symbols same as in A.

this and the next stop, we will walk generally eastward (see traverses shown on Fig. 7b) downsection through the Upper Triassic volcanic succession into older Triassic carbonates and clastics. Along the traverse we will see more evidence for Late Triassic volcanism. In addition, we will see a downsection increase in metamorphic grade and strain, although the amount of strain recorded by specific rocks or rock units is highly dependent on lithology.

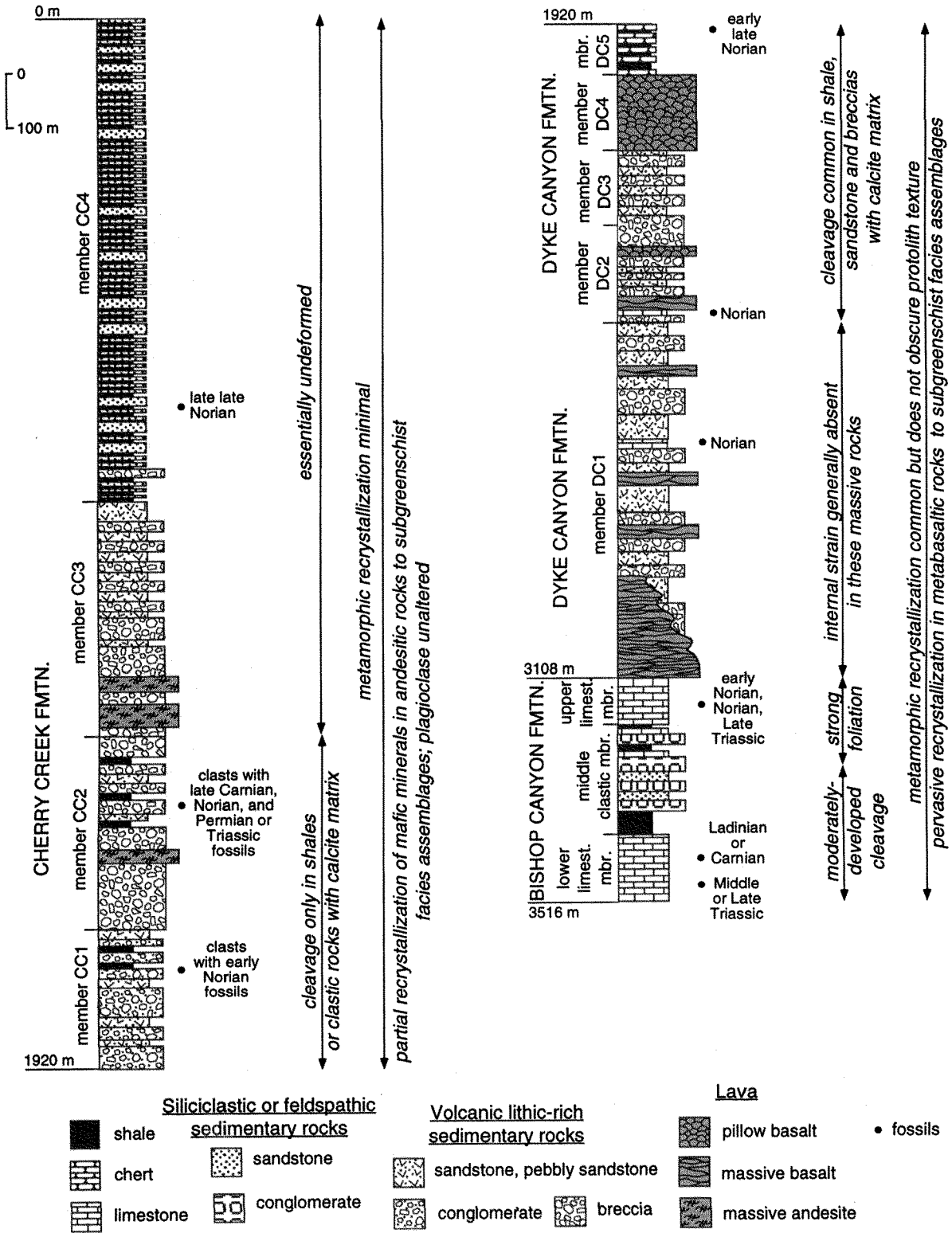


Figure 8. Composite stratigraphic section of Triassic rocks in the southeastern Pine Forest Range showing approximate stratigraphic level of age-diagnostic fossils and showing general structural and metamorphic characteristics at varying stratigraphic levels. Unit thicknesses are averages.

Begin by first hiking about 300 m west up the road to the lower member of the Cherry Creek formation (unit Trc1; see Fig. 7b). This unit consists mostly of turbidite-facies conglomerate and sandstone, interspersed with spectacular debris flow deposits (Fig. 8). Clasts were derived from an active andesitic volcanic center, a source of incompletely-lithified carbonate, and the underlying Dyke Canyon formation (Wyld, 1991a). Strain is minimal and heterogeneously-distributed in this unit: the coarsest-grained strata (e.g., the debris flow deposits) have no visible cleavage, whereas finer-grained rock types locally display a slaty cleavage. Metamorphic grade is very low in this unit; there is only limited metamorphic recrystallization of protolith mineralogy and mafic crystals are commonly unaltered.

Proceed eastward into the upper part of the Dyke Canyon formation, looking at outcrops above the road on the north side (Fig. 7b). The Norian Dyke Canyon formation has been divided into 5 members (Figs. 7, 8), four of which we will see at this stop. These include member Trd5, which consists mostly of interbedded chert and argillite; member Trd4, composed of basaltic pillow lava and pillow breccia; member Trd3, characterized by turbidite-facies, tuffaceous volcanic sandstone and breccia; and member Trd2, which contains a heterogeneous mixture of basaltic volcanic breccia and sandstone, massive to pillowed basaltic lava, and rare limestone. Member Trd1, described in more detail below, consists mostly of basaltic lava and volcanic breccia. Facies data from throughout this ~1200 m thick formation indicate deposition in a relatively deep marine environment, with proximal active volcanism recorded by the lower four members (Wyld, 1991a).

Relations to observe as you hike through the upper four members of the Dyke Canyon formation, ultimately dropping back downhill to the vehicles (Fig. 7b), are as follows. First, there is an abundance of proximal volcanogenic strata in this part of the section. Second, cleavage is evident in most rock types, although this cleavage varies in character from spaced in chert-argillite rocks to more penetrative in tuffaceous sandstones, and it is only locally evident in the coarser breccias and lavas (Fig. 8). This more pervasive cleavage development is a distinct difference between the Dyke Canyon and Cherry Creek formations and reflects a gradual increase in strain downsection. Second, basaltic rocks that make up most of the Dyke Canyon formation have a distinct green color, reflecting the widespread presence of metamorphic chlorite + epidote ± actinolite (Figs. 6, 8). This well-developed subgreenschist facies assemblage contrasts with the minor degree of metamorphic recrystallization in the Cherry Creek formation and reflects increasing metamorphism downsection.

Return to vehicles, turn around and drive east 0.2 miles.
90.9 Park in small grassy area.

Stop 1.5: Older, higher strain Triassic strata, continued

This stop is located on the Dyke Canyon 7.5 minute topographic map and is shown on Figure 7. We will start out in the lower member of the Dyke Canyon formation, looking at outcrops reached by hiking 100–200 feet up the ravine north of the parking area, and we will then sidehill east through the Bishop Canyon formation, dropping back down to the road near the range front (see traverse located on Fig. 7b).

The lower member of the Dyke Canyon formation here (Trd1) is composed entirely of massive basaltic lava with prominent large pyroxene phenocrysts (augite porphyry), and coarse volcanic breccia of similar composition (Fig. 8). The underlying Bishop Canyon formation, ~200–400 m thick, has been divided into three members (Figs. 7, 8; Wyld, 1990); an upper, generally fine-grained limestone member that contains early Norian conodonts (Trb3); a middle clastic member (Trb2) that consists of siliciclastic conglomerate, sandstone and shale; and a lower, generally fine-grained limestone member that contains Ladinian or Carnian conodonts (Trb1). There is no primary volcanic detritus in the Bishop Canyon formation, and facies relations (turbidite facies siliciclastic rocks and turbidite to pelagic/hemipelagic facies carbonates) indicate deposition in a relatively deep marine environment near the base of the slope (Wyld, 1990, 1991a).

Our main purpose on this traverse is to look at structural relations. We start out in the lavas and breccias of the lower Dyke Canyon formation which show no sign of any internal strain, although they do contain a subgreenschist facies metamorphic assemblage (Fig. 6) that imparts a pronounced green color to the rocks. Lack of any internal strain is characteristic of the bulk of unit Trd1 (Fig. 8) and is interpreted to be related to the fact that this unit is composed almost entirely of massive lava flows and coarse massive breccias that are locally tens of meters thick. These rock types would be rheologically strong and difficult to deform under the relatively low-strain conditions that affected Triassic rocks in the east-central Pine Forest Range. In contrast to the lower part of the Dyke Canyon formation, rocks of the Bishop Canyon formation exhibit a well-developed cleavage that is evident in all rock types. This is interpreted to reflect the fact that the Bishop Canyon formation is composed of rock types that were rheologically weaker, such as comparatively thin-bedded limestone and siliciclastic strata. Thus, during deformation, strain in this part of the stratigraphic section was accommodated primarily in the incompetent Bishop Canyon formation.

On a more detailed level, there is a local increase in strain within the Bishop Canyon formation upsection from the base of the formation to the contact with the unde-

formed basal member of the Dyke Canyon formation (see text adjacent to stratigraphic column in Fig. 8); this increase in strain is manifested primarily by an increasingly intense foliation upsection toward the Dyke Canyon formation. This pattern is interpreted to reflect differential strain during layer-parallel shear, with the highest strain occurring in incompetent rocks (Bishop Canyon formation) near their boundary with strong rocks (lower Dyke Canyon formation).

In summary, the relations we have seen in stops 1.3–1.5 indicate the following. First, there was a prolonged episode of basaltic to andesitic marine volcanism in this area in the latest Triassic (Norian). Second, the extent of regional metamorphic recrystallization increases downsection within the Triassic strata of the east-central Pine Forest Range. Finally, Jurassic strain recorded in these strata is heterogeneous and is controlled both by rheology and by stratigraphic depth, with deformation most pronounced in finer-grained sedimentary rocks and in rocks that occupy the deeper stratigraphic levels.

Return to vehicles and drive east along Solo Canyon road.

91.5 Intersection with two roads leading down the alluvial fan at the mouth of Solo Canyon (Fig. 7). Take the northern road.
91.9 Intersection with Woodward road. Turn north.
95.8 Intersection with Big Creek road. Turn right (east).
102.6 Junction with Highway 140. Turn left (north).
127.5 Denio Junction.

DAY 2: CENTRAL PINE FOREST RANGE—JURASSIC PLUTONISM AND DEEPER LEVEL JURASSIC DEFORMATION

0.0 Denio Junction. Proceed south on Highway 140.
24.8 Intersection with Big Creek road. Turn west (right).
31.6 Junction with Woodward road.

STOP 2.1. Overview

This stop is located on the Dyke Canyon 7.5 minute topographic map and is shown on Figure 4. It is an overview stop to provide information relevant for connecting relations seen yesterday with stop 2.2 today.

Paleozoic rocks that form the stratigraphic basement for the Triassic section we saw yesterday underlie a much larger area in the east-central Pine Forest Range than the Triassic rocks (Fig. 4). From the vantage point of this stop, Paleozoic strata underlie most of the area from ~6 km to the north (to Big Creek), ~5 km to the south (to south of Bishop Canyon), and ~3 km to the west (Fig. 4). Most of this area is underlain by the Pass Creek unit (Fig. 4). This distinctive unit consists of a monotonous assemblage of metamorphosed shales with less common sandstone and minor conglomerate. The coarser clastic rocks are commonly graded and can be classified as turbidite facies where not extensively metamorphosed and recrystallized; thus deposition is inferred to have occurred in a relatively deep marine environment, probably a deep sea fan. Clastic detritus in coarser rocks includes abundant mono-crystalline quartz, polycrystalline quartz, chert and argillite (Wyld, 1990). Calculated structural thickness of the unit is ~4 km (Fig. 5), but this can only be considered approximate. As noted earlier, age constraints for the Pass Creek unit indicate deposition in the Late Devonian(?) to Late Mississippian.

As discussed by Wyld (1990), the Pass Creek unit appears very similar in composition, facies, thickness, and age range to the Bragdon Formation of the eastern Klamath Mountains (Miller and Harwood, 1990). This suggests that the two units may have been deposited within the same basin and argues for relatively close proximity between the Black Rock Desert and Klamath Mountains provinces in the mid-Paleozoic. In contrast, the Jurassic evolution of these provinces is very different, as discussed in the introduction and further below, which suggests that the two provinces were more distant from one another in the Jurassic (Wyld et al., 1996). We hypothesize that the eastern Klamath Mountains and Black Rock Desert provinces were located very near to one another in the mid-Paleozoic, but that they were displaced from one another sometime between the mid-Paleozoic and the Jurassic. Insofar as both provinces record intermittent arc volcanism during this time interval (Miller and Harwood, 1990; Wyld, 1990), we suggest that the most likely explanation for displacement is broadly arc-parallel strike-slip faulting. One possible interpretation is that the eastern Klamath Mountains province was displaced to the south relative to the Black Rock Desert province during the Pennsylvanian left-lateral strike-slip faulting episode which resulted in truncation of the continental margin farther south (Walker, 1988). This is an appealing sense of displacement because it positions the eastern Klamath Mountains province correctly to be moved northward again along the right-lateral Mojave-Snow Lake strike-slip fault (Schweickert and Lahren, 1990, 1993) in the Cretaceous into its current position west of the Black Rock Desert (Figs. 1, 2), as hypothesized by Wyld et al. (1996).

From the Triassic units north into mid-Paleozoic strata, the pattern of downsection increasing Jurassic strain and synmetamorphic grade that we saw yesterday persists, but is obscured to some extent by metamorphism and deformation around the Cretaceous Duffer Peak pluton (Figs. 4, 6). The Duffer Peak pluton underlies the high country that can be seen to the west from this stop. It is surrounded by an aureole of metamorphism and deformation related to intrusion (Fig. 6; Wyld, 1991a). This aureole is particularly wide on its east side near the range front where the pluton margin dips outward (east) at a moderate angle; in con-

trast, the pluton margin is more steeply dipping elsewhere and the metamorphic and deformational aureole is correspondingly more narrow (Fig. 6). Within the wide eastern aureole zone, Jurassic structures (described further below) are refolded by tight to isoclinal Cretaceous folds, and overprinted by a generally pronounced Cretaceous foliation and by upper greenschist to amphibolite-facies Cretaceous metamorphism (Wyld, 1991a). Farther from the pluton margin, Cretaceous strain and metamorphic grade diminish and eventually die out (Wyld, 1991a).

Because the Cretaceous overprint features are prominent from the vicinity of Buckaroo Canyon to Pass Creek (Figs. 4, 6), and because patterns of Jurassic deformation and metamorphism can only be discerned in this area by detailed microscopic studies coupled with macroscopic observations at multiple localities, it is impractical to take traverses through this region to observe the effects of Jurassic deformation and metamorphism. Key relations within the area of Cretaceous overprint are, however, described below. Because most of this area is underlain by the Pass Creek unit, we focus on relations in this unit.

Detailed outcrop-scale and microtextural studies within metapelites of the Pass Creek unit between its upper and lower contacts indicate that the Jurassic regional fabric in this unit varies progressively from a low-grade slaty cleavage in the younger (southern) part of the unit, to a phyllitic foliation in the middle part of the unit, to a fine-grained schistose foliation defined by visible mica in the older (northern) part of the unit (Wyld, 1996). Accompanying this downsection change in the character of the regional foliation is the appearance of a stretching lineation, defined by pebbles in conglomerates which vary from unstrained at high stratigraphic levels to highly elongate at deep stratigraphic levels. Metamorphic grade also increases with stratigraphic depth from subgreenschist or lower greenschist-grade at the top of the unit to amphibolite-grade at the base of the unit where it is in gradational contact with the Short Creek amphibolite (Figs. 4, 6; Wyld, 1996). It is important to emphasize that these changes in the character of structures and syntectonic metamorphism downsection within the Pass Creek unit are gradual and are not associated with any structural break. Our next field trip stop (see below) will be in the more highly strained and metamorphosed lower Pass Creek unit. For those future users of this guide who may be interested in seeing the relatively low-grade and low-strain rocks of the upper Pass Creek unit, good exposures can be found locally in the area north of Bishop Canyon (Fig. 4).

Turn right and proceed 1.3 miles north on the Woodward road.

32.9 Intersection with dirt road heading west toward Pass Creek (shown on Fig. 4); cattle fence gate at entrance of road. Turn west (left). Be sure to close the cattle gate.

33.9 Mouth of Pass Creek. Park.

STOP 2.2. Deformation in mid-Paleozoic units; syntectonic metamorphism around Big Creek pluton

This stop is located on the Howard Hot Spring 7.5 minute topographic map and is shown on Figures 4 and 9a. There are two main goals to this stop. One is to look at the character and metamorphic grade of regional Jurassic structures in rocks at this stratigraphic level and compare this with what was seen in younger rocks yesterday. The Pass Creek metasedimentary rocks we will look at here were deformed under amphibolite grade conditions; this is based on the metamorphic assemblage in metabasaltic rocks of the adjacent Short Creek amphibolite (Figs. 4, 6) which is in gradational contact with the Pass Creek unit. The Pass Creek rocks at this stop are also more highly strained than the rocks we saw yesterday.

Another goal is to observe metamorphic and structural evidence that the nearby 201 Ma Big Creek pluton is syntectonic with respect to the regional deformation. The lower part of the Pass Creek unit and the Short Creek amphibolite are intruded by the Big Creek pluton (Fig. 4). Detailed studies within metapelites of the lower Pass Creek unit indicate that the regional (syntectonic) metamorphic assemblage in these rocks varies with distance from the Big Creek pluton (Fig. 6; Wyld, 1996). Farther from the pluton, biotite always occurs with muscovite and there are no aluminosilicates. Closer to the pluton, however, first andalusite and then cordierite porphyroblasts appear as part of the regional metamorphic assemblage (Fig. 6) and biotite locally occurs in the absence of muscovite. These relations suggest that the regional metamorphic assemblage is upgraded toward the Big Creek pluton margin. In addition, aluminosilicate porphyroblasts exhibit syntectonic textural relations with respect to the regional fabric. These textural relations can be summarized as follows: the regional foliation (defined largely by aligned mica) can be traced into the ends of the andalusite and cordierite porphyroblasts but also bends weakly around them; there are no pressure shadows at the ends of the porphyroblasts; the long axes of the andalusite porphyroblasts are parallel to the regional foliation and define a stretching lineation (Wyld, 1991a, 1996). Collectively, these relations provide evidence that regional deformation and metamorphism occurred during emplacement of the Big Creek pluton at 201 Ma. This Early Jurassic age for regional deformation is further corroborated by $^{40}Ar/^{39}Ar$ age data from nearby samples of the Short Creek amphibolite; syntectonic hornblende from these samples yielded $^{40}Ar/^{39}Ar$ plateau ages of ~193–205

Ma (Wyld and P. Copeland, unpublished data). We will also see further structural evidence for syntectonic intrusion of the Big Creek pluton at stop 2.3.

Hike north along the range front ~800 m to the first major drainage (Fig. 9a). The traverse proceeds west up the southern ravine of the drainage to about the 4800 feet elevation, then sidehills to the north around to the northern ravine of this drainage, then sidehills farther north to a point where one can look northward down into the next big drainage at the Short Creek amphibolite. Outcrops along this traverse are in the lower part of the Pass Creek unit and show the following features. (1) Schistose metapelites with a foliation defined by fine grained biotite and muscovite. Compare this with the low-grade slaty cleavage seen yesterday. (2) Highly strained and lineated chert-pebble conglomerates. Compare this with the generally unstrained pebbles in coarse clastic rocks seen yesterday. (3) Andalusite porphyroblasts in metapelites in the aureole of the Big Creek pluton whose long axes lie within the plane of the regional foliation and define a stretching lineation.

Return to vehicles and drive back east on the Pass Creek road.

34.9 Intersection with Woodward road. Turn south (right), closing the cattle gate.

36.2 Intersection with Big Creek road. Turn east (left).

43.0 Intersection with Hwy. 140. Turn north (left).

51.0 Intersection with dirt road leading west toward the range front. This road initially goes off from the highway at an acute angle to the southwest and is difficult to see coming from the south along the highway. A helpful hint is to look for the small hill that it angles to the top of. Turn onto this road and proceed westward through a cumbersome cattle gate—be sure to close the gate.

53.9 Intersection with north-south dirt road along the range front. Turn south (left).

55.0 Y-intersection with a dirt road going southeast. Take right track (i.e., continue south).

55.9 Junction with dirt road that proceeds west up Short Creek (see Figs. 4 and 9b). Turn west (right).

56.9 Park along the side of the road. Four wheel drive is needed to proceed farther.

STOP 2.3. Jurassic ductile shear zone and syntectonic Big Creek pluton.

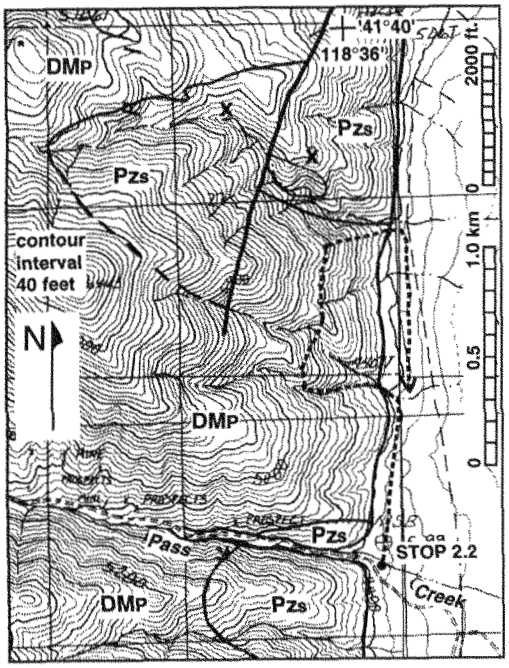
This stop is located on the Howard Hot Springs 7.5 minute topographic map and is shown on Figures 4 and 9b. At this stop, we will see rocks that experienced the highest regional strain and metamorphic grade in the east-central Pine Forest Range. The units involved are the Short Creek amphibolite and the Willow Creek schist, both of which were metamorphosed to amphibolite grade during deformation, based on the syntectonic assemblage in metamafic rocks (Fig. 6). Deformation occurred within the Short Willow shear zone (Figs. 4, 6) and produced a mylonitic foliation with a well-developed stretching lineation; this mylonitic fabric is obvious in hand-sample but recrystallized on a microscopic scale and is therefore best described as annealed or blastomylonitic. Recrystallization is interpreted to be due in part to high temperatures attained during deformation, but is probably also related to some extent to thermal effects associated with intrusion of the Cretaceous Granite Mountain plutonic complex (Fig. 4).

Foliation within the Short Willow shear zone strikes, on average, to the west-northwest, and dips, on average moderately to the south-southwest (Fig. 4); however, there is some variation in foliation attitude due to the presence of Cretaceous folding related to intrusion of the Granite Mountain plutonic complex. These folds are generally gentle to open, and are visible at the outcrop scale or manifested by larger-scale variations in the attitude of the Jurassic foliation.

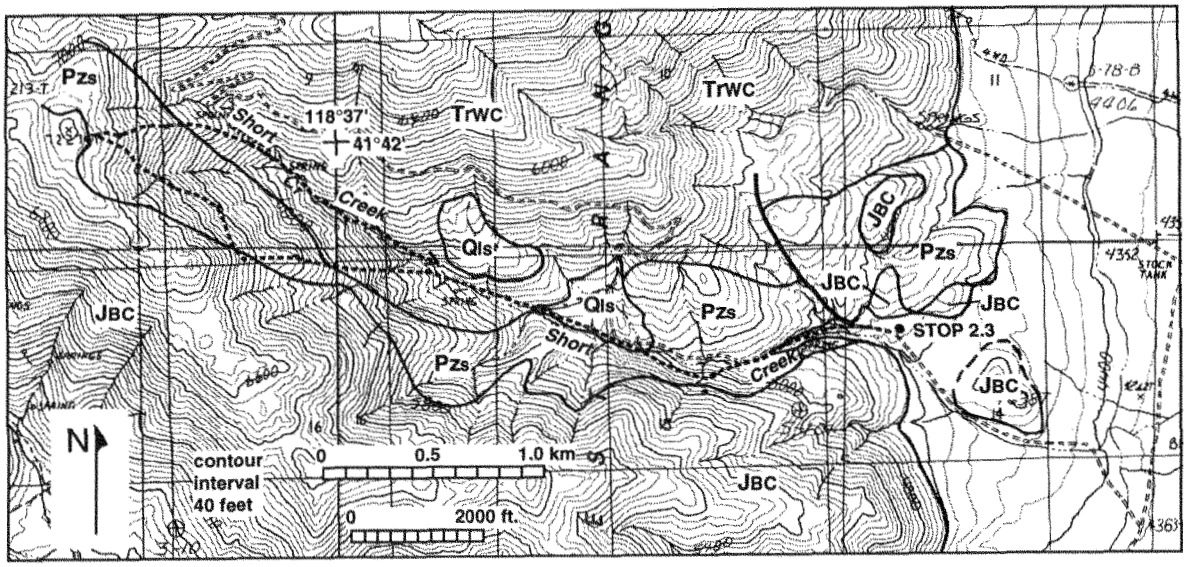
At this stop, we will observe structures and shear sense indicators within the shear zone. The Willow Creek metaplutonic schist is granodioritic in overall composition but is so deformed that protolith textures are completely unrecognizable. Foliation in these rocks is defined by highly-strained quartz ribbons, preferred alignment of mafic minerals (biotite ± hornblende ± epidote), and a pronounced metamorphic segregation layering between mafic and felsic minerals. The subhorizontal stretching lineation is defined by strained quartz and clusters of mafic minerals. Shear sense indicators are not obvious at the outcrop scale, but S-C fabrics are locally evident in thin section and indicate a dextral sense of motion (in present-day coordinates). The Short Creek amphibolite consists mostly of metabasaltic rocks, but also includes minor metapelitic biotite schists and rare marbles. Deformation has greatly obscured protolith textures in the Short Creek amphibolite but it is locally possible to discern that the metabasalts were derived from a combination of lava and volcaniclastic rock (Wyld, 1990). Foliation in these rocks is defined by the preferred alignment of metamorphic minerals, including, depending on rock type, hornblende, epidote, sphene, clinopyroxene, plagioclase, biotite, muscovite, and andalusite. The subhorizontal stretching lineation is defined by long axes of hornblende and strained volcanic clasts in amphibolites, and by long axes of andalusite in metapelitic rocks. Shear sense is defined by sigma structures at various scales (formed around relict phenocrysts, amygdules and volcanic clasts in metavolcanogenic rocks) and consistently indicates dextral sense of shear (in present-day coordinates). Recall, however, that original shear sense can only be determined by reconstructing the orientation of Jurassic structures prior to post-Jurassic tilting of the stratigraphic section (see further discussion below).

We will also examine relations indicating that the Short Willow shear zone developed in the Early Jurassic at ~201 Ma. Relevant relations include the intensely deformed and metamorphosed state of the ~230 Ma Willow Creek metaplutonic schist, which indicates that this pluton was clearly intruded prior to deformation, and a variety of structural and metamorphic relations in and around the 201 Ma Big Creek pluton indicating that this pluton was intruded during deformation, as described further below (Wyld, 1996).

As noted in an earlier section (see text for stop 1.2), tilting of the stratigraphic section in the east-central Pine Forest Range to the south is interpreted to have occurred after regional deformation and metamorphism, and the Short Willow shear zone is interpreted to have originated as a subhoriztontal ductile thrust, with the lower-strain, lower-grade rocks located to the south of the shear zone representing the less deformed upper plate of the shear zone. Apparent dextral shear sense indicators in the shear zone therefore reconstruct to an original sense of shear (i.e., prior to tilting) that is broadly top-to-the-northwest, as discussed earlier. A schematic model showing interpreted Early Jurassic relations (prior to tilting of the stratigraphic succession) is presented in Figure 10. The ductile thrust is inferred to have developed at a depth of about 8–10 km. This depth is based on the structural thickness of the upper plate and the presence of syntectonic andalusite and cordierite in metapelites near the upper boundary of the shear zone (Figs. 4, 6) which requires metamorphism at depths of <~9 km. Intrusion of the syntectonic Big



A.



B.

Figure 9. Locations of stops 2.2 and 2.3 and respective traverses on topographic base. A. Shows Pass Creek area near range front and stop 2.2. B. Shows Short Creek drainage and stop 2.3. Unit symbols same as in Figure 4, except Qls which is Quaternary landslide.

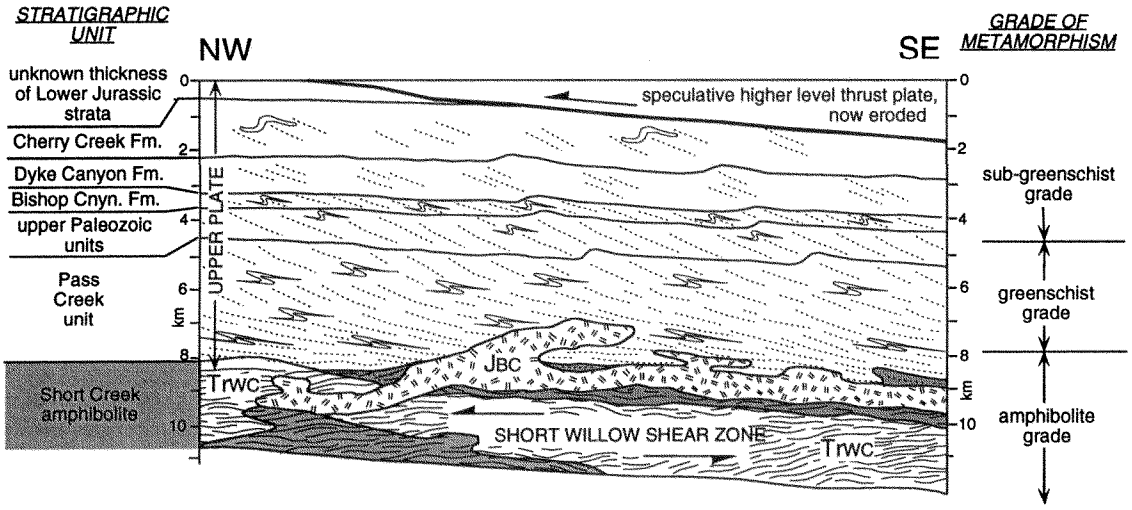


Figure 10. Schematic cross-section showing inferred structural and metamorphic relations with depth in the Pine Forest Range during Early Jurassic deformation and metamorphism. Depths to boundaries between different metamorphic facies is approximate. Squiggly lines depict morphology of megascopic folds at different crustal depths. Dotted lines depict orientation (with respect to stratigraphic layering) and relative pervasiveness of regional foliation at different crustal depths. Wavy lines are shear zone. JBC and TrWC are the Big Creek pluton and Willow Creek meta-plutonic schist, respectively.

Creek pluton appears to have been localized along the upper levels of the shear zone (Fig. 10). Strain and metamorphic grade decrease upsection within the upper plate of the shear zone.

We speculate that there may have been a higher level thrust plate emplaced during regional deformation onto the younger part of the section, but that this thrust is now eroded or covered by Tertiary strata (Fig. 10). The basis for this suggestion is that the regional foliation in even the lowest strain Triassic rocks of the east-central Pine Forest Range is generally at a low angle to layering (Figs. 4, 7a; Wyld, 1996). While a small angle between layering and foliation is reasonable within or near a layer-parallel shear zone due to high shear strain (Ramsay, 1980), this relation is not expected many kilometers away from a shear zone. The vertical load imparted by a higher level thrust plate, however, could contribute to layer-perpendicular flattening during a regime of layer-parallel simple shear (e.g., Northrup, 1996).

From the parking location, hike west up Short Creek (Fig. 9b). At about 700 m west of the parking spot, go south uphill toward the Big Creek pluton (Fig. 9b). Metapelites occur in the Short Creek amphibolite unit near the pluton contact in this area, and these rocks have a well-developed foliation defined by mica and a stretching lineation defined by aligned crystals of syntectonic andalusite. From here, return back downhill and then proceed west up the ravine formed by (ephemeral) Short Creek (Fig. 9b). At about 5400 feet elevation, angle westward uphill. From here, we will hike generally westward toward the Big Creek pluton, eventually returning back downhill to the Short Creek ravine. Along this part of the traverse we will be able to examine excellent exposures of the Willow Creek schist and metamafic rocks of the Short Creek amphibolite, observing the mylonitic foliation, the stretching lineation, and a variety of shear sense indicators. Near the pluton contact key relations indicating syntectonic intrusion can be found. These include dikes, emanating off the pluton, that were intruded parallel to the regional wall rock foliation and which contain an internal foliation parallel to that in the wall rocks. Some blebs of plutonic rock also occur near the contact that were apparently sheared away from the main pluton body during deformation and incorporated as lozenge-shaped lenses in wall rock amphibolites; these blebs have foliated margins and are oriented with their long axes parallel to the wall rock foliation (Wyld, 1996). Finally, we will observe that the main body of the Big Creek pluton is generally unfoliated in this area, although locally there is a weak foliation near the pluton margin (Fig. 3). Collectively, these relations are consistent only with an interpretation that the Big Creek pluton was intruded during regional deformation and shear zone development, but that it was probably late syntectonic.

Return to vehicles and retrace route to Highway 140.

57.9 Junction with range front road. Turn north (left).
58.8 Y-junction. Continue north.
59.8 Junction with east-west road. Turn east (right).
62.8 Intersection with Hwy 140. Turn north (left).
79.6 Denio Junction.

DAY 3: JACKSON MOUNTAINS AND BASINAL TERRANE

0.0 Denio Junction. Proceed south on Highway 140.

29.8 Junction with paved road leading south to the Jackson Mountains (Fig. 3). This road is a short distance beyond a ranch on the south side of the highway. Turn south (right).

31.4 Intersection with gravel road. Stay on paved road going south along the west side of the Jackson Mountains.

37.4 Junction with gravel road going west. Stay on range front road going south.

49.8 Pullover spot on side of road.

Stop 3.1. Geology of the Jackson Mountains

This stop is on the Hobo Canyon 7.5 minute topographic map and is shown on Figure 11. Here we will discuss the geology of the Jackson Mountains which is similar in many respects to that of the Pine Forest Range, except that more Mesozoic rocks and fewer Paleozoic rocks are exposed, and Mesozoic strata as young as Cretaceous are present (Figs. 11, 12; Russell, 1984; Quinn, 1996; Quinn et al., 1997). The Jackson Mountains thus preserves higher Mesozoic crustal levels than the Pine Forest Range.

The geology of the Jackson Mountains is dominated by the Happy Creek igneous complex (Fig. 11). This complex was previously considered to consist in large part of basaltic to andesitic volcanic strata, which were believed to have been deposited over the time span from latest Triassic or Early Jurassic through the Late Jurassic (Willden, 1964; Russell, 1984; Maher, 1989). Recent studies by Quinn et al. (1997), however, have demonstrated that the Happy Creek complex is instead composed largely of hypabyssal subvolcanic intrusions with only minor supracrustal volcanogenic strata (Fig. 11). Furthermore, Quinn et al. (1997) have shown that the complex was emplaced entirely during the latest Triassic(?) to Early Jurassic because it cuts strata as young as Norian and is cut by plutons dated at 196–190 Ma (U–Pb zircon) (Figs. 11, 12).

Flanking the Happy Creek complex to the east and west are older strata (Triassic Boulder Creek beds) that were bowed up and cross-cut by the complex during its emplacement (Figs. 11–13). These strata are entirely of Late Triassic (Carnian to Norian) age (Russell, 1984; Maher, 1989) and are divisible into three main units that are correlative with Triassic units in the Pine Forest Range based on composition and facies (Quinn, 1996; Quinn et al., 1997). Units Trb1 and Trb2 on the west side of the Jackson Mountains (Fig. 11) are correlated with the Bishop Canyon and Dyke Canyon formations of the Pine Forest Range, respectively, and unit Trb3 on the east side of the Jackson Mountains correlates with the Cherry Creek formation in the Pine Forest Range.

Paleozoic rocks occur only on the west side of the Jackson Mountains, where they are everywhere in fault contact with Triassic strata (Fig. 11, 13; Quinn, 1996). The most widespread Paleozoic unit is the McGill Canyon formation, which contains Mississippian fossils and consists of metamorphosed shale, less common quartz-rich sandstone and minor chert-pebble conglomerate (Figs. 11, 12). This unit correlates with the Pass Creek unit of the Pine Forest Range based on composition, facies and age. Other Paleozoic or inferred Paleozoic units on the west side of the Jackson Mountains include the following (Quinn, 1996). (1) A Permian volcaniclastic unit, which consists of volcaniclastic rocks and minor carbonates with Permian fossils, and which unconformably overlies the McGill Canyon formation (Figs. 11, 12). (2) The Hobo Canyon amphibolite, which structurally overlies the McGill Canyon formation on the west side of the Jackson Mountains (Fig. 11–13), and consists mostly of amphibolite-grade metamafic schist. This unit is not directly dated, but is inferred to be older than the McGill Canyon formation because it structurally overlies the latter unit along a thrust fault and because this thrust fault is one of a stacked series of thrusts that elsewhere demonstrably place older, higher-grade rocks over younger, lower-grade rocks (Figs. 11, 13; Quinn, 1996). This unit is considered likely to be correlative with the compositionally similar Short Creek amphibolite of the Pine Forest Range, and is therefore inferred to be of Devonian age (Fig. 12).

Unconformably overlying the Happy Creek complex along the central axis of the Jackson Mountains is the King Lear Formation (Figs. 11–13). This formation consists of subaerially-deposited, Lower Cretaceous sandstone and conglomerate, with minor shale and felsic volcaniclastic rock (Willden, 1964; Russell, 1984; Quinn et al., 1997).

Early Mesozoic Magmatism and Deformation

Lower Mesozoic igneous rocks in the Jackson Mountains include Norian volcanogenic strata of the Boulder Creek beds, the uppermost Triassic(?) to Lower Jurassic Happy Creek complex, and several Jurassic plutons (Figs. 11, 12). Three of these plutons have yielded Early Jurassic U–Pb zircon ages (196–190 Ma) and two have yielded Middle Jurassic U–Pb zircon ages (~170 and 163 Ma) (Fig. 11, including inset map; Quinn et al., 1997). Thus, like the Pine Forest Range, the most voluminous early Mesozoic magmatism in the Jackson Mountains occurred in the latest Triassic to Early Jurassic, whereas there is only minor evidence of magmatism in the Middle to Late Jurassic time frame.

Two phases of penetrative Mesozoic shortening deformation, D1 and D2, are recognized in the Jackson Moun-

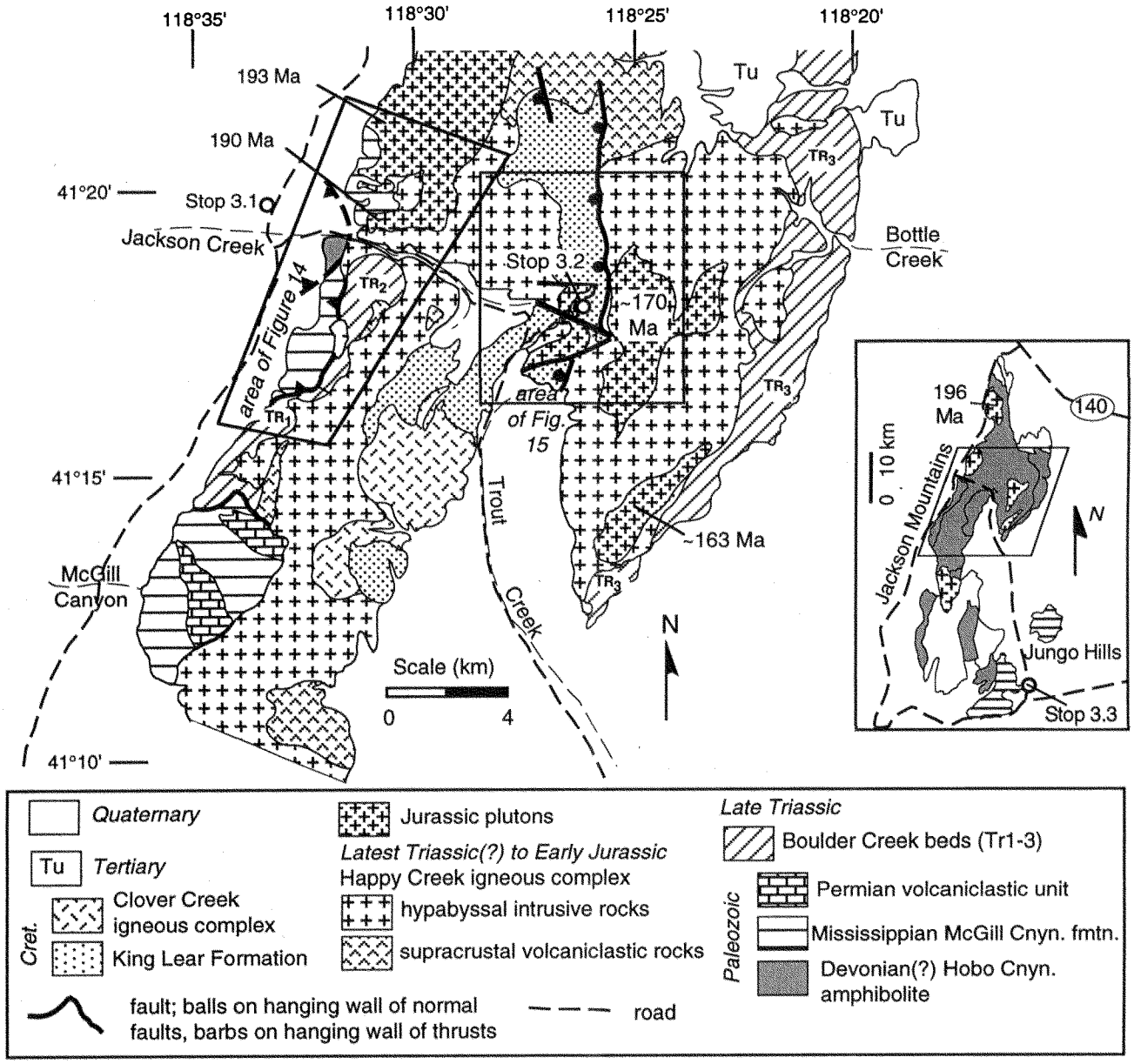


Figure 11. Generalized geology of the central Jackson Mountains (from Quinn, 1996, and Quinn et al., 1997; see outlined area in inset map), and locations of stops 3.1–3.3. Inset map shows entire Jackson Mountains and Jungo Hills area (see Fig. 3 for locations of these features). Patterns in inset map as follows: shading is Paleozoic-Mesozoic magmatic arc and related rocks exclusive of Jurassic plutons which are shown with cross pattern; striped pattern is Triassic basinal terrane; unpatterned is Tertiary. Ages of plutons are U–Pb zircon dates (Quinn et al., 1997).

tains (Quinn, 1996; Wyld et al., 1996). D1 deformation is evident only in the two older Triassic units (Tr1-2) and the Paleozoic units on the west side of the range, whereas D2 deformation affects the Happy Creek complex and older units (Quinn, 1996). D1 and D2 were separated by an episode of faulting that juxtaposed Paleozoic and Triassic rocks on the west side of the range (Fig. 11; Quinn, 1996).

The main manifestation of D1 deformation is a foliation that varies in strain and metamorphic grade with stratigraphic depth (Quinn, 1996). Thus, S1 is a low-grade spaced cleavage in Triassic Boulder Creek beds units Tr1–2 and in the Permian volcaniclastic unit, a generally phyllitic to schistose foliation defined by subgreenschist to greenschist grade metamorphic minerals and stretched pebbles in the

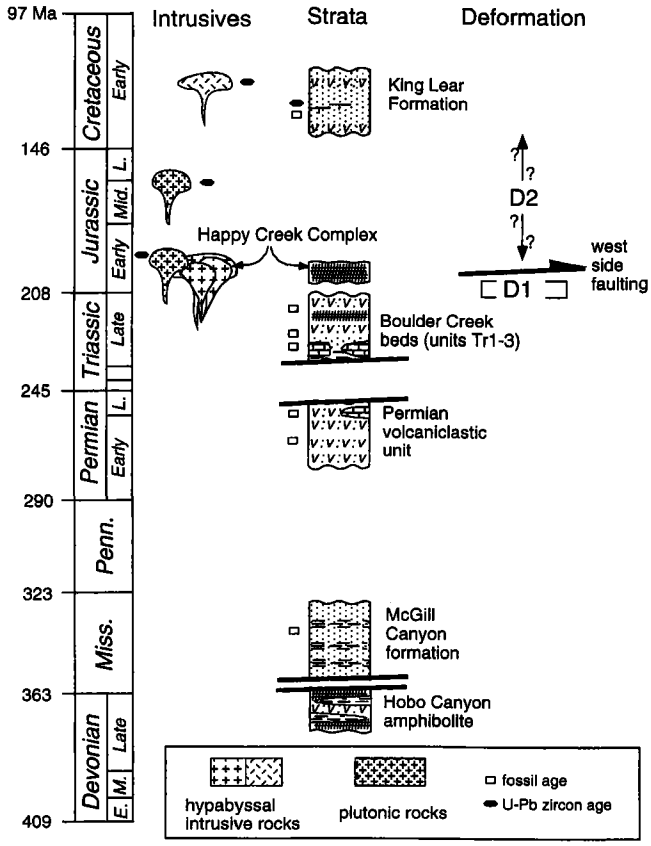


Figure 12. Summary diagram showing stratigraphy of Jackson Mountains, age span of intrusive rocks, and timing of deformation and faulting. Modified from Quinn (1996). Symbols same as in Figure 5. Age data from Willden (1964), Russell (1984), and Quinn et al. (1997).

Mississippian McGill Canyon formation, and an amphibolite-grade, schistose to locally mylonitic foliation in the Devonian(?) Hobo Canyon amphibolite. Orientation of the S1 foliation varies due to later D2 deformation. The D1 fabric affects rocks as young as Norian, is overprinted by contact metamorphism in the aureole of the 193 Ma Parrot Peak pluton, and is cut by hypabyssal rocks of the Happy Creek complex (Quinn, 1996). D1 deformation therefore occurred in the latest Triassic to mid-Early Jurassic (Fig. 12). In addition, syntectonic hornblendes from the Hobo Canyon amphibolite yielded a $^{40}Ar/^{39}Ar$ plateau age of ~200 Ma (Quinn, 1996).

D1 deformation in the Jackson Mountains is identical in both timing and style to the regional Jurassic deformation of the Pine Forest Range, and is thus considered to have occurred in response to the same phenomena—development of a broadly layer-parallel shear zone at the stratigraphic depth of mid-Paleozoic rocks coupled with a pattern of decreasing strain and metamorphic grade upsection away from the shear zone in its upper plate (Quinn, 1996). Lack of evidence for D1 deformation in the youngest Boulder Creek beds strata on the east side of the Jackson Mountains (unit Tr3) is interpreted to reflect the fact that these units were at high stratigraphic levels during D1 deformation and did not accumulate sufficient strain to be recognizable through the effects of subsequent D2 deformation.

Several faults are present on the west side of the Jackson Mountains, including a stacked series of west-dipping thrust faults near Jackson Creek and a south-dipping reverse fault south of Bliss Canyon (Figs. 11, 13). The thrusts place older rocks deformed to higher grades during D1 deformation progressively over younger rocks deformed at lower grades during D1 deformation (Quinn, 1996). The reverse fault places the McGill Canyon formation over Triassic strata. All of these faults affect rocks as young as Norian, cut D1 structures, and are cut by hypabyssal rocks of the Happy Creek complex (Fig. 11; Quinn, 1996). They were therefore all active during the early Early Jurassic.

D2 deformation produced a NE-striking, steeply-dipping cleavage that is variably developed in the Happy Creek complex and all older rocks units, and was accompanied by subgreenschist grade metamorphism with no evidence of any variation in strain or metamorphic grade with stratigraphic depth (Quinn, 1996). In general, the D2 cleavage is best developed in unit Tr3 and in Paleozoic and Triassic strata near the western margin of the Happy Creek complex, and is less prominent or pervasive in rocks of the Happy Creek complex. This deformation was previously considered to have affected the Cretaceous King Lear Formation, and was therefore previously considered to be Cretaceous in age (Russell, 1984; Maher, 1989). Recent work by Quinn et al. (1997), however, has shown that the King Lear Formation is neither deformed nor metamorphosed and that deposition of this unit therefore postdates D2 deformation. D2 deformation thus occurred sometime in the interval between the mid-Early Jurassic and the mid-Early Cretaceous (Fig. 12). The structural significance of this deformation remains unclear at present because its age is uncertain and because no equivalent deformation is recognized in the Pine Forest Range. One possibility is that it is related to strain associated with intrusion of the massive hypabyssal suite making up most of the Happy Creek complex (Fig. 11)—this could explain its heterogeneous development within the Jackson Mountains and its absence in the Pine Forest Range where no equivalent large intrusive complex is present. Further study will be required, however, to resolve this problem.

At this stop, we have a good view of the key elements of the geology of the western Jackson Mountains (Fig. 14). The Jackson Creek valley is the large valley directly to the east. To the far south are prominent limestone cliffs of

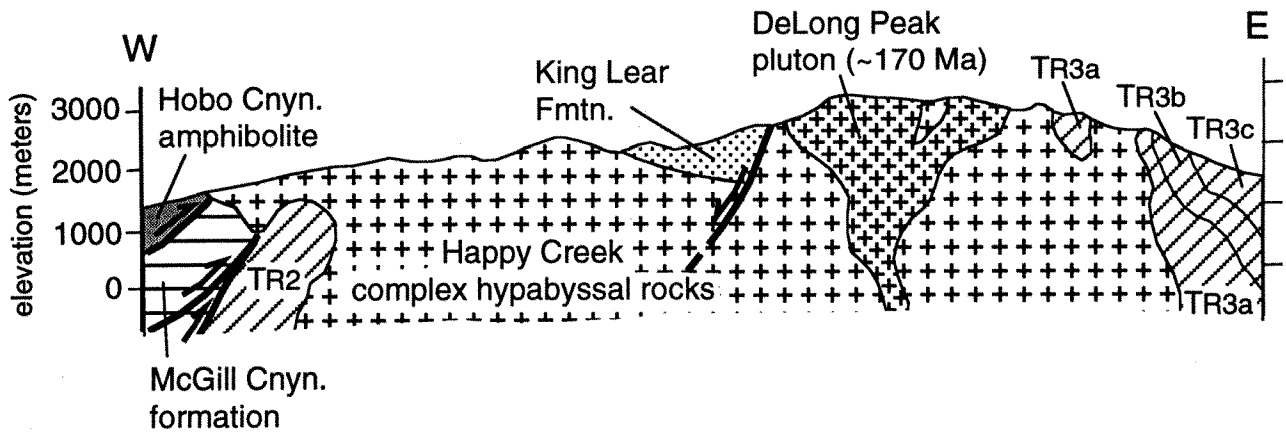


Figure 13. Generalized east-west cross section through the central Jackson Mountains near Jackson Creek (modified after Quinn et al., 1997). Patterns same as in Figure 11. Units TR3a–c are different members of Boulder Creek bed unit TR3.

Boulder Creek beds unit Tr1, equivalent to the Bishop Canyon formation of the Pine Forest Range. Previously considered to include coherent beds or units of carbonate (Russell, 1984; Maher, 1989), detailed mapping by Quinn (1996) indicates that the limestones occur as large to small, irregularly-shaped bodies completely surrounded by clastic rocks. Based on these relations, Quinn (1996) argued that the limestones are large olistoliths that slumped or slid from a carbonate bank downslope. Structurally above Triassic strata, just south of Jackson Creek, are a series of west-dipping thrust sheets of metamorphic rock (Figs. 13, 14; Quinn, 1996). The two structurally lower thrust sheets imbricated the McGill Canyon formation and emplaced these greenschist-grade rocks over subgreenschist-grade Boulder Creek beds unit Tr2. The structurally higher thrust sheet, which underlies the low hills just south of the mouth of Jackson Creek, places the amphibolite-grade Devonian(?) Hobo Canyon amphibolite over the McGill Canyon formation (Figs. 13, 14). The prominent high topography behind (to the east of) these Paleozoic and Triassic rocks is underlain mostly by hypabyssal basalts and andesites of the Happy Creek complex, which intrude across the thrust faults in several places (Fig. 14).

To the north of Jackson Creek, the high country near the range front is underlain by the composite Parrot Peak pluton and Harrison Grove stock (Figs. 11, 14), which yielded U–Pb zircon ages of 193 Ma and 190 Ma, respectively (Quinn et al., 1997). These plutons intrude the hypabyssal rocks of the Happy Creek complex, which are exposed primarily south and east of the plutons. The low, poorly exposed hills west of the plutons are underlain by the McGill Canyon formation (Fig. 14). The greenschist facies regional foliation in these metaclastic rocks is overprinted by contact metamorphism in the aureole of the plutons.

Straight east, up the Jackson Creek valley, you can see the massive, prominent, dark cliffs of the Happy Creek hypabyssal complex. When we drive up the valley (see directions below), you will notice that there is no sign of stratigraphic layering or lithologic variation within these cliffs. This is because the cliffs are made up of monotonously homogeneous hypabyssal intrusive rocks rather than the layered volcanic succession inferred by previous workers (Quinn et al., 1997).

Drive south 0.6 miles along the range front road.

50.4 Junction with gravel road going east up Jackson Creek (Fig. 14). Turn east (left).

56.2 Intersection with dirt road to the north. Stay on main road going east.

56.9 This is the pass between the east-west trending Jackson Creek drainage and the south-trending Trout Creek drainage (Fig. 11). Here, the main road turns south down the Trout Creek valley. Here also there is an intersection with a dirt road going uphill to the east (Fig. 11, 15). Turn onto the latter road and proceed uphill.

59.1 Pullover spot at hairpin turn (Fig. 15).

STOP 3.2. King Lear Formation and basal unconformity.

This stop is on the Parrot Peak 7.5 minute topographic map and is shown on Figure 15. Here we will look at the King Lear Formation and the unconformity separating this unit from older rocks of the Jackson Mountains.

The King Lear Formation was deposited in a relatively short interval in the Early Cretaceous because a tuff near the base of the formation has been dated at ~125 Ma, and the Clover Creek igneous complex, which intrudes the upper part of the formation, has been dated at ~123 Ma

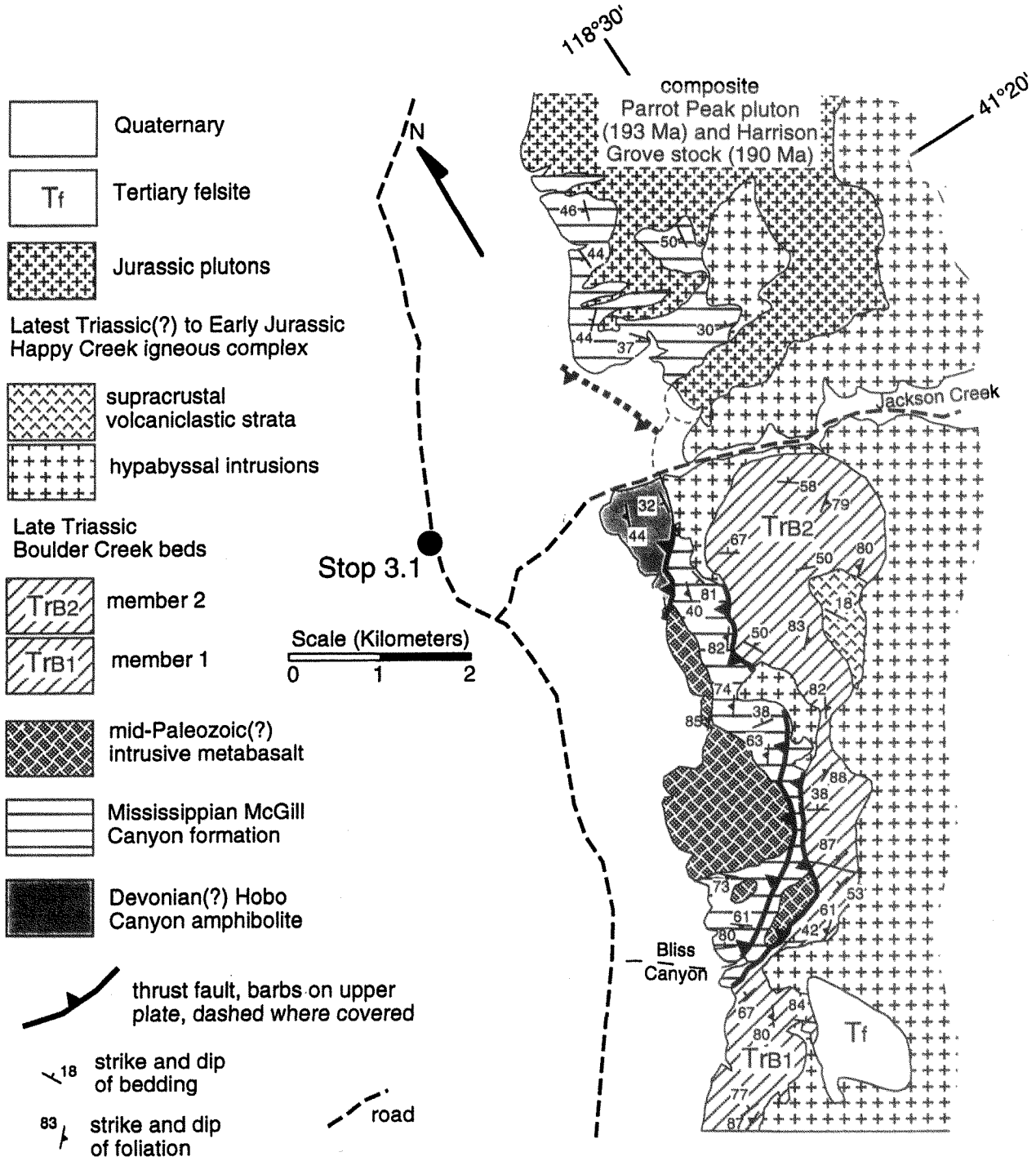


Figure 14. Geology of the western Jackson Mountains, near Jackson Creek (see boxed area in Fig. 11); figure modified from Quinn (1996). Note that the Jurassic plutons cut the Happy Creek complex and the Happy Creek complex cuts across the thrust faults.

(U–Pb zircon; Quinn et al., 1997; Fig. 15). The basal contact of the formation is an unconformity which spans the Middle and Late Jurassic as well as perhaps portions of the Early Jurassic and/or Early Cretaceous (Figs. 12, 15; Russell, 1984; Quinn et al., 1997). The King Lear Formation was previously interpreted to have been deposited as a molasse sequence shed from a west-vergent thrust belt that imbricated the Happy Creek complex and older strata (Russell,

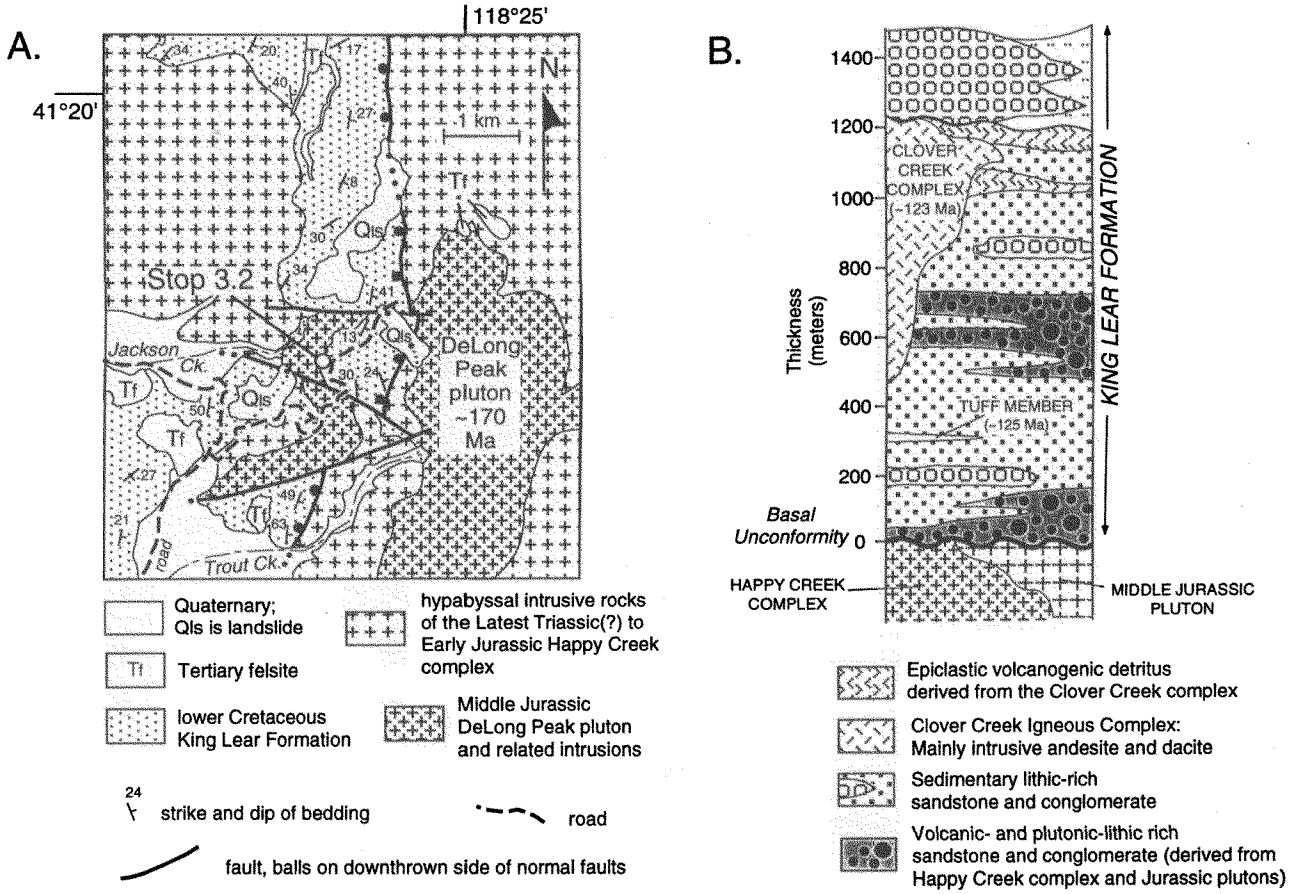


Figure 15. A. Geologic map showing the central part of the King Lear Formation (simplified from Quinn, 1996); age data (U–Pb zircon) from Quinn et al. (1997). B. Generalized stratigraphic architecture of the King Lear Formation (from Quinn et al., 1997).

1984; Maher, 1989). This interpretation has recently been shown to be incorrect. Specifically, data reported in Quinn (1996) and Quinn et al. (1997) indicate that the hypothesized west-vergent thrust system does not exist. Instead, it is now recognized that the King Lear Formation was deposited in an extensional half-graben, whose eastern boundary fault is a high-angle normal fault (Figs. 11, 13, 15a).

Walk to the west and drop downhill several meters to see spectacular outcrops of the King Lear Formation and the unconformity between this unit and underlying Jurassic plutonic rock. The basal part of the King Lear Formation, here and elsewhere in the Jackson Mountains, consists of conglomerate containing rounded pebbles and cobbles of andesitic rock and dioritic plutonic rock (Fig. 15b; Quinn et al., 1997). These clasts are identical to rocks of the Happy Creek complex and Jurassic plutons and are inferred to have been derived from erosion of these rocks. Underlying the King Lear Formation, along an unconformity that is well exposed in this area, are fine-grained diorites associated with the ~170 Ma DeLong Peak pluton (Fig. 15; Quinn et al., 1997).

Two features are interesting about this basal King Lear unconformity. First, the King Lear Formation sits unconformably on only three rock units in the Jackson Mountains: hypabyssal rocks of the Happy Creek complex, supracrustal rocks of the Happy Creek complex, and part of a fine-grained Middle Jurassic pluton (Figs. 11, 15a). Second, the basal part of the King Lear Formation contains clasts derived from erosion of the Happy Creek complex and Jurassic plutonic rocks (Fig. 15b; Quinn, 1996). It therefore appears that Jurassic tectonism in the Jackson Mountains had not resulted in widespread uplift and exposure of Triassic or Paleozoic basement rocks by the time the King Lear Formation was first deposited in the Early Cretaceous. This is consistent with relations in the Pine Forest Range indicating that the Black Rock Desert was not subjected to any major shortening deformation in the Middle to Late Jurassic—in other words, if the Black Rock Desert had been

subjected to two major phases of Jurassic shortening deformation, one in the Early Jurassic as documented above, and one in the Middle to Late Jurassic as is seen in many other arc assemblages of the western U.S. Cordillera, then it seems likely that the combined effects of this shortening would have resulted in significant uplift and widespread exposure of Triassic and Paleozoic basement rocks. The relations described above thus add further support to the conclusion, based on relations in the Pine Forest Range, that there is no significant Middle to Late Jurassic deformation recorded in the Black Rock Desert segment of the continental arc.

To the west down the Jackson Creek valley are good views of hypabyssal rocks of the Happy Creek complex. From here, it can clearly be seen that there is no stratification in these rocks and that they crop out in the massive, homogeneous, cliffy exposures typical of plutonic rocks. To the north, strata of the King Lear Formation dip homoclinally at a low angle to the east on top of the Happy Creek complex (Fig. 13a), attesting to the lack of any significant deformation during or after King Lear deposition.

Return to vehicles and drive back downhill to the main road at the pass.

61.5 Junction with main road at the pass. Turn south (left) down Trout Creek valley (Figs. 11, 15a).

72.4 Y-intersection with gravel road going north. Stay on road going south.

87.1 Pull over by low hills and park.

STOP 3.3. Triassic basinal terrane and Luning-Fencemaker fold-and-thrust-belt in the Jungo Hills area

This stop is located on the Jungo 7.5 minute topographic map and is shown on Figure 11 (inset). At this stop, we will look at deformed sedimentary strata of the Triassic basinal terrane (Figs. 3, 11 inset), part of the Luning-Fencemaker fold-and-thrust belt. Strata in this area are typical of the basinal terrane and consist mostly of metamorphosed mudstone with less common quartz-rich sandstone and minor calcareous rocks, although some dacitic sills are also locally abundant (Wyld, work in progress). Similar sedimentary strata crop out from here east to Winnemucca (Fig. 3; Willden, 1964).

This part of the basinal terrane lies very close to magmatic arc and related strata of the Black Rock Desert province (Fig. 11 inset) and is one of the few places where the contact between the arc province and the back-arc basinal terrane may be exposed. The only published geologic map of this area is that of Willden (1964). His study shows the contact between Jackson Mountains Paleozoic-Mesozoic arc rocks and basinal terrane rocks as a high-angle fault. Wyld (unpublished data) mapped this area in detail in 1996, and found that a high-angle fault is present, but that it separates strata typical of the basinal terrane (metapelite the dominant rock type) to the east from strata that are similar but contain more abundant carbonate. It is thus not clear that this fault, whose sense of displacement is indeterminate at this time, actually forms the boundary between the arc province to the west and the back-arc basinal terrane to the east. Rather, it appears that the fault may separate mudstone-dominated facies of the basinal terrane from a more carbonate-dominated facies of the same depositional basin. Further mapping is needed to the west of this area to determine whether the arc-basinal terrane contact is anywhere exposed or whether any facies or structural changes can be found that would shed light on the nature of the arc-basinal terrane boundary. At present, the nature of the arc-basinal terrane boundary remains unknown.

Hike from the parking spot west a short distance up the low hills through deformed strata of the basinal terrane. The principal structure seen in these rocks is a well-developed slaty to phyllitic cleavage (S1) parallel to bedding which strikes northeast and dips to the northwest. Our ongoing studies in basinal terrane strata throughout the Jungo Hills area, in Blue Mountain and in the Santa Rosa Range (Fig. 3) indicate that this cleavage is axial planar to tight to isoclinal folds of bedding that are overturned to the southeast (F1), and that these structures (S1 and F1) are the dominant structures in this part of the Luning-Fencemaker fold-and-thrust belt. Younger phases of folding are locally evident in these rocks, including D2 open folds and D3 crenulation folds, but these later structures are heterogeneously distributed and represent substantially less shortening strain.

It is difficult at present to conclusively link deformation in this part of the Luning-Fencemaker belt with Early Jurassic deformation in the Black Rock Desert for three reasons. First, the original orientation of Early Jurassic structures in the Pine Forest Range are uncertain due to substantial post-Jurassic tilting, and Early Jurassic structures in the Jackson Mountains have been reoriented by younger deformation that may be related to intrusion of hypabyssal rocks of the Happy Creek complex, as described above. Thus, it is difficult to link structures between the arc and the back-arc on the basis of orientation. Second, little is known about the boundary between Black Rock Desert arc rocks and back-arc basinal strata, as noted above, and it is therefore not possible at present to examine how structures may be related across this boundary. Finally, the timing of deformation in the Luning-Fencemaker fold-and-thrust belt is poorly constrained and it is therefore difficult to relate structures between the arc and the back-arc on the basis of age.

Previous studies in the Luning-Fencemaker belt have inferred that most deformation probably occurred in the

Middle to Late Jurassic (Oldow, 1984; Speed et al., 1988; and references therein). This inference is based in part on imprecise dating (K–Ar ages) of intrusive rocks that may be syntectonic, although structural relations are somewhat ambiguous. This inference has probably also been influenced by the fact that Middle to Late Jurassic deformation is widespread in parts of the western U.S. Cordillera. Our new data from the Black Rock Desert, indicating that this area was affected by major Early Jurassic shortening with little evidence for Middle to Late Jurassic deformation, suggests that the timing of deformation in the Luning-Fencemaker fold-and-thrust belt needs to be examined more carefully. Specifically, since the arc region immediately west of the Luning-Fencemaker fold-and-thrust belt was deformed primarily in the Early Jurassic, then it would appear likely that this may also be the age of major shortening in the adjacent back-arc, particularly since our ongoing studies indicates that most shortening in this part of the Luning-Fencemaker fold-and-thrust belt was accommodated during a single phase of deformation.

This problem is an important one to resolve because the Luning-Fencemaker belt is a major belt of shortening within the western U.S. Cordillera, and one which reflects the final closure of marine basins within the more internal parts of the continental margin (Fig. 1). Shifting the timing of main-phase shortening within this belt from Middle or Late Jurassic to Early Jurassic, some 30–50 m.y. earlier, would thus have major implications for paleogeographic, structural and tectonic reconstructions of the Cordillera during the Mesozoic. In particular, if most shortening in the Luning-Fencemaker fold-and-thrust belt occurred in the Early Jurassic, this would support the model depicted in Figure 2 and described in Wyld et al. (1996) arguing that the early Mesozoic arc and back-arc in the vicinity of the Black Rock Desert experienced a different Jurassic structural evolution from the Klamath Mountains-to-Arizona segment of the continental arc because these provinces occupied different parts of the continental margin prior to arc-parallel displacement on the Cretaceous Mojave-Snow Lake strike-slip fault. Further work on the timing of deformation in the Luning-Fencemaker fold-and-thrust belt is clearly needed to resolve this important question.

Return to vehicles and proceed south.

88.5 Intersection with major gravel road. Turn east (left).

123.1 Intersection with Highway 95/Melarkey street. Turn east (right).

123.6 Intersection with Winnemucca Boulevard. Downtown Winnemucca.

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