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STRUCTURE OF THE NORTHERN SIERRA NEVADA, CALIFORNIA

by

WILLIAM J. SCHMIDT

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

DOCTOR OF PHILOSOPHY

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May, 1985
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STRUCTURE OF THE NORTHERN SIERRA NEVADA, CALIFORNIA

I. INTRODUCTION

A. Preamble

Since the advent of the theory of plate tectonics, numerous tectonic models that attempt to explain the tectonic evolution of the North American Cordillera have been proposed (Burchfiel and Davis, 1981; Dickinson, 1977; Schweickert, 1981). Paleontological, paleomagnetic, geochemical and structural data support the hypothesis that much of western North America is allochthonous in nature, having been accreted to the western edge of the continent during Phanerozoic time (Burchfiel and Davis, 1981; Speed, 1979; Jones and others, 1978). While the exotic nature of some of the rocks of the Cordilleran is well established, there are two major problems that thwart attempts to unravel the details of the accretionary history of this region. The first and foremost problem is the incomplete nature of the existing rock record: important pieces of an extremely complex puzzle have been permanently lost via subduction, rifting, strike-slip faulting, erosion, or other geologic processes. The second problem is the lack of detailed data. Only by close examination of the existing rock record can we hope to gain a more detailed picture of the tectonic evolution of the Cordilleran.

B. Thesis Objective

The Sierra Nevada of California is one area within the Cordilleran that has received a relatively large amount of attention from geologists over the past 100 years. Although much progress has been made in piecing together the tectonic history of this portion of the Cordilleran, much work needs to be done before a more accurate account of the Phanerozoic tectonic activity can be given. Research dealing with the detailed structure of deformed rock bodies
has received little regard in the northern portion of the Sierra Nevada even though these data are extremely useful. The objective of this study is to investigate in detail the structure of a portion the northern Sierra Nevada in order to determine the timing and kinematics of deformation for this area. An attempt will also be made to correlate the deformational history of the rocks in the study area with the tectonic evolution of the Sierra Nevada as a whole.

C. Procedure

In order to determine the structural history of the deformed rocks in the study area, structural features on megascopic, mesoscopic, and microscopic scales and the metamorphic history of the area have been examined. Megascopic analysis involved field mapping of rock units and large-scale structures. Data collected for mesoscopic analysis include the orientation of planar and linear features and the analysis of fold profile geometries. On a microscopic scale lattice preferred orientation of quartz and calcite have been examined as well as the relationship between the various fabric elements. The finite strain ellipsoid for one area was determined from measurements made on the microscopic scale (deformed amygdules in a metabasalt).

Field work was carried out during the summers of 1979 and 1980. As shown in Figure 1, detailed data have been collected on an east-west traverse (normal to the regional structural trend) and in several areas along a line trending north-northwest (approximately on-strike with the regional structural trends). Data from the east-west traverse were collected along road cuts and in the canyons of the Middle Fork of the American River and the Rubicon River; heavy vegetation prohibited data collection outside of these areas. This traverse begins at the Melones fault zone near the town of Foresthill and ends where undeformed granitic rocks of the Sierra Nevada batholith crop out about
Figure 1. Map showing location of study areas in the northern Sierra Nevada.
2 km east of the Ellicott bridge in the Robbs Peak quadrangle. Due to the relatively good exposure of rock units, it was possible to make detailed geologic maps of three areas along the north-northwest traverse near Bowman Lake, Cisco Butte, and Loon Lake (see Figure 1 and Plates 1, 2, and 7). In addition, data were collected along a portion of the North Fork of the American River, about 8 km south of Cisco Butte.
II. REGIONAL GEOLOGY

A. Nature and Distribution of Principal Rock Systems

The general composition and distribution of the pre-Mesozoic rocks that the Sierra Nevada batholith intrudes have been outlined by Bateman and Clark (1974). They note that the wall rocks exposed on opposite flanks of the Sierra Nevada batholith (Figure 2) are quite different. The strata on the eastern flank of the batholith consist of Late Precambrian stratified rocks, which are overlain by a structurally conformable sequence of Paleozoic rocks. This package of pre-Mesozoic rocks is composed primarily of carbonates, shales, and clean quartzites. On the other hand, the Paleozoic section on the western flank consists of a suite of rocks that apparently owe their origin to an oceanic or island-arc environment. Rock types include thick sections of volcanic and volcanioclastic rocks, chert, argillite, and lesser amounts of carbonates.

Mesozoic strata of both the eastern and western flanks of the Sierra Nevada are made up primarily of volcanic and volcanioclastic rocks. Bateman and Clark (1974) note that the eastern strata, which were deposited in shallow water or subaerially, have an average rhyodacitic composition. The western Mesozoic strata, which were deposited in a marine environment in close proximity to (or within) an island arc, have an average composition somewhere between that of andesite and dacite (Bateman and Clark, 1974).

The emplacement history of the Sierra Nevada batholith is constrained by numerous K-Ar and U-Pb age dates. Problems with argon loss due to reheating make K-Ar dates for older igneous events suspect (Stern and others, 1981; Chen and Moore, 1982). The U-Pb age dates indicate three major plutonic epochs, excluding allochthonous (?) small plutons (e.g., Don Pedro and Chinese Camp plutons) that crop out west of the Melones fault zone (Chen and Moore, 1982; Stern and others, 1981). Granitoids from the oldest epoch were intruded 215-
Figure 2. Generalized geologic map of the Sierra Nevada from Bateman and Clark (1974).
202 m.y.B.P. in the eastern portion of the Sierra Nevada (near the towns of Lee Vining and Bishop) (Chen and Moore, 1982). These Triassic–Early Jurassic (the DNAG time scale of Palmer, 1983, is used in the present work) granitoids comprise a relatively small volume compared to the voluminous granitoids of the two subsequent epochs (Stern and others, 1981). The next intrusive epoch (185–155 m.y.B.P.) forms a belt that trends N 40° W (Bateman, 1981; Stern and others, 1981). This Jurassic–age belt is cross-cut by rocks of the Cretaceous intrusive epoch (125–88 m.y.B.P.) (Stern and others, 1981; Bateman, 1981). The Cretaceous granitoids form a belt that trends N 20° W (Bateman, 1981). Cretaceous plutonism migrated steadily to the east–northeast and terminated at about 80 m.y.B.P. (Chen and Moore, 1982). Gross compositional differences between the quartz dioritic rocks in the western Sierra Nevada and granodioritic rocks to the east (Kistler and others, 1971) are apparently due to differences in the composition of the lower crust and upper mantle (Bateman, 1981).

B. Major Structures

The Foothills Fault system, located on the western flank of the Sierra Nevada, comprises a complicated system of anastomosing faults that follow the north–northwest trend of the Sierra Nevada; the faults dip steeply to the east–northeast. The fault system is over 300 km in length ranging in width from 10 to 40 km. Conclusive evidence indicating the sense of displacement is apparently lacking because the faults of this area have been interpreted as both dip–slip and strike–slip. At least two workers (Clark, 1960; Cebull, 1972) in the area who initially supported the likelihood of strike–slip displacement now favor dip–slip displacement along faults of the area (Clark, 1976; Bateman and Clark, 1974; Cebull and Russell, 1979; Russell and Cebull, 1974). Davis (1969)
suggested that the Melones fault was a southwest-directed low-angle thrust, before it was rotated about a northwest axis to the vertical during Late Jurassic times. The last large-scale displacements along the Foothills Fault system coincide with the Nevadan orogeny, an intense but short-lived event (less than 10 m.y.), which ended just prior to Cretaceous time (Bateman and Clark, 1974; Schweickert and others, 1984). This important structural feature will be discussed in more detail later in this chapter and in the concluding chapters.

The interpretation of the gross structural configuration of the wall rocks of the Sierra Nevada has been a bone of contention among Sierran geologists. The most frequently stated interpretation is that the wall rocks form a faulted synclinorium (Bateman and others, 1963; Clark, 1964; Bateman and Wahrhaftig, 1966; Bateman and Eaton, 1967; Bateman and Clark, 1974; Schweickert, 1981). This interpretation is based upon the observation that the wall rocks on both flanks of the Sierra Nevada dip inward, toward the central axis of the Sierra Nevada. While this may be true, Hamilton (1969) aptly points out that the available data are insufficient to substantiate the existence of a synclinorium. Hamilton notes that the faults, which juxtapose profoundly different rock assemblages, in the wall rocks on the western flank of the Sierra Nevada "do not represent stratigraphic displacements within a continuous sedimentary section as the synclinorial conjecture would require" (1969, p. 2421). Hamilton explains the faulted east-dipping metamorphic rocks of the western flank in terms of an accretionary model. He infers that most of the pre–Upper Jurassic rocks have an oceanic origin and have been accreted via subduction to the western edge of the continent during Mesozoic time. Russell and Nokleberg (1977) have offered yet another view. They propose that the wall rocks of the

C. Regional Deformation

Effects of as many as five regional deformations are recognized in the wall-rocks of the Sierra Nevada (Nokleberg and Kistler, 1980). The following discussion is limited to the contiguous wallrocks of the western metamorphic belt (Clark, 1964) and the roof pendants to the east. The isolated roof pendants to the south are discussed by Saleeby (1981) and Schweickert (1981).

Early Paleozoic deformation in the northern Sierra Nevada is indicated by the angular discordance between the Ordovician-Silurian Shoo Fly Formation and overlying Late Devonian rocks (D'Allura and others, 1977; Varga and Moores, 1984). Nokleberg and Kistler (1980) have summarized the evidence for early to middle Paleozoic deformation in the roof pendants of the central Sierra Nevada. They note five different localities where Devonian or Mississippian deformation most likely occurred. The age of deformation is reasonably well constrained at the Mount Morrison roof pendant where Silurian(?) and Ordovician rocks were deformed prior to deposition of Pennsylvanian-age rocks. Similarities between rock types and the orientation of structures (northerly striking) in other roof pendants in the central Sierra Nevada implies that this event may have been of regional extent.

During late Paleozoic to early Mesozoic time at least two deformational events affected the rocks of the western metamorphic belt. Evidence for these two events is supported by the fact that the Permo-Carboniferous(?) Calaveras Complex was metamorphosed and twice deformed prior to the intrusion of the 170 m.y. Standard pluton (Schweickert, 1981). Further support for these two events is found along the North Fork of the American River where mid-Permian
rocks are twice deformed prior to deposition of Late Triassic (Norian) and Jurassic age rocks (Harwood, 1983).

Isotopic dates on the late Paleozoic–early Mesozoic deformational events yield Triassic to Early Jurassic ages. A Late Triassic age (215±10 m.y.) for deformation and metamorphism of orthogneisses that intrude the Shoo Fly Formation of the western metamorphic belt has been determined from isotopic dating of zircon, sphene, and amphibole (Sharp and others, 1982). An Ar40/Ar39 age determination of 236±4 m.y. (Middle Triassic) for deformation and metamorphism in the Feather River ultramafic complex is reported by Weisenberg (1979). Blueschist facies metamorphism near Downieville occurred prior to 174 m.y.B.P. and may be older than 190 m.y.B.P. (Early Jurassic) based upon K–Ar age determinations (Schweickert and others, 1980). Melange formation and associated regional metamorphism occurred in the Foothills Fault system (Bear Mountains fault zone) during Early Jurassic time (190–200 m.y.B.P.) based on numerous K–Ar age determination (Saleeby, 1982).

Late Paleozoic to early Mesozoic structures in the Sierra Nevada have a strikingly similar orientation; planar features throughout this area strike north to northwest, dipping steeply (Figure 3 and Table 1).

Deformational features resulting from the Late Jurassic Nevadan orogeny are evident throughout the Sierra Nevada. Schweickert and others (1984) review the age constraints for Nevadan deformation. An age of 150–158 m.y. for the Nevadan orogeny is well constrained by radiometric and stratigraphic data.

The orientation of Nevadan structures for many localities in the Sierra Nevada are shown in Figure 4. Most Nevadan structures trend north to northwest, although data from three localities indicate a northeasterly trend.
Figure 3. Strike of late Paleozoic–early Mesozoic structures in the northern Sierra Nevada. References, average orientation of planar features, and age constraints are listed in Table 1.
Late Paleozoic-Early Mesozoic Trends
## TABLE I

LATE PALEozoIC-EARLy MesozoIC TRENDs

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</tr>
<tr>
<td>2a</td>
<td>Standlee, 1978</td>
<td>330/90</td>
<td>Deforms Ordo-Silurian Shoo Fly Fm.</td>
</tr>
<tr>
<td>2b</td>
<td>Standlee, 1978</td>
<td>5/90</td>
<td>Deforms Ordo-Silurian Shoo Fly Fm.</td>
</tr>
<tr>
<td>3a</td>
<td>Schmidt, present work</td>
<td>355/90</td>
<td>Deforms Early Permian Peale Fm.</td>
</tr>
<tr>
<td>3b</td>
<td>Schmidt, present work</td>
<td>5/85E</td>
<td>Deforms Early Permian Peale Fm. but not Early Jurassic Sailor Canyon Fm.</td>
</tr>
<tr>
<td>3c</td>
<td>Schmidt, present work</td>
<td>0/85W</td>
<td>Deforms Ordo-Silurian Shoo Fly Fm.</td>
</tr>
<tr>
<td>4</td>
<td>Chandra, 1961</td>
<td>335/85E</td>
<td>Deforms Lower Carboniferous but not Oxfordian age rocks Early to Middle Triassic</td>
</tr>
<tr>
<td>5</td>
<td>Nokleberg and Kistler, 1980 (Saddlebag Lake)</td>
<td>335/90</td>
<td>Early to Middle Triassic</td>
</tr>
<tr>
<td>6</td>
<td>Nokleberg and Kistler, 1980 (Mount Dana)</td>
<td>335/?</td>
<td>.</td>
</tr>
<tr>
<td>7</td>
<td>Nokleberg and Kistler, 1980 (Ritter Range)</td>
<td>335/90</td>
<td>Early to Middle Triassic</td>
</tr>
</tbody>
</table>
In the southern portion of the western metamorphic belt northwest-striking
Nevadan structures are folded about structures with a northeasterly trend.
Schweickert and others (1984) believe that a late pulse of the Nevadan orogeny
is responsible for the formation of the northeasterly trending structures.
Definitive data constraining the age of the so called "late-phase" Nevadan
structures (Schweickert and others, 1984) are lacking. The fact that
Schweickert and others (1984) assign these structures to a late phase of the
Nevadan orogeny is apparently based upon radiometric dating of plutons (154
m.y.) that cross-cut northeast-trending foliation in the Bowman Lake vicinity.
Yet, the northeast-trending structures in the Bowman Lake area are most likely
"main-phase" Nevadan structures as argued in a later chapter (Chapter VII.B.3).
Within the Late Jurassic (Oxfordian to Kimmeridgian) Mariposa Formation in
the southern portion of the western metamorphic belt, northeast-trending
structures cross-cut the northwest Nevadan structures (Figure 4). A regionally
extensive northeast-trending Cretaceous deformational event post-dates the
northeast-trending event, thus bracketing the age of deformation as latest
Jurassic or Early Cretaceous.

The last regional deformation in the Sierra Nevada occurred during
Cretaceous time resulting in structures that have a consistent northwest strike
(Figure 5, Table III). The age of this event has been reasonably well constrained
in the roof pendants of the central Sierra Nevada (Nokleberg and Kistler, 1980).
Nokleberg and Kistler (1980) summarize the radiometric and stratigraphic data
that support a late Early to early Late Cretaceous (100–80 m.y.) age for this
event. Tobisch and Fiske (1982) report an age of 92$^{+7}$ m.y.B.P. for Cretaceous
deformation in the eastern Sierra Nevada based upon radiometric dating of
deformed units and undeformed plutons.
Figure 4. Strike of Nevadan and late Nevadan structures in the northern Sierra Nevada. References, average orientation of planar features, and age constraints are listed in Table II.
Nevadan and Late Nevadan Trends
**TABLE II**

<table>
<thead>
<tr>
<th>Number</th>
<th>Reference</th>
<th>Average Orientation</th>
<th>Age Constraints</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Varga, 1980</td>
<td>340/90</td>
<td>Deforms Permo-Triassic Arlington Fm. 175 m.y. or younger (isochron on blueschists deformed during D2)</td>
</tr>
<tr>
<td>2</td>
<td>Cebull and Russell, 1979</td>
<td>355/80E</td>
<td>Deforms Early Permian Peale Fm. Deforms Early Jurassic Sailor Canyon Fm.</td>
</tr>
<tr>
<td>3a</td>
<td>Schmidt, present work</td>
<td>40/85E</td>
<td>Deforms late Paleozoic(?) rocks Deforms Ordo-Silurian Shoo Fly Fm. Folds Early Jurassic or younger(?) rocks</td>
</tr>
<tr>
<td>3b</td>
<td>Schmidt, present work</td>
<td>40/85W</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>3c</td>
<td>Schmidt, present work</td>
<td>40/90</td>
<td>Deforms late Paleozoic(?) rocks Deforms Ordo-Silurian Shoo Fly Fm. Folds Early Jurassic or younger(?) rocks</td>
</tr>
<tr>
<td>3d</td>
<td>Schmidt, present work</td>
<td>0/90</td>
<td>Deforms late Paleozoic(?) rocks Deforms Ordo-Silurian Shoo Fly Fm. Folds Early Jurassic or younger(?) rocks</td>
</tr>
<tr>
<td>4</td>
<td>Loomis, 1961</td>
<td>335/90</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>5</td>
<td>Chandra, 1961</td>
<td>335/75E</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>6a</td>
<td>Duffield and Sharp, 1975</td>
<td>352/80E</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>6b</td>
<td>Duffield and Sharp, 1975</td>
<td>342/90</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>7a*</td>
<td>Ave Lallemant, unpublished data</td>
<td>55/85N</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>7b*</td>
<td>Ave Lallemant, unpublished data</td>
<td>55/85N</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>8</td>
<td>Baird, 1962</td>
<td>330/80E</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>9</td>
<td>Eric et al., 1955</td>
<td>325/80E</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>10a*</td>
<td>Schweickert et al., 1984</td>
<td>25/90</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>10b*</td>
<td>Schweickert et al., 1984</td>
<td>30/90</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>11</td>
<td>Nokleberg and Kistler, 1980</td>
<td>330/90</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
<tr>
<td>12</td>
<td>Best, 1962</td>
<td>320/80E</td>
<td>Deforms Oxfordian—Kimmeridgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Deforms Late Oxfordian to Early Kimmersidgian rocks Second phase in Late Jurassic Mariposa Fm. Third Phase in late Paleozoic(?) Calaveras Fm. Deforms late Paleozoic(?) Calaveras Fm. Deforms Callovian Logtown Ridge Fm. Deforms Nevada structures in Shoo Fly and Calaveras Fms. Second phase in Late Jurassic Mariposa Fm. Deforms Jurassic rocks, Early Cretaceous rocks undeformed Deforms Oxfordian—Kimmeridgian rocks</td>
</tr>
</tbody>
</table>

*Late Nevadan localities denoted by asterisk.*
Figure 5. Strike of Cretaceous structures in the northern Sierra Nevada. References, average orientation of planar features, and age constraints are listed in Table III.
### TABLE III

**CRETACEOUS TRENDS**

<table>
<thead>
<tr>
<th>Number</th>
<th>Reference</th>
<th>Average Orientation</th>
<th>Age Constraints</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a</td>
<td>Schmidt, present work</td>
<td>315/90</td>
<td>Last phase in Paleozoic sequence</td>
</tr>
<tr>
<td>1b</td>
<td>Schmidt, present work</td>
<td>315/90</td>
<td>Last phase in late Paleozoic sequence</td>
</tr>
<tr>
<td>1c</td>
<td>Schmidt, present work</td>
<td>315/90</td>
<td>Last phase in early Paleozoic sequence</td>
</tr>
<tr>
<td>1d</td>
<td>Schmidt, present work</td>
<td>320/90</td>
<td>Last phase in early Paleozoic sequence</td>
</tr>
<tr>
<td>2</td>
<td>Eric et al., 1955</td>
<td>340/90</td>
<td>Second phase in Callovian Logtown Ridge Fm.</td>
</tr>
<tr>
<td>3</td>
<td>Nokleberg and Kistler, 1980 (Saddlebag Lake)</td>
<td>300/90</td>
<td>—</td>
</tr>
<tr>
<td>4</td>
<td>Nokleberg and Kistler, 1980 (Mount Dana)</td>
<td>295/7</td>
<td>Lower Cretaceous isochron</td>
</tr>
<tr>
<td>5</td>
<td>Nokleberg and Kistler, 1980 (Ritter Range)</td>
<td>300/7</td>
<td>Lower Cretaceous isochron</td>
</tr>
<tr>
<td>6</td>
<td>Best, 1962</td>
<td>338/30E</td>
<td>Second phase in Oxfordian—Kimmeridgian rocks</td>
</tr>
</tbody>
</table>
The Cretaceous deformational event folds older structures throughout the Sierra Nevada. The fact that all older structures exhibit varied orientations (Figures 3 and 4) whereas Cretaceous structures exhibit a consistent northwest trend (Figure 5) supports the hypothesis that older structures throughout the Sierra Nevada were reoriented prior to or during the Cretaceous deformational event.

D. Summary of the Regional Geologic History

Much of the geologic history of the rocks that comprise the Sierra Nevada province is poorly constrained. Paleogeographic reconstructions of this province and adjacent areas over the past decade have relied heavily upon plate-tectonic models to aid in deciphering the complex geologic history of this portion of the Cordilleran. The following paragraphs briefly outline the geologic history of the Sierra Nevada province and adjacent areas. Emphasis will be placed on examining the major tectonic events. Various plate-tectonic models that have been set forth to explain the tectonic evolution of this region will be examined in the light of the supporting geologic data.

Latest Precambrian to Late Devonian rocks of the southern Cordilleran were deposited along a passive continental shelf (Stewart and Suczek, 1977). Latest Precambrian to Lower Cambrian volcanic and terrigenous clastic rocks are overlain by Middle Cambrian to Upper Devonian carbonates (Stewart and Suczek, 1977; Ross, 1977; Poole and others, 1977).

During the latest Devonian to Early Mississippian Antler orogeny, a eugeosynclinal assemblage of rocks (clastic rocks and chert intercalated with volcanic rocks) was thrust eastward along the Roberts Mountains thrust over shelf deposits (miogeosynclinal rocks consisting primarily of limestone and dolomite), which had formed along the passive margin of the North American
continent (Roberts and others, 1958; Speed and Sleep, 1982). This event, which records the thrusting of oceanic rocks over shelf deposits, is considered by Dickinson (1977) to be a form of obduction. Basically, two tectonic models have been called upon to explain the Antler orogeny (Figure 6). One model involves backarc thrusting (Burchfiel and Davis, 1972, 1975; Silberling, 1973; Poole, 1974; Poole and others, 1977; Poole and Sandberg, 1977; Miller and others, 1984) while the other explains this event in terms of crustal collision (Moores, 1970; Burchfiel and Davis, 1972; Churkin, 1974; Schweickert, 1976; Dickinson, 1977; Speed and Sleep, 1982; Oldow, 1984).

The backarc-thrusting model assumes a Japan-type margin for the southern Cordilleran prior to the Antler orogeny. This configuration involves east-dipping subduction beneath an island arc that was separated from the North American continent by a marginal basin. Obduction occurs as the result of closure or collapse of the marginal basin. Devonian-age metamorphism and westward vergent folding in the Klamath Mountains may have occurred as a result of normal-polarity subduction (east-dipping) beneath an island arc adjacent to the North American continent (Burchfiel and Davis, 1972, 1975). Siluro-Devonian arc volcanic rocks in the Klamath Mountains (Murray and Condie, 1973) and Devono-Carboniferous arc volcanic rocks in the northern Sierra Nevada (Schweickert and Cowan, 1975) also may be related to west-facing island arc subduction (Dickinson, 1977).

Not all the available data support the backarc-thrusting hypothesis. The presence of arc-derived detritus within the Siluro-Devonian oceanic sequence of the Roberts Mountains allochthon has not been documented (Stewart and Poole, 1974). One would expect arc-derived detritus to make up a significant portion of the sediments of a collapsed marginal basin of modest dimensions. Further-
Figure 5. Two possible tectonic models for the Late Devonian Antler orogeny.
more, the short duration of the Antler orogeny discounts the backarc-thrusting model. Johnson (1971) notes that the Roberts Mountains allochthon was probably displaced several centimeters per year. This is an order of magnitude faster than Dickinson's (1976) determined strain rates for Mesozoic backarc thrusting in the Rocky Mountains.

Many workers in the southern Cordilleran believe that the Antler orogeny is best explained by the collision of an exotic east-facing island arc with the North American continent (Dickinson and others, 1983; Speed and Sleep, 1982; Oldow, 1984). Structures in the Roberts Mountains allochthon and the juxtapositioning of allochthonous oceanic rocks with autochthonous shelf rocks are best explained by the collisional model (Oldow, 1984; Dickinson and others, 1983). Foreland basin deposition is also compatible with the arc-continent collisional model (Speed and Sleep, 1982; Dickinson and others, 1983). The crustal collision model is supported by the probable rapid displacement (several centimeters per year) of the Roberts Mountains allochthon, because in this model displacement is the result of thrusting along a west-dipping subduction zone. The paucity of arc-derived detritus in the Roberts Mountains allochthon is also in better accordance with the collisional model.

The collisional model relies on west-dipping subduction for emplacement of the Roberts Mountains allochthon. Although there is no conclusive evidence favoring reversed polarity during the Late Devonian, there is support for this view. The imbricate oceanic assemblage that comprises the Roberts Mountains allochthon is very similar to accretionary prism terranes that form on the leading edge of the overriding plate at consuming plate boundaries (Speed and Sleep, 1982; Dickinson and others, 1983; Oldow, 1984). If the Roberts Mountains allochthon represents an accretionary prism that was thrust eastward
over shelf deposits during the Antler orogeny, a collisional event involving west-dipping subduction is the favored model for Antler orogenesis.

There is abundant support for a Carboniferous rift event in the southern Cordilleran (Dickinson, 1977). The anomalous evolution of the Antler foreland basin, with flysch deposits overlain by shelf deposits (Bissel, 1974) and not molasse, implies that normal foreland basin evolution was aborted by some tectonic process. The many local basins of Permian-Pennsylvanian age within the Antler orogen (Roberts and others, 1958) may indicate that a fault-controlled rift topography existed at this time, while the turbidites of the Oquirrh basin (Larson, 1976) may represent fill of a Carboniferous aulacogen. Further support for a Carboniferous rift event is found in the Permian-Pennsylvanian oceanic assemblages (e.g., the Havallah sequence) of the Golconda allochthon, which was thrust eastward over the Antler allochthon and its terrestrial and marine cover (Silberling and Roberts, 1962; Dickinson and others, 1983). Dickinson (1977) notes that the Permo-Carboniferous arc volcanics now exposed in northern California are difficult to relate to this rift event.

The onset of the Mesozoic Epoch in the southern Cordilleran was marked by the Late Permian to Early Triassic Sonoma orogeny (Schweickert and Snyder, 1981). Striking similarities between this event and the Antler orogeny are apparent. Both orogenies involved eastward thrusting of imbricate oceanic assemblages over a previously quiescent continental margin. The Golconda allochthon is composed of ocean-floor strata consisting primarily of turbidites, mafic volcanic rocks, and hemipelagic and pelagic rocks (Speed, 1979). Speed (1979) notes that arc-derived debris, depositional melange, and blueschist are absent from this assembly of thrust nappes. Emplacement of the Golconda allochthon was completed by Early Triassic time (Dickinson and others, 1983).
Due to the obvious similarities of the Antler and Sonoma orogenies, the models for these events are also similar (Figure 6). The Sonoma orogeny has been explained in terms of backarc thrusting (Burchfiel and Davis, 1972, 1975; Silberling, 1973; Miller and others, 1984) and crustal collision of a distant arc (Burchfiel and Davis, 1972; Churkin, 1974a, 1974b; Schweickert, 1976a; Dickinson, 1977; Speed, 1979; Dickinson and others, 1983).

Two backarc-thrusting models have been proposed for the Sonoma orogeny. In the first model (Burchfiel and Davis, 1972, 1975) the assumed backarc-thrusting of the Sonoma orogeny is thought to be a continuation of backarc-thrusting associated with the Antler orogeny. In this scenario the trapped marginal basin of Antler time was not completely closed during the Antler orogeny. Following a period of tectonic quiescence, backarc-thrusting resumed during the Sonoma orogeny. The other backarc-thrusting model (Burchfiel and Davis, 1972; Silberling, 1973) calls upon backarc-spreading following the Antler orogeny to create a marginal basin, which subsequently closed during Sonoman backarc-thrusting.

Most data appear to be in better agreement with the distant-arc collisional model while discounting the backarc-thrusting models. The distant-arc collisional model is supported by the absence of arc-derived sediments in the Golconda allochthon, the relatively undeformed nature of the suballochthonous rocks and the similarity of composition and structure of the Golconda allochthon to subduction complexes (Speed, 1979).

A major change in Cordilleran paleogeography occurred during the Mesozoic. Paleozoic and early Mesozoic structural and depositional trends, which strike northeast, were replaced in Late Triassic time with a northwest-trending magmatic arc (Burchfiel and Davis, 1972; Hamilton and Meyers, 1966;
Hamilton, 1969). The truncation of northeast trends in the southern Cordilleran can be explained by one of three models: continental subduction, rifting, or major strike-slip faulting. The continental subduction model is unlikely due to difficulties in initiating and sustaining subduction of continental crust. Convergent-plate motion implied by the Sonoma orogeny and the subsequent Late Triassic magmatic arc (Evernden and Kistler, 1970) argues against the possibility of a rifting event. Thus, the truncation of Paleozoic and early Mesozoic trends by strike-slip faulting is the preferred model for this event. The truncation event probably occurred in Early Triassic time (Burchfiel and Davis, 1981).

The early Mesozoic truncation event has been related to the proposed suturing of a distant microplate, Sonomia, to the western edge of North America during the Sonoma orogeny by Speed (1979). Speed’s model depicts the truncation event as a result of left-lateral motion along a transform fault initiated at an unstable triple junction that juxtaposed North America, Sonomia, and an oceanic plate. Davis and others (1978) also view this truncation event as the result of left-lateral strike-slip motion, which is supported by the apparently continuous magmatic activity in the Klamath Mountains during Permian and Triassic time. Apparent sinistral offset of continental basement and Paleozoic strata also supports left-lateral displacement (Saleeby, 1981).

In the western metamorphic belt of the Sierra Nevada the location of the line of continental truncation appears to coincide with the westernmost exposure of the Shoo Fly Formation (Burchfiel and Davis, 1981). Most workers (Davis and others, 1978; Buchfiel and Davis, 1981; Schweickert, 1976; Saleeby, 1981) agree that in the northern Sierra Nevada the line of continental truncation coincides with the Melones fault zone, which also coincides with the
westernmost exposure of the Shoo Fly Formation. In the central Sierra Nevada the westernmost exposures of the Shoo Fly Formation are thrust westward over the Calaveras Complex (Schweickert, 1981). Schweickert (1981) calls this fault, which separates the lower Paleozoic Shoo Fly Formation from the upper Paleozoic (?) Calaveras Complex, the Calaveras-Shoo Fly thrust. The Calaveras-Shoo Fly thrust likely coincides with the line of truncation in the central Sierra Nevada (Burchfiel and Davis, 1981; Saleeby, 1981). Extensive plutonism in the southern Sierra make it difficult to locate the position of the line of truncation (Burchfiel and Davis, 1981).

It is important to note that Speed's (1978, 1979) model for the Sonoma orogeny and the subsequent Early Triassic truncation event requires no continental truncation in the northern and central Sierra Nevada (i.e., the western metamorphic belt). In Speed's (1979) model, Paleozoic rocks of the western metamorphic belt are part of his allochthonous microplate, Sonomia. Speed (1978, 1979) believes that the present northwest trend along the western boundary of Sonomia, which Speed correlates with the Melones fault zone, is probably related to the western edge of "pre-collisional" Sonomia. Structural and stratigraphic data from the northern and central Sierra Nevada are in agreement with Speed's model. The basis for the truncation event, the truncation of northeast-striking Paleozoic and Early Triassic structural and depositional trends by northwest trends, is lacking in the northern and central Sierra Nevada, where Paleozoic and early Mesozoic trends strike northwest (Figures 3 and 7). The location of the 0.706 contour for the initial strontium isotopic ratios (\(^{87}\)Sr/\(^{86}\)Sr) (Kistler, 1978) also supports Speed's model. The initial strontium ratios imply that the southern Sierra Nevada is underlain by
Figure 7. Generalize geologic map of the northern Sierra Nevada from D'Allura and others, 1977.
Generalized Map, Paleozoic Rocks
Northeastern Sierra
J.A. D'Allura, E. M. Moores, L. Robinson
1977

- faults
- contacts
- bedding attitudes

- ultramafic rocks
- Felsic Intrusives, mostly presumably Mesozoic
- Jurassic Rocks
- Arlington
  - Hasselkus
  - Reeve
  - Goodhue
- Upper Poole
- Lower Poole
- Taylor
- Sierra Buttes, Elwell
- Upper Shoo Fly
- Shoo Fly, middle sheared unit
- Shoo Fly, lower quartz sandstone unit

SCALE

50 50 50 50 100 Miles
5 5 5 5 10 Km
continental lithosphere, whereas the northern Sierra Nevada (part of Speed's Sonomia) is not.

Prior to emplacement of Late Triassic to Early Jurassic plutonic rocks of the Sierra Nevada (Stern and others, 1981), plate motion along the western margin of North America had developed a component of convergent motion sufficient to produce the observed Late Triassic arc (Speed, 1979). Plutonic rocks of the Sierra Nevada and Klamath Mountains, which range in age from Late Triassic to Late Cretaceous, support Hamilton's (1969) interpretation that the Mesozoic western margin of North America was dominated by a convergent plate boundary with an east-dipping Benioff zone. The now popular idea (Nokleberg, 1983; Schweickert, 1981; Schweichert and Cowan, 1975; Burchfiel and Davis, 1972, 1975, 1981) that exotic terranes have been accreted onto the margin of western North America was first proposed by Hamilton (1969).

The juxtapositioning of the allochthonous late Paleozoic (?) Calaveras Complex with the Shoo Fly Formation to the east occurred prior to emplacement of the Standard pluton 170 m.y.B.P. (Schweickert, 1981). The Standard pluton cross-cuts two sets of structures in the Calaveras Complex, indicating that the Calaveras Complex was deformed twice prior to 170 m.y.B.P. (Schweickert, 1981). Schweickert (1981) argues convincingly for restricting the extent of the Calaveras Complex to the southern portion of the western metamorphic belt, east of the Melones fault (Figure 8). Others who work in the northern Sierra Nevada believe that the Calaveras Complex also occurs west of the Melones fault (Burchfiel and Davis, 1981; Hietanen, 1973, 1977). Schweickert (1981) makes a reasonable case for a late Paleozoic age for the Calaveras Complex, which is comprised of a chaotic assemblage of oceanic rocks (cherts, argillites, basalts, and minor marble lenses). Others consider the
Figure 3. Geologic map for the northern Sierra Nevada from Schweickert, 1981.
Calaveras Complex to be of Mesozoic age (Burchfiel and Davis, 1981; Saleeby, 1981). The age of the Calaveras-Shoo Fly thrust, which juxtaposes the Calaveras Complex with the Shoo Fly Formation to the east, is older than the 215 ± 10 m.y. metamorphic event that overprints the thrust (Sharp and others, 1982). The Calaveras Complex is bound to the west by the Melones fault zone (Schweickert, 1981) (Figure 8).

Ophiolitic rock sequences that crop out in the Foothills fault system formed about 300, 200, and 160 m.y.B.P., based upon Pb/U age dating (Saleeby, 1982). Saleeby (1982) notes that the field setting of samples from the 300 m.y. age group (plagiogranites that occur as isolated blocks in melange) differ significantly from those of the 200 and 160 m.y. age groups, which are taken from intact igneous bodies and stratigraphic sequences. Saleeby (1982) believes that 200 and 160 m.y. ophiolites formed as result of intra-arc rifting in close proximity to the melange basement of the Foothills fault system. A regional metamorphic-deformational event, which resulted in ophiolitic melange development, occurred about 190 to 200 m.y.B.P. (Saleeby, 1982).

Evidence for a Late Jurassic rifting event is not limited to the Sierra Nevada. In the Klamath Mountains, the Josephine ophiolite formed around 157 m.y.B.P. in a back-arc setting (Harper, 1984). This age is similar to the age of the Smartville ophiolite of the northern Sierra Nevada. Menzies and others (1980) note that the Smartville ophiolite formed 155 to 160 m.y.B.P. in an island-arc setting. The island-arc tectonic setting is deduced from geochemical and sedimentological constraints (Menzies and others, 1980).

The Late Jurassic Nevadan orogeny severely deformed rocks throughout the Sierra Nevada (Schweickert and others, 1984). Two dramatically different models have been proposed for the Nevadan orogeny. The collisional model
views the Nevadan orogeny as the result of a collision between an exotic east-facing island arc and a west-facing arc on the edge of North America (Schweickert, 1981; Schweickert and Cowan, 1975). The other model states that the Nevadan orogeny resulted from intraarc deformation along a west-facing arc on the edge of North America (Saleeby, 1981; Burchfiel and Davis, 1981). Both models have strengths and weaknesses but the available data seem to better support Schweickert's collisional model.

Support for Schweickert's (1981) collisional model is found in the distribution of petrotectonic terranes in the western Sierra Nevada, where western and eastern arc terranes are separated by a major fault zone (Figure 8). Schweickert (1981, p. 125) notes that there is no evidence that volcanic rocks of the western arc developed on Paleozoic rocks east of the Melones fault zone. The short-lived, intense deformation that occurred during the Nevadan orogeny is compatible with the collisional model (Schweickert and others, 1984). The collisional model also offers the best explanation for the abrupt change in relative plate motion between the North American and Farallon plates that occurred in Nevadan time (Page and Engebretson, 1984). The distribution of ophiolites that formed 150 to 160 m.y.B.P. also supports Schweickert's collisional model. The Coast Range, Josephine, and Smartville ophiolites all formed at about the same time, probably in association with Schweickert's exotic east-facing island arc (Schweickert, 1981). Schweickert (1981) notes that the stratigraphy in the Franciscan Complex makes it difficult to explain the origin of the Coast Range ophiolite in terms of Saleeby's (1981) complex arc model.

The intraarc-deformation model is supported by the extreme narrowness of the arc terranes exposed in the Sierra Nevada (Burchfiel and Davis, 1981). Saleeby (1981) contends that the basement of the western arc was emplaced
prior to arc formation, while Schweickert (1981) argues otherwise. As noted by Burchfiel and Davis (1981, p. 69), the location of the western arc basement during Jurassic volcanism "is of extreme importance to an understanding of Sierran Mesozoic tectonics." Unpublished work by Warren Sharp implies that at least a portion of the western arc basement was part of the North American continent during Jurassic volcanism (Ave Lallemant, personal communication, 1985). Jurassic plutons that crop out west of the Melones fault zone also appear to favor the intraarc-deformation model, although Schweickert (1981, p. 129) notes just because "Jurassic plutonic rocks occur both east and west of the Melones fault zone does not require that all plutonic rocks represent parts of the same arc."

Andean-type subduction occurred at the western margin of North America following the Nevadan orogeny. The convergent plate boundary had a significant left-lateral component, evidenced by the probable hundreds of kilometers of left-lateral displacement observed along the Jura-Cretaceous Pine Nut fault (Oldow and others, 1984). The Pine Nut fault is located in western Nevada and occurs in a back-arc basin setting (Oldow and others, 1984). The left-lateral strike-slip component of convergence eventually was eventually replaced by a right-lateral component. The effects of this right-lateral component are expressed in the dextral rotation of large crustal blocks along the North American plate margin. The paleomagnetic record for rocks along the western edge of North America indicate that this region has experienced a consistent clockwise rotation (Beck, 1980). Paleomagnetic data from Upper Jurassic to Lower Cretaceous intrusive rocks of the Blue Mountains of northeastern Oregon indicates that the region has experienced a $60^\circ \pm 20^\circ$ clockwise rotation since its formation (Wilson and Cox, 1980). Beck (1980) observes that the rocks north
and south of the Pacific Northwest have moved northward relative to the stable craton.

Contrary to the conclusions drawn by Hannah and Verosub (1980), paleomagnetic data from the northern Sierra Nevada are insufficient to rule out post-Jurassic oroclinal flexures in the northern Sierra Nevada. In fact, Hannah and Verosub’s (1980) paleomagnetic data support post-Jurassic oroclinal flexure! The mean declination in Hannah and Verosub’s (1980) Clio site varies about 150° to 500° from the Homer and Evans sites farther north (Figure 9). The change in orientation of the mean declination is consistent with a clockwise rotation. Hannah and Verosub (1980) erroneously conclude that the oroclinal flexure in the area predates Late Jurassic remagnetization because the 500° change in strike of bedding, between the Clio site and the northern Homer and Evans sites, is larger than the change in mean declination. Hannah and Verosub (1980) incorrectly assume a one to one relationship between change in strike of bedding and change in mean declination. In my opinion, this is a poor assumption, especially in multiply deformed areas such as the Sierra Nevada. The regional strike of formation contacts would be much more likely to exhibit a direct relationship to declination orientation in this area because the oroclinal flexure is defined by regional changes in the strike of formation contacts (Figures 7 and 8). Therefore, it is significant that the change in strike for the contact between the Peale and Reeve Formation is similar to the observed change in mean declination orientation (Figure 9).

Paleomagnetic data from the central and southernmost Sierra Nevada indicate contrasting rotational histories. In the central Sierra Nevada, no significant rotation has occurred since 90 m.y.B.P. (Frie and others, 1984); this
Figure 9. General geology of the northernmost Sierra Nevada (slightly modified from Hannah and Verosub, 1980). Paleomagnetic sample sites of Hannah and Verosub (1980) are indicated by squares. Arrows show the orientation of mean declination (without tilt correction) for the Clio, Evans, and Homer "B" sites (cf. table 1 from Hannah and Verosub, 1980).
contrasts sharply with the $45^\circ \pm 14^\circ$ rotation observed in the southernmost Sierra Nevada from 80 to 20 m.y.B.P. (Kanter and McWilliams, 1982).

The right-lateral convergent plate boundary was replaced by the San Andreas fault system, which is a right-lateral transform fault, about 30 m.y.B.P. (Atwater, 1970).
III. ROCK UNITS

A. Introduction

Work on the stratigraphy in the northern Sierra Nevada began in the late 1800’s with the pioneering work of Diller (1892, 1908) and Turner (1895, 1896, 1897). McMath’s (1966) work in the Taylorsville area helped resolve many of the stratigraphic problems of the northern Sierra Nevada. More recently, workers at the University of California at Davis (Durrel and D’Allura, 1977; D’Allura and others, 1977) have systematically described the Paleozoic rock units that crop out throughout the northern Sierra Nevada. The following discussion summarizes the descriptions of the rock units that are exposed in the northern Sierra Nevada. Subsequently, pertinent information from the present investigation will be discussed. The correlation of the rock units exposed in the Loon Lake area with similar rocks to the north will be dealt with at the end of this chapter. The prefix meta- will not be used in rock descriptions although it should be understood that pre-Tertiary rocks throughout the area have experienced varying degrees of metamorphism (most commonly greenschist facies).

B. Rock Units

The following discussion gives a general description of the rock units that crop out in the study area. The descriptions, unless noted otherwise, are taken from D’Allura and others (1977). The areal extent of the rock units of the northern Sierra Nevada is shown in Figure 7.

The Shoo Fly Formation is the oldest rock unit exposed in the northern Sierra Nevada. This formation comprises a wide variety of rock types that include siltstone, quartz sandstone, mudstone, conglomerate, and carbonate rocks with lesser amounts of chert and serpentinite. Turbidity current structures, such as graded bedding and cross bedding, are common in this unit. Standlee (1973) notes that the Shoo Fly Formation is a heterogeneous
assemblage of eugeosynclinal rocks characterized by abundant siliciclastic and a lack of volcaniclastic rocks. The Shoo Fly Formation crops out in a continuous belt throughout the northern Sierra Nevada (Figure 7). The lower contact of this formation is not exposed. The upper contact with the Sierra Buttes Formation is an angular unconformity. Varga and Moores (1980) note that in situ fossils from the Lakes Basin area indicate an Ordovician-Silurian age for the Shoo Fly Formation.

The Sierra Buttes Formation is made up of 100 to 1250 m of dacitic intrusives, volcaniclastic rocks, bipyramidal quartz keratophyre, and lesser amounts of chert. There is no direct evidence bearing on the age of this unit; however, the conformable overlying Late Devonian Elwell Formation implies that the Sierra Buttes Formation is Middle to Late Devonian in age. The Sierra Buttes Formation is most likely the result of Middle to Late Devonian (?) volcanic activity in the northern Sierra Nevada.

The Elwell Formation consists of from 1 to 700 m of black phosphatic radiolarian chert with local volcanic rocks. Ammonoid fossils found within this formation indicate a Late Devonian age. Schweickert (1981) notes that the Elwell Formation may actually be interbedded with the Sierra Buttes Formation and, thus coeval with the Sierra Buttes Formation. The contact between the Elwell Formation and the overlying Taylor Formation varies; in some places it is conformable, in others unconformable. The unconformities are apparently due to local areas of non-deposition or erosion.

The Taylor Formation is made up of 900 to 3000 m of andesitic volcaniclastic rocks and pillowed flows. Its age is bracketed between Late Devonian and Early Mississippian, the age of fossils taken from the underlying and overlying rocks, respectively.
The Peale Formation conformably overlies the Taylor Formation. The lower member, which is found only locally, consists of felsic extrusive rocks whereas the upper member, which is more widely distributed (Figure 7), is made up of 250 to 500 m of chert, shale, intraformational breccia, and lesser amounts of tuffaceous material. The rocks of the Taylor Formation and the lower member of the Peale Formation formed within, or adjacent to, an active volcanic arc during Late Devonian to Early Mississippian time. D’Allura and others (1977) note that Early Mississippian and Early Permian fossils occur in the upper member of the Peale Formation, although Schweickert (1981, p. 102) includes the Early Permian fossil locality in the overlying Reeve Formation.

The Arlington Formation and its stratigraphic equivalents (the Goodhue, Reeve, and Robinson Formations) consist primarily of volcanic and volcanioclastic rocks characteristic of volcanic arc terranes. The contact between these arc-related rocks and the underlying formations varies. In most exposures the contact appears conformable, in places even gradational; elsewhere the contact is an angular unconformity. D’Allura and others (1977) believe that the unconformity could be due to the Sonoman orogeny. Fossils from the Arlington Formation indicate an age ranging from Early Permian (Wolfcampian) to Late Triassic (Norian). Fossils from the Goodhue and Reeve Formations are of Late Pennsylvanian to Early Permian and Middle Permian age, respectively.

It is important to note that there is only one regionally extensive unconformity present in the Paleozoic and early Mesozoic section of the northern Sierra Nevada: the contact between the Shoo Fly Formation and the overlying rock units. Although other unconformable surfaces exist, none of them are regional in extent. Therefore, when considering the northern Sierra
Nevada as a whole, the Late Devonian to Late Triassic sequence is, at least in part, conformable.

The Late Triassic (Norian) Hosselkus Limestone, which crops out near Taylorsville (McMath, 1966), is similar in age to the black shale that is exposed near the top of the Arlington Formation. McMath (1966) notes that the Hosselkus Limestone rests with angular discordance on the underlying rocks in the Taylorsville area. The upper contact of the Hosselkus Limestone is the Taylorsville thrust. A probable equivalent to the Hosselkus Limestone is exposed along the North Fork of the American River (Clark, 1976 and Plate 3). The limestone at this locality has been dated as Late Triassic based upon the presence of late Karian to Norian conodonts (Harwood, 1983). This important locality will be discussed in detail in the following section.

A thick (4000 m) volcanic section exposed in the Taylorsville area ranges in age from Early Jurassic to early Late Jurassic (Callovian) (McMath, 1966). This sequence is characterized by andesitic volcanioclastic and volcanic rocks with several conglomerate beds.

Following the deposition of the Jurassic arc sequence, the rocks of the Sierra Nevada were deformed in the latest Jurassic Nevadan orogeny. Nevadan diastrophism resulted in the formation of an angularly unconformable surface throughout the Sierra Nevada. Cenozoic volcanic and sedimentary rocks unconformable overlie deformed pre-Tertiary rock units.

C. Description of Rocks at Localities in the Study Area

1. Bowman Lake area

The Shoo Fly Formation in the Bowman Lake area is made up primarily of quartz sandstone, siltstone, mudstone, limestone, and chert with minor amounts of quartz-pebble conglomerate. The overlying Sierra Buttes Formation is in
intrusive contact with the Shoo Fly Formation (Plate 1). The Sierra Buttes Formation is comprised of quartz keratophyre and dacitic(?) intrusive rocks with minor amounts of chert and black shale. The Elwell Formation, which is made up of thinly-bedded black nodular chert (Figure 10) has a conformable contact with the underlying Sierra Buttes Formation. The Taylor Formation, resting unconformably upon the Elwell and Sierra Buttes Formations, is made up of volcanioclastic and volcanic rocks with lesser amounts of conglomerate and chert (Figure 11). Graded bedding is observed in the Taylor Formation as are pillow lavas. The upper member of the Peale Formation appears to conformably overlie the Taylor Formation at this locality. The Peale Formation consists of interbedded chert and siltstone. Intraformational chert breccia and chert-pebble conglomerate (Figure 12) makes up a significant portion of the exposed upper portion of this unit. The upper contact of the Peale is in part intrusive but may be conformable where overlain by extrusive rocks of unknown age.

2. Cisco Butte area

In the area mapped around Cisco Butte four sedimentary units are exposed. The oldest is the Taylor Formation, which consists of bedded volcanioclastic rocks that grade into the bedded chert and siltstone of the overlying Peale Formation (Figure 13). The Triassic(?) limestone that overlies portions of the Peale is in fault contact at one locality (Plate 2). The general nature of this contact is thought here to be an angular unconformity, although data from this locality are insufficient to substantiate this interpretation (for further discussion see below). The Lower Jurassic (late Sinemurian) to Middle Jurassic (Bajocian) Sailor Canyon Formation (Clark and others, 1962) is made up of massive beds of volcanioclastic graywacke, siltstone, and mudstone. The contact between the Sailor Canyon Formation and the underlying limestone is
Figure 10. Bedded black nodular chert of the Elwell Formation, Bowman Lake area.

Figure 11. Folded chert layers in the Taylor Formation, Bowman Lake area.
Figure 12. Chert-pebble conglomerate of the Peale Formation, Bowman Lake area.

Figure 13. Folded bedded chert of the Peale Formation, Cisco Butte area.
disconformable. The depositional contact between the Jurassic rocks and the Peale Formation at this locality is apparently an angular unconformity.

3. North Fork of the American River

Although this area (Plate 3) has received a relatively large amount of attention from geologists over the past 20 years, interpretations of the geology of this area vary considerably. For example, early workers (Clark and others, 1962; Clark, 1976) did not recognize the upper Paleozoic volcanogenic rocks and chert (Sierra Buttes, Taylor, and Peale Formations) that have been observed by more recent studies (D'Allura and others, 1977; Harwood, 1980, 1983). Limited exposure, precipitous canyon walls, and poor accessibility have no doubt hindered attempts to map this area without relying heavily upon interpretation.

Even though there are discrepancies between much of the published data for this area, the salient aspects of the stratigraphy can be ascertained with some degree of certainty. Deformation during Permian or Triassic time resulted in the pronounced angular unconformity between the Permian Peale Formation and the overlying Late Triassic limestone (D'Allura and others, 1977; Harwood, 1983; and Plate 3). The sedimentary clasts of the Shoo Fly Formation that are observed at the base of the Triassic sequence (Clark, 1976) suggest that there is not a fault contact between the Paleozoic rock units and the overlying Triassic rocks.

The contact between the Late Triassic limestone and the Early Jurassic Sailor Canyon Formation has been described in detail by Clark (1976). He notes that "the available evidence indicates that the Sailor Canyon Formation is structurally concordant with the Triassic(?) rocks, and the paleontologic record indicates but a very short time break at the unconformity" (1962, p. 19). Clark
also states that the disconformable surface, between the Sailor Canyon Formation and the underlying limestone, has a local relief of about 3 m.

4. East-west traverse

The rocks exposed along the east-west traverse, beginning in the Foothills near Foresthill and ending east of the Ellicott Bridge in the Robbs Peak Quadrangle (Figure 1, and Plates 4, 5, and 6), are comprised solely of the Ordovician Shoo Fly Formation. Common rock types are quartzose siltstone, graywacke, mudstone, quartzite, carbonate rocks, pebble conglomerate, serpentine, and chert. Volcanogenic rocks are rare.

5. Loon Lake area

Three multiply deformed rock units are exposed in the Loon Lake area (Plate 7). The lowermost unit is made up of a fining-upward, volcanoclastic turbidite sequence. Graded, convolute, and cross beds are common (Figure 14). Detrital components include quartz and large (up to 1 cm) angular to sub-rounded feldspar grains. Minor volcanic rocks are also present.

The contact between the turbidite sequence and the overlying chert sequence is gradational. The chert is characterized by a distinctive thinly bedded black nodular chert, which is identical in appearance to the Elwell Formation that crops out 50 km to the northwest (Figure 15). Siltstone, white chert, and argillite also make up a significant portion of this unit.

Structurally overlying and in fault contact with the chert sequence is a chert breccia, which consists of poorly-sorted, angular chert clasts (up to 10 cm in length) in a siliceous matrix (Figure 16), similar to the intraformational breccia and chert-pebble conglomerate of the Peale Formation (Figure 12). Black and white chert beds that crop out near the exposed base of this unit
Figure 14. Turbidite sequence of the Loon Lake area showing graded bedding, cross-bedding and horizontal laminations.
Figure 15a. Folded black nodular chert of the Elwell Formation, Bowman Lake area.

Figure 15b. Folded black nodular chert of the chert sequence, Loon Lake area. Note refolded, rootless isoclinal fold.
Figure 16. Poorly sorted chert breccia from the Loon Lake area.
suggest that this sequence is conformable with the underlying bedded chert sequence.

The age of these three units can only be inferred indirectly since fossils were not found in this area. These rocks are distinctly different from rocks of the Shoo Fly Formation, which crops out less than 10 km to the west. The multiply deformed nature of the sequence (see Chapter IV) and the presence of a relatively thick chert sequence argues against a post-Late Triassic age since Upper Triassic and younger rocks throughout the northern Sierra Nevada exhibit neither of these characteristics. Although exact stratigraphic correlation is not possible, the rocks exposed in the Loon Lake area are remarkably similar to rocks deposited during intra-arc sedimentation that occurred during the late Paleozoic throughout the northern Sierra Nevada. Likely equivalents for the turbidite unit include the Sierra Buttes and Taylor Formations. The Elwell and/or Peale Formations may represent stratigraphic equivalents to the bedded chert, siltstone, and chert breccia found near Loon Lake. Thus, a late Paleozoic age for the Loon Lake sequence seems to be reasonable.
IV. STRUCTURE AT LOCALITIES WITHIN THE STUDY AREA

A. Comment on Nomenclature

All linear and planar deformational features are designated \( L_i \) and \( S_i \) respectively, with the subscript indicating the deformational event \( (d_i) \) that formed the given features. Therefore, the axis of a fold formed during the first deformational event, \( d_1 \), would define an \( L_1 \) lineation as would the intersection of bedding \( (S_0) \) with the axial plane of the fold \( (S_1) \) or axial-plane cleavage (also \( S_1 \)). Deformational events observed locally are designated \( d_j \), whereas events of regional extent are indicated by \( D_i \).

B. Criteria for Assigning Structures to a Deformational Event

Deformational features were correlated to deformational events using three criteria: orientation, style, and overprinting relationships. The assumptions and limitations of these criteria are discussed below.

When using orientation as a criterion, it is assumed that fold axial planes and associated axial-plane cleavage have a relatively constant orientation. While this may hold for some areas it is not true in all cases. For example, conjugate and box folds generate two \( S \)-surfaces that are oblique to each other and most likely are principal planes of finite strain as well. Folding of structures during subsequent deformation limits the use of this method. Heterogeneities in a deformational event also limit the reliability of this criterion.

The deformational-style criterion assumes that fold geometry and intensity of deformation (indicated, for example, by the presence or absence of a penetrative cleavage) are similar for a deformational event. While this may be true in many cases, it is important to note that, when dealing with regional events, the style of structures formed in one event may vary and the style of different events may be the same.
Overprinting relationships are the most reliable means for determining relative age of structures. Unfortunately, it is often not possible to observe overprinting relationships. Limited exposure of deformatonal features and obliteration of older structures by younger events limit the use of this criterion.

Because all three criteria discussed have limitations, the criteria must be used jointly to establish the most likely association between deformatonal features and events. Due to lack of overprinting relationships, the assignment of a given feature to a deformatonal event may involve a large degree of uncertainty. In spite of these limitations, with enough reliable field data it is often possible to piece together a reasonably consistent picture relating deformatonal features to deformatonal events.

C. Description of Structures

1. Foothills area

The effects of at least two deformatonal events (d₁ and d₂) are recognized in the Foothills portion of the study area (Plate 4). The first set of folds (d₁) are tight to isoclinal, exhibit class 2 or class 3 fold geometries (Ramsay, 1967), and have axial-plane cleavage. The d₁ linear features lie on a north-trending vertical great circle (Figure 17). The poles to planar d₁ features (cleavage and fold axial planes) exhibit the same great-circle distribution as do poles to bedding (Figure 17); the pole to this great circle is coaxial with d₂ linear elements.

The second set of folds (d₂) fold previous (d₁) structures. Both open, concentric and tight, similar folds are observed. Axial-plane cleavage is often, but not always, developed. Planar d₂ features strike northwest, dipping steeply (Figure 12).
Figure 17. Mesoscopic structures from the Foothills area.
Local deformational event
- Pole to axial plane of fold or bedding \((S_0)\)
- Fold axis
- Intersection lineation \((S_0 \text{ w}/S_i)\)
- Pole to foliation

Equal-area, lower hemisphere projections

Foothills
2. Devil Peak area

Structures from at least two deformational events are recognized in the Devil Peak area (Plate 5 and Figure 18). Although folds associated with d1 are not observed, the fact that d1 cleavage is sub-parallel to S0 suggests isoclinal folding during d1. The d1 penetrative cleavage is much better developed than the fracture cleavage associated with d2. Most d2 folds are tight to isoclinal; folds with either dextral or sinistral asymmetry are common (Plate 5). The d2 folds fold bedding and bedding-parallel d1 cleavage (S1). As shown in Figure 18, nearly all planar features (S0, S1, and S2) in the Devil Peak area strike northwest, dipping steeply to the northeast. The d2 fold axes plunge moderately to steeply (Figure 18).

3. Robbs Peak area

Overprinting relationships indicate that structures related to at least three deformational events are present in Robbs Peak area (Plate 6). The oldest set of structures (d1) are folded by the two younger events. The intensity of deformation for the three events decreases with decreasing age. The oldest event (d1) produced isoclinal, similar-type folds and associated penetrative axial-plane cleavage. The second event (d2) produced tight to isoclinal, similar-type folds but did not form a penetrative cleavage. The last event (d3) gently folded all previous structures; there is no cleavage associated with d3.

The orientation of planar and linear features from the Robbs Peak area is shown in Figure 19. The d1 and d2 structures have a similar orientation; axial planes have a northerly strike, dipping to the east and fold axes exhibit a north-south, subhorizontal orientation. Folds from the third deformational event (d3)
Figure 18. Mesoscopic structures from the Devil Peak area.
\( d_1 \) Local deformational event
- Pole to axial plane of fold or bedding \((S_0)\)
  - Fold axis
  - Intersection lineation \((S_o w/S_i)\)
  - Pole to foliation

Equal-area, lower hemisphere projections

Devil Peak
Figure 19. Mesoscopic structures from the Robbs Peak area
Robbs Peak

- Local deformational event
- Pole to axial plane of fold or bedding ($S_0$)
- Fold axis
- Intersection lineation ($S_0$ w/s)
- Pole to foliation

Equal-area lower hemisphere projections

$N$
have northwest-striking axial planes with axes that plunge moderately to the northwest. Bedding in this area generally dips steeply to the west.

4. Loon Lake area

Structures from at least two deformational events are recognized in the Loon Lake area (Plate 7). Deformation associated with d₂ resulted in folding of d₁ folds and d₁ cleavage throughout the area (Figure 20). Both d₁ and d₂ involved tight to isoclinal folding and the formation of penetrative axial-plane cleavage. Asymmetric z-folds are common for both events (Plate 7). Class 2 and class 3 fold geometries are common for d₁; class 1C, 2, and 3 fold geometries are observed for d₂ (Figures 21-23). Cleavage associated with d₁ is usually parallel to bedding. The poles to bedding and d₁ cleavage (S₁) have a great circle distribution; the pole to this great circle is coaxial with d₁ and d₂ linear elements (Figure 24). The d₂ cleavage often forms at high angles to bedding and usually strikes northwest, dipping steeply. A few conjugate folds associated with d₂ have axial surfaces that are upright and strike northeasterly (Figures 24 and 25). Faults, which occur throughout the area (Plate 7), offset d₁ and d₂ structures. The sense of displacement along the faults has not been determined due to the complexity of this multiply deformed area.

5. North Fork of the American River

The effects of at least three deformational events are observed in the rocks along the North Fork of the American River. Only a few d₁ folds (tight, dextral, similar folds with axial-plane cleavage) were observed in this area, although penetrative, bedding-parallel cleavage (S₁) associated with this event is common in the Shoo Fly Formation and the overlying late Paleozoic rocks. The angular unconformity between Paleozoic and Late Triassic rocks, which also truncates d₁ structures (Plate 3), is related to d₂. A large megascopic fold
Figure 20a. Interference pattern resulting from superimposed folding in the turbidite sequence, Loon Lake area. Steeply plunging D₃ folds are folded about a steeply plunging D₄ fold.

Figure 20b. Same interference pattern observed in the chert sequence, Loon Lake area. Isoclinal, rootless D₃ fold (fold axis oriented 295/41) is folded about a D₄ fold (fold axis oriented 283/43).
Figure 21. This d1 fold occurs within mudstone beds of the turbidite unit in the Loon Lake area. The axial plane and fold axis are oriented 327/69S and 307/62, respectively. Layer A lies entirely within the class III field. Layer B has a more complex geometry. Although layers A and B lie within the class III field, the geometry of the fold defined by these two layers (A+B) is very near to class II. For a description of the method used to measure fold class see section A of the Appendix.
Figure 22. This d\textsubscript{1} fold occurs within mudstone beds of the turbidite unit in the Loon Lake area. The fold axis and axial plane are oriented 292/68 and 344/71 W, respectively. The T\textsuperscript{'} values for layer A indicate that this layer lies entirely within the class III field.
Figure 23. This d2 fold occurs in interbedded mudstone and siltstone in the turbidite unit of the Loon lake area. The fold axis and axial plane are oriented 260/70 and 301/73W, respectively. Layer A and layer B consist of mudstone and plot in the class II field, as does layer A+B.
Figure 24. Mesoscopic structures from the Loon Lake area.
Local deformational event

- Pole to axial plane of fold or bedding ($S_0$)
- Fold axis
- Intersection lineation ($S_0$ w/$S_1$)
- Pole to foliation

Equal-area, lower hemisphere projections

Loon Lake
Figure 25. Conjugate d$_2$ box fold in the turbidite sequence, Loon Lake area.
(d₂), which folds d₁ cleavage as well as upper Paleozoic rocks, is also truncated at the unconformable surface (Plate 3). Mesoscopic structures associated with d₂ are not observed. The above stratigraphic and structural relationships allow the following age constraints for d₁ and d₂ deformation. The deformation associated with d₁ and d₂ occurred sometime after deposition of the Early Permian Peale Formation and prior to deposition of the late Triassic Hosselkus Limestone, which is unaffected by d₁ and d₂ deformation.

The youngest deformational event, d₃, is younger than the Lower Jurassic Sailor Canyon Formation, which was folded during d₃. A penetrative d₃ cleavage is developed locally; d₃ cleavage cross-cuts d₁ and d₂ structures in the Paleozoic rocks. Folds associated with d₃ are open to tight and occasionally exhibit sinistral asymmetry.

The orientation of d₁ and d₃ mesoscopic structures is shown in Figure 26. Poles to S₀ and S₁ lie on a great circle whose pole is coaxial with L₃. S₃ is upright and strikes north to northeast. As noted above, d₂ mesoscopic structures, including cleavage, were not observed.

6. Cisco Butte area

Structures related to at least three deformational events occur in the Cisco Butte area (Plate 2). Tight d₁ folds with weakly developed axial-plane cleavage were observed; d₁ folds were not observed in the Sailor Canyon Formation. Axial planes of d₁ folds striked north to northeast, dipping steeply. Fold axes related to this event plunge moderately to the south (Figure 27).

The Lower Jurassic Sailor Canyon Formation was deformed for the first time during d₂. The effects of d₂ are evident throughout the Cisco Butte area. Penetrative d₂ cleavage is common as are asymmetric (sinistral), gentle-to-tight d₂ folds. A northeast-trending zone of very intense shearing and/or
Figure 26. Mesoscopic structures from the North Fork of the American River.
Local deformational event

- Pole to axial plane of fold or bedding ($S_0$)
- Fold axis
- Intersection lineation ($S_0$ w/$S_i$)
- Pole to foliation

Equal-area, lower hemisphere projections

North Fork
American River
Figure 27. Mesoscopic structures from the Cisco Butte area.
Local deformational event

- Pole to axial plane of fold or bedding ($S_0$)
- Fold axis
- Intersection lineation ($S_0$ w/$S_1$)
- Pole to foliation

Equal-area, lower hemisphere projections

Cisco Butte
flattening that occurs in the Taylor Formation is a result of the \( d_2 \) deformation (see SZ in Plate 2). A large megascopic \( d_2 \) s-fold, which folds Jurassic and older rocks, is also observed in the area (Plate 2). The \( d_2 \) planar features have a relatively consistent northeast strike, dipping steeply. The \( d_2 \) fold axes plunge moderately to the northeast. The attitude of bedding seems to be primarily controlled by \( d_2 \) evidenced by the fact that poles to bedding lie on a great circle whose pole is coincident with \( d_2 \) folds axes.

A northwesterly striking cleavage is the only \( d_3(?) \) structure observed; the deformational event that formed this cleavage is questionable because overprinting relationships were not observed. The \( d_3(?) \) cleavage was observed only in the Paleozoic rocks (Plate 2) and thus may be reoriented \( d_1 \) cleavage.

7. Bowman Lake area

Overprinting relationships indicate that structures from at least three deformational events are present in the Bowman Lake area (Plate 1). Similarities in morphology and orientation of \( d_1 \) and \( d_2 \) folds often make it difficult to differentiate between \( d_1 \) and \( d_2 \) structures where overprinting relationships were not observed. Tight to isoclinal, class 2 to class 3 folds characterize \( d_1 \) (Figures 28). Open to tight, class 1c to class 2 \( d_2 \) folds are common (Figures 29 and 30). Both \( d_1 \) and \( d_2 \) have associated penetrative axial-plane cleavage.

The structural style for \( d_3 \) varies markedly from \( d_1 \) and \( d_2 \). Folds for this event are gentle to tight, concentric, and often exhibit dextral asymmetry (Figure 31). The \( d_3 \) folds, which occasionally exhibit weakly developed axial-plane cleavage, refold \( d_1 \) and \( d_2 \) structures.

The orientation of mesoscopic structures for the Bowman Lake area are shown in Figure 32. Bedding has a variable orientation but for the most part dips steeply to the east. Poles to \( S_1 \) lie on a great circle whose pole is
Figure 28. This $d_1(?)$ fold occurs within a limestone member of the Shoo Fly Formation in the Bowman Lake area. The axial plane and fold axis are oriented 022/80E and 029/75, respectively. Layer A and layers A+B exhibit class II (near hinge) and class III morphology.
Figure 29. This $d_2$ fold occurs within the bedded chert of the Elwell Formation in the Bowman Lake area. The axial plane and fold axis are oriented 046/76S and 218/63, respectively. Layer C for the most part lies well within the class IC field whereas layers A+B lie within 3 different fields. Layers A–C exhibit class II (near hinge) and class IC morphology.
Figure 30. This $d_2$ fold occurs within the bedded chert of the Elwell Formation in the Bowman Lake area. The axial plane and fold axis are oriented 041/78E and 088/65, respectively. Layer A is of class II and class III morphology while layer B is of class II and class IC geometry. Layer A+B lies on or very near the line representing class II morphology.
Figure 31. This open d3 fold occurs in bedded chert of the Elwell Formation in the Bowman Lake area. The axial plane and fold axis are oriented 355/90 and 175/63, respectively. Layers A+B for the most part exhibit class IA morphology whereas layer C has a class III morphology. Layers A-C have a class IC geometry.
Figure 32. Mesoscopic structures from the Bowman Lake area.
$d_1$ Local deformational event
- Pole to axial plane of fold or bedding ($S_o$)
- Fold axis
- Intersection lineation ($S_o$ w/$S_i$)
- Pole to foliation

Equal-area, lower hemisphere projections

Bowman Lake
approximately coaxial with \( d_2 \) and \( d_3 \) fold axes. Planar structures of \( d_2 \) generally strike northeasterly, and dip steeply to the southeast. Planar \( d_3 \) structures are upright and strike northwest. As might be expected, \( d_1 \) structures show the greatest variability in orientation due to subsequent deformation; structures from the last event, \( d_3 \), have the most consistent orientation.
V. MICROFABRIC ANALYSIS

A. Microfabric Defined

The microfabric of a rock is defined by the microstructure and lattice preferred orientation of the minerals that comprise the rock (Hobbs and others, 1977). The shape, arrangement, and internal substructure (such as kink bands and deformation lamellae) of the constituent grains of a rock delineate the microstructure.

Although microfabric data can be extremely useful in determining the deformational evolution of a rock unit, it is necessary to keep in mind the limitations of these data. Heterogeneous deformation along with biased sampling (which is often unavoidable due to the small sample population and the limited exposure of rock units) can lead to erroneous conclusions concerning the microfabric of a rock unit.

B. Lattice-Preferred Orientation in Experimentally and Naturally Deformed Rocks

1. Quartz

Workers have shown that useful information concerning the strain history of quartz-bearing rocks can be obtained by studying lattice preferred-orientation of quartz (Tullis, 1977; Miller and Christie, 1981; Lister and Williams, 1979; Lister, 1974, 1981). Lattice-preferred orientation of deformed quartz aggregates can form by translation glide or syntectonic recrystallization. Experimental and theoretical studies show how preferred orientation of quartz relates to strain, strain path, and active slip systems. Experimental work indicates that the strain and active slip systems are important factors in lattice-preferred orientation of quartz. Theoretically predicted and experimentally observed quartz-preferred orientations for given slip systems and state of strain are generally in agreement (Tullis, 1977).
The preferred orientation pattern produced by translation glide depends on the slip systems active during deformation. Experimental data indicate that the various active slip systems depend strongly upon temperature, strain, and strain rate (Heard and Carter, 1968; Ave Lallemant and Carter, 1971; Tullis and others, 1973).

Studies of axial-symmetric, experimentally deformed quartz aggregates indicate the importance of temperature and strain rate upon the lattice-preferred orientation of quartz. Tullis and others (1973) note that at temperatures below 850°C and strain rates of 10^{-5}/sec, deformation lamellae (which are assumed to be parallel to active slip planes) have many orientations and a maximum of c-axes develops about the compression axis. At higher temperatures and slower strain rates only sub-basai and prismatic deformation lamellae are observed and c-axes form a small-circle girdle about the compression axis. With either an increase in temperature or a decrease in strain rate the radius of the small circle increases, ranging from 20° to 45°.

Data from experimentally and naturally deformed quartz aggregates indicate that preferred orientation of quartz is strongly influenced by strain (see Table IV and Figure 33). Figure 33 shows various preferred orientation patterns for quartz c-axes that result from different types of strain and relative strain rates. Note that in every instance the extension direction (X) coincides with a pole-free area. It is also important to note that in nearly every case the principal planes of strain are planes of symmetry. Therefore, it is often possible to determine the orientation of the principal axes of finite strain from the lattice-preferred orientation of quartz.

Work by Miller and Christie (1981) indicates that the degree of lattice-preferred orientation (i.e. the strength of the fabric) is a function of total
**Table IV.** Outline of various types of homogeneous strain.
TABLE IV

TYPES OF HOMOGENEOUS STRAIN

1) Orthorhombic strain
   a) General strain \((X > Y > Z)\)
   b) Plane strain \((X > Y = 1 > Z)\)

2) Axial symmetric strain
   a) Extension \((X > Y = Z)\)
   b) Shortening \((X = Y > Z)\)

3) Pure shear: coaxial orthorhombic strain

4) Simple shear: non-coaxial plane strain
Figure 33. Figure showing the various types of preferred orientation patterns observed for quartz for different types of strain and strain rates. Shaded portions represent areas of relatively high concentrations of c-axes. The principal axes of the strain ellipsoid ($X \gtrsim Y \gtrsim Z$) are indicated in the stereograms to the left.
strain. They note that "microfabrics in quartzites may be cautiously interpreted as qualitative indicators of strain intensity."

Fabrics produced by syntectonically recrystallized quartz during non-coaxial strain are similar to fabrics that form by translation glide under similar conditions (Green and others, 1970). It is not known whether fabrics resulting from syntectonic recrystallization are related to stress or strain. This problem can be partially overcome by considering the fact that the equilibrium grain size for syntectonically recrystallized minerals is reached after only a relatively small amount of strain (about 4% in many cases) (Ross, 1979). Therefore, the preferred orientations of syntectonic recrystallized grains may be related to the last increment of strain, not the finite strain. For small incremental strains, the principal axes of stress and strain are approximately parallel (assuming an isotropic material). Therefore, in a quartzite that has experienced a relatively large amount of non-coaxial strain, the fabric observed in the original grains would be expected to differ from the fabric formed by syntectonically recrystallized grains, because neither the principal incremental strain axes or the principal stress axes coincide with the axes of finite strain.

2. Calcite

Important information concerning the deformational history of calcite bearing rocks can be obtained from the lattice-preferred orientation of calcite. Fabrics with orthorhombic and axial symmetry are commonly observed in naturally and experimentally deformed limestones (see Figure 34). Orthorhombic fabrics, related to orthorhombic strain, are characterized by a girdle (or partial girdle) distribution of c-axes normal to the extension direction (X) with a maximum in the direction of maximum compressive finite strain (Z) (Weiss, 1954; Neumann, 1969; Friedman and Higgs, 1981). Axially symmetric
Figure 34. Commonly observed lattice preferred orientations of calcite.

The symmetry in the sample above is nearly axial (S is foliation, L is a lineation) while the pattern below is orthorhombic (after Hobbs and others, 1976).
fabrics, which are observed in S-tectonites (deformed rocks that are characterized by a planar structure) and limestones deformed in uniaxial experiments, have c-axes that form a maximum in the Z direction or, in some cases, a small circle (radius of about 25° to 30°) about the Z direction (Neumann, 1969; Wenk and others, 1973) (see Figure 34).

Friedman and Higgs (1981) note that calcite fabrics produced in experimental shear zones are symmetric with respect to the axes of finite strain. In contrast, Rutter and Rusbridger's (1977) experimental work with non-coaxially deformed marble indicates that the concentrations of calcite c-axes are oblique to the axes to finite strain. The fact that lattice-preferred orientations in naturally deformed limestones often exhibits a symmetry that is oblique to the dimensional-preferred orientation (Weiss, 1954; Rutter and Rusbridger, 1977) seems to favor Rutter and Rusbridger's work.

C. Microfabric of Samples from the Study Area

The microfabrics of eighteen samples have been obtained from samples from the study area. The methods used to measure the microfabrics are discussed in the Appendix (Sections B and C). Most of the fabrics exhibit strong preferred orientations, which made it possible to determine the orientation of the principal axes of finite strain (X > Y > Z). Five of the samples have weak fabrics that are not useful for strain analysis. These weak fabrics are presented in the Appendix (Section B).

Three quartz c-axis fabrics and one calcite c-axis fabric were obtained from samples from the Foothills area (Figures 35-38). The finite strain for one sample was determined by measuring the shape of deformed amygdules in a metabasalt (Figure 39). Textural relationships indicate that all of these fabrics formed during d1 deformation in the Foothills area (see previous chapter for
Figure 35. Sample FH-33: This quartz-rich phyllite from the Foothills east of the Melones fault zone has a well developed $d_1$ foliation and a $d_2$ lineation, which is defined by the axis of small crenulations on the foliation plane. The orientations of 100 quartz $c$-axes of pre-tectonic grains, which exhibit a moderate-to-strong dimensional-preferred orientation, have been measured. The weakly developed cross-girdle fabric with associated pole-free area indicates that $X$ is approximately oriented $175/45$ with $XY$ oriented $018/88E$. The data are contoured at 1, 2, and 3% frequency density using a counting circle of 2% area. The data are plotted with a thin section (upper) and geographic (lower) reference. For a description of the method used to measure the $c$-axes see section B of the Appendix.
Figure 36. Sample TH-5A: This quartzose phyllite that crops out in the Foothills east of the Melones fault zone has a well developed d$_1$ foliation and a d$_2$ lineation defined by crenulation cleavage. The orientations of 199 quartz c-axes from both pre-tectonic and recrystallized grains have been measured. Old grains exhibit a strong dimensional-preferred orientation. This somewhat unusual single girdle fabric is very similar to fabrics observed by Hara and others (1973) in granites from a shear zone in Japan. The foliation plane is a plane of symmetry and is parallel to the XY plane derived from the quartz fabric. X is oriented 000/20 with XY oriented 000/90. The data are contoured at 1, 2, and 2.5% frequency density using a 4% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure 37. Sample GS-42b: This quartz greywacke that crops out in the Foothills east of the Melones fault zone lacks any well defined foliation or lineation. The $d_1$ foliation and $d_2$ lineation plotted on the stereograms are taken from measurements made near the sample locality (this may explain why $S_1$ is not a plane of symmetry). The orientations of 180 quartz c-axes of pre-tectonic (and most likely a small number of syn-tectonic) grains that have a moderate-to-strong dimensional-preferred orientation have been measured. This fabric can be considered as either a cross-girdle of small-circle girdle distribution. In either case the XY plane is oriented approximately 325/90 with X oriented 145/20. The data are contoured at 0.5, 1, 1.5, and 2% frequency density using a 4% area counting circle. The data are plotted with a thin section (upper) and geographic reference.
Figure 38. Sample GS-52: This calcitic marble is from the Foothills east of the Melones fault zone. A well defined \( d_1 \) lineation and \( d_1 \) foliation are readily observable in this sample. The calcite in this sample is recrystallized and has a strong dimensional preferred orientation. The orientations of 120 calcite c-axes have been measured. The single girdle fabric with a maximum normal to the foliation plane is characteristic of calcite fabrics. The XY plane is oriented 350/80E and corresponds to the foliation plane. The lineation is parallel to the X direction and is oriented 155/50. The data are contoured at 0.5, 2, 4, and 8\% frequency density using a 2\% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure 39. Orientation of the principal axis of finite strain for sample (TH-2) from the Foothills area (Plate 4). The strain was determined by measuring the shape of deformed amygdules in a metabasalt (see Appendix, section C). The pole to d₁ cleavage in this sample is nearly parallel to the Z-axis of the finite strain ellipsoid.
Orientations:
X : 60/77
XY : 5/79SE
Cleavage : 0/75E

Magnitude of Principal Axes of Strain (Assuming ΔV=0):
X = 2.4
Y = 1.0
Z = 0.41
more details on $d_1$ Foothills deformation). The $d_1$ foliation is nearly coplanar with the XY plane of the finite strain ellipsoid (determined from the micro-fabrics) in all but one sample (Figure 37). The XY plane for all samples strikes north to north-northwest and dips steeply. The extension direction ($X$) for these samples varies; some samples have a subhorizontal extension direction while others have an extension direction that is nearly vertical. The significance of this variation will be discussed in Chapter VII. (The relationship between local deformational events ($d_1$) to regional events ($D_1$) is also discussed in Chapter VII. In the following discussion it is important to remember that $d_1$ at one locality may not be related to $d_1$ at another locality.)

The two samples from the Devil Peak area both have XY planes of strain that strike northwest, dipping steeply (Figures 40 and 41). The extension direction for these samples is also similar, plunging $60^\circ-85^\circ$ to the northwest. These samples are taken from a locality where $S_1$ and $S_2$ are coplanar (Plate 5). The fabrics may be related to either $d_1$ (which is much more penetrative than $d_2$) or $d_2$ (which isoclinally folds $S_1$ at this locality). The similarity between the fabrics suggests that they both formed during the same event, either $d_1$ or $d_2$.

Textural features suggest that the fabrics (Figures 42 and 43) from the Robbs Peak area formed during $d_1$. The fabrics from these two samples indicate that the XY plane of strain strikes north-northeast, dipping steeply. The extension direction for both samples is subhorizontal. The strain in these samples is markedly different to the strain observed in the Devil Peak area, five miles to the southwest.

The two well developed fabrics (Figures 44 and 45) from the Cisco Butte area formed during intense $d_2$ deformation, evidenced by the style and orientation of structures associated with these samples. One sample (Figure
Figure 40. Sample DP-46: This schistose stretched pebble conglomerate from the Devil Peak area has a well defined $d_1$ foliation. The $d_2$ lineation is derived from the orientation of local fold axes. There were no $d_1$ lineations at this locality. The orientation of 200 quartz c-axes of recrystallized quartz grains have been measured. The fabric consists of two girdles, one being much more strongly developed. The XY plane is approximately parallel to the foliation and is oriented 320/70S with X oriented 275/60. The data are contoured at 1, 2, 3, 4, and 5% frequency density using a counting circle of 2% area. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure 41. Sample DP-46b: This quartz-mica schist from the Devil Peak area has a well-developed foliation and lineation. The axial plane and fold axis of a small fold are parallel to the $d_2$ foliation and $d_2$ lineation, respectively. The orientations of 149 quartz $c$-axes of recrystallized grains were measured. The foliation is approximately parallel to a plane of symmetry in this orthorhombic fabric. The XY plane is oriented 315/90 with X oriented 315/65. The data are contoured at 0.5, 1, 1.5, and 2% frequency density using a 4% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Sample RP-18a: This quartzite from the Robbs Peak area has a weakly developed $d_1$ foliation. The $d_1$ lineation is from a local measurement not from the sample (this explains why $l_1$ does not lie in $S_1$). The orientations of 180 quartz $c$-axes of pre-tectonic (and a few recrystallized) grains that exhibit a strong dimensional-preferred orientation have been measured. The cross-girdle fabric indicates that $XY$ is oriented 025/70W with $X$ oriented 020/10. The data are contoured at 1, 2, and 3% frequency density using a counting circle of 2% area. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure 43. Sample RP-19: This quartzite from the Robbs Peak area has a foliation that crosscuts folded bedding. The axial plane and fold axis of the $d_1$ fold are plotted on the stereograms. The orientations of 200 $c$-axes from recrystallized quartz grains have been measured. The XY plane for this strong cross-girdle fabric is oriented 020/90, parallel to the axial plane of the observed fold. $X$ is oriented 200/25, parallel to the fold axis. The data are contoured at 1, 2, 3, 4, and 5% frequency density using a counting circle of 2% area. The data plotted with a thin section (upper) and geographic (lower) reference.
Figure 44. Sample CG-5: This quartz-biotite schist from the Cisco Butte area has a well developed $d_2$ foliation; the $d_2$ lineation is parallel to local fold axes. The orientations of 122 c-axes of pre-tectonic and syn-tectonic (?) grains, which exhibit a slight dimensional-preferred orientation, have been measured. The fabric consists of a single girdle distribution with a maxima oriented around the pole to the foliation plane. This pattern is similar to fabrics formed in uniaxial strain tests with relatively fast strain rates and/or low temperatures. The $XY$ plane is oriented 020/70W, parallel to the foliation. $X$ is oriented 005/35, approximately parallel to the local fold axes. This fabric may indicate that $X$ is approximately equal in magnitude to $Y$. If this is true, the deformation recorded in this sample is the result of flattening. The data are contoured at 1, 2, and 3% frequency density using a 4% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure 45. Sample CG-7: This quartzite from the Cisco Butte area has a well developed lineation and foliation (parallel to bedding). The $d_1$ lineation is defined by mineral streaks on the $d_1$ foliation plane. The orientations of 199 quartz c-axes of recrystallized grains have been measured. $XY$, oriented 020/50W and $X$, oriented 245/45, are parallel to the foliation and lineation, respectively. The data are contoured at 1, 2, 3, and 4% frequency density using a 1% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
44), from a zone of extreme deformation (the shear zone in Plate 2), has an extension direction that plunges shallowly to the north. The other sample (Figure 45) has an extension direction that plunges moderately to the southwest, parallel to a lineation defined by elongate mineral streaks. In both samples the XY plane is parallel to a well developed cleavage that strikes north-northeast, dipping steeply to the west. The significance of the two different extension directions for d2 at this locality is discussed in Chapter VII.

The last two fabrics (Figures 46 and 47) are from the Bowman Lake area. These fabrics formed during either d1 and d2 deformation. Similarities in style and orientation of d1 and d2 structures make it difficult to distinguish between these events. The fabric shown in Figure 47 is from a sample from an intensely deformed area. The lineation and foliation have the same orientation as the extension direction and XY plane, respectively. The subhorizontal lineation favors a d1 origin for this fabric because subhorizontal lineations are associated with d1, not d2, deformation (Figure 32).
Figure 46. Sample EG-36: This quartzite from the Bowman Lake area has no foliation or lineation. The $d_1$ or $d_2$ foliation and lineation plotted on the stereograms are from local measurements. The orientations of 100 quartz c-axes of pre-tectonic grains that exhibit a slight dimensional-preferred orientation have been measured. This fabric is similar to fabrics observed by Hara and others (1973) in a shear zone in Japan. The XY plane is oriented 040/30E with X oriented 048/6. The data are contoured at 1, 2, and 3% frequency density using a 4% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure 47. Sample EG-66: This quartzite from the Bowman Lake area has a double maxima fabric. The $d_1$ or $d_2$ foliation and lineation plotted on the stereograms are taken from local measurements. The orientations of 150 quartz $c$-axes of reccrystallized grains have been measured. The $XY$ plane for this sample probably bisects the two maxima and is thus oriented 010/80E, approximately parallel to the local foliation. The $X$ direction is oriented 015/10. The data are contoured at 0.5, 1, 2, and 3% frequency density using a 4% counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
VI. METAMORPHISM

The metamorphism of the study area has been examined to determine the relationship between metamorphism and deformational history. Mineral paragenesis for samples collected within the study area are shown in Section D of the Appendix (Tables A.I-A.VI). Sample localities are indicated on Plates 1-7. A mineral paragenesis is an assemblage of coexisting minerals that are in contact (Winker, 1976). Therefore, one thin section may have more than one mineral paragenesis (Figure 48a). Mineral parageneses for the study area have been divided into two categories: syntectonic and post-tectonic. The distinction between syntectonic versus post-tectonic mineral parageneses is based upon the dimensional preferred orientation of platy minerals; micas in syntectonic mineral parageneses are foliated whereas micas in post-tectonic mineral parageneses are generally randomly oriented. Where possible, syntectonic mineral parageneses have been assigned to a given deformational event (d). For example, in an area with two deformational events, foliated micas that are folded would be assigned to d1 (Figure 48b).

Metamorphic mineral parageneses within the study area can be assigned to one of three facies: greenschist, albite-epidote-hornfels, and hornblende-hornfels (see Appendix, section D). The paragenesis "quartz + muscovite + epidote" is characteristic of greenschist facies metamorphism (Turner, 1976). This paragenesis is found in rocks of the Foothills and Bowman Lake areas (Appendix, Tables A.IV and A.VI). In both of these areas this paragenesis is observed as syntectonic and post-tectonic.

The paragenesis "quartz + biotite + muscovite + chlorite" is also characteristic of the greenschist facies, although this paragenesis occurs in the albite-epidote-hornfels facies as well. Winkler (1976) notes that this paragenesis is
Figure 48a. Sketch showing two different mineral parageneses (from Winkler, 1976). One paragenesis consist of A, B, and C; the other paragenesis is B, C, and D. All four minerals together do not constitute a paragenesis because all the minerals are not in contact with each other.

Figure 48b. Sketches showing d1 minerals (parallel to S1) overprinted by d2 minerals (parallel to S2) (from Spry, 1969).
diagnostic of the "biotite zone" of the greenschist facies lying above the 
isoreaction-grad

(stilpnomelane + muscovite)-out/(biotite + muscovite)-in

and below the stability field of almandine. In the Devil Peak area the "quartz + 
biotite + muscovite + chlorite" paragenesis formed syntectonically, indicating 
that this area most likely experienced medium-grade (i.e. biotite zone) 
greenschist facies metamorphism. This paragenesis occurs post-tectonically in 
the Loon Lake and Devil Peak areas, probably as a result of albite-epidote- 
hornfels facies metamorphism.

Contact metamorphism of the hornblende-hornfels facies occurred at the 
Loon Lake and Robbs Peak areas evidenced by the parageneses "quartz + 
muscovite + biotite + andalusite" and "muscovite + biotite + cordierite + quartz 
+ plagioclase." Both of these parageneses are characteristic of the hornblende- 
hornfels facies (Turner, 1976). The Loon Lake and Robbs Peak areas are both 
adjacent to large intrusive bodies.

In conclusion, syntectonic mineral parageneses indicate that rocks within 
the study area were subjected to lower to middle greenschist facies meta- 
morphism. The effects of contact metamorphism of the hornblende-hornfels 
facies are observed locally, adjacent to large granitic intrusives.
VII. INTERPRETATION OF STRUCTURES

A. Introduction

At least four regionally extensive deformational events have deformed rocks in the northern Sierra Nevada (Chapter II). The correlation between regional events and the deformational events observed locally is shown in Table V. Nevadan (D3) structures were relatively easy to correlate because structures from this event occur at all the localities and are overprinted by only one phase of deformation (D4). Where overprinting relationships were not observed it was often not possible to differentiate between Nevadan (D3) and older (D1 or D2) structures. Cretaceous (D4) structures were relatively easy to correlate from one locality to another because D4 is the last event (folding all previous structures) and D4 structures have a relatively consistent orientation (axial planes and cleavage strike northwest, dipping steeply). The style and orientation of structures of the regional events from the study area is shown in Table VI and Plate 10, respectively. A discussion of the effects of the regional events in the study area follows.

B. Regional Deformation in the Study Area

1. Post-Ordovician/Silurian and pre-Late Devonian deformation (D1)

Early Paleozoic deformation in the northern Sierra Nevada is indicated by the angular discordance between the Ordovician-Silurian Shoo Fly Formation and overlying Late Devonian rocks (D'Allura and others, 1977; Varga and Moores, 1984). Structures associated with this post-Ordovician/Silurian and pre-Late Devonian event have not been identified in the study area during this investigation. Girty and Schweickert (1984) believe they observe D1 folds in the Bowman Lake area. They note that D1 folds are isoclinal, northwest-
Table V. Correlation between regional deformational events \( (D_i) \) and deformational events observed locally \( (d_i) \).
<table>
<thead>
<tr>
<th>Regional $D_i$:</th>
<th>$D_2$</th>
<th>$D_3$</th>
<th>$D_4$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Local $d_i$</td>
<td>$d_1$</td>
<td>$d_2$</td>
<td></td>
</tr>
<tr>
<td>Foothills</td>
<td></td>
<td>$d_1$</td>
<td>$d_2$</td>
</tr>
<tr>
<td>Devil Peak</td>
<td>$d_1$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Robbs Peak</td>
<td></td>
<td>$d_2$</td>
<td>$d_3$</td>
</tr>
<tr>
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<td></td>
<td>$d_1$</td>
<td>$d_2$</td>
</tr>
<tr>
<td>North Fork</td>
<td>$d_1$ $d_2$</td>
<td></td>
<td>$d_3$</td>
</tr>
<tr>
<td>Cisco Butte</td>
<td>$d_1$</td>
<td></td>
<td>$d_3$</td>
</tr>
<tr>
<td>Bowman Lake</td>
<td>$d_1$</td>
<td>$d_2$</td>
<td>$d_3$</td>
</tr>
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Table VI. Style of regional deformational events (Dt) at the seven detailed study areas. The abbreviations O, G, T, and I stand for open, gentle, tight, and isoclinal, respectively.
<table>
<thead>
<tr>
<th>Location</th>
<th>Limb appearance</th>
<th>A. P. cleavage?</th>
<th>Asymmetry</th>
<th>Fold class</th>
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<td><strong>TABLE VI</strong></td>
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<tr>
<td></td>
<td>$D_2$</td>
<td>$D_3$</td>
<td>$D_4$</td>
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<td><strong>Foothills:</strong></td>
<td></td>
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<td></td>
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<tr>
<td>Limb appearance</td>
<td>T-I</td>
<td>O-T</td>
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<td>Yes</td>
<td>Yes</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Asymmetry</td>
<td>-</td>
<td>-</td>
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<tr>
<td>Fold class</td>
<td>II and III</td>
<td>Concentric &amp; Similar</td>
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<td><strong>Devils Peak:</strong></td>
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<td>Limb appearance</td>
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<td>G-I</td>
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<td>Yes</td>
<td></td>
<td></td>
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<tr>
<td>Asymmetry</td>
<td>-</td>
<td>S &amp; Z</td>
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<td>No</td>
<td></td>
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<td>S &amp; Z</td>
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<td><strong>Loon Lakes:</strong></td>
<td></td>
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<td>T-I</td>
<td></td>
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<td>Yes</td>
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<td>Z common</td>
<td>Z common</td>
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<td>Fold class</td>
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<td>IC, II, and III</td>
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<td>O-T</td>
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<td>Yes (locally)</td>
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<td>Limb appearance</td>
<td>T</td>
<td>G-T</td>
<td></td>
<td></td>
</tr>
<tr>
<td>A. P. cleavage?</td>
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<td>Yes</td>
<td></td>
<td>Yes(?)</td>
</tr>
<tr>
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<td>-</td>
<td>S</td>
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<td>T-I</td>
<td>O-T</td>
<td>G-T</td>
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<td>S</td>
<td>Z</td>
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<td></td>
</tr>
<tr>
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<td>Similar</td>
<td>Similar</td>
<td>Concentric</td>
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trending and lack penetrative axial-plane cleavage. Girty and Schweickert (1984) believe D₁ is related to the Antler orogeny.

Structures associated with D₁ may be present in the Shoo Fly Formation throughout the study area but were not identified due to overprinting and similarities in style and orientation with later events. At least three regional deformational events followed D₁. Two of the post-D₁ events have tight to isoclinal folds and penetrative cleavage (Table VI).

2. Late Paleozoic–early Mesozoic deformation (D₂)

Structures from this event are recognized at four of the localities in the study area (Table V and Plate 10) and can be inferred to be present in the Foothills, Devil Peak, and Loon Lake areas as well. The Lower Permian Peale Formation is the youngest rock unit deformed during this event. Upper-Triassic rocks, which crop out along the North Fork of the American River, lie with angular discordance upon Paleozoic rocks deformed during D₂ (Plate 3). Thus, within the study area, the age of D₂ is constrained as Middle Permian to Upper Triassic. A depositional hiatus of regional extent that occurred in Permo-Triassic time (see Chapter III) is probably related to D₂.

A summary of the style of deformation for D₂ is listed in Table VI. Tight to isoclinal similar folds with associated axial-plane cleavage characterize this event. Folds with upright north-striking axial planes and subhorizontal fold axes are common (Plate 10). The north-striking, steeply dipping girdle distribution of fold axes may be related to non-coaxial rotational D₂ deformation or to earlier (D₁) or later (D₃ or D₄) events.

Plate 9 is a composite diagram illustrating the inferred orientation of the principal strain axes (derived from lattice and dimensional-preferred orientation) for given deformational events in the area. Due to the similar
orientation of D2 and D3 structures, microfabric data from only a few samples were correlated to a given deformational event with a relatively large degree of certainty. Both fabrics from the Robbs Peak area are most likely the result of D2. The extension direction for D2 in the Robbs Peak area is horizontal, parallel to the axes of mesoscopic folds observed in the area (Plate 10). The XY plane of strain derived from the microfabrics is parallel to D2 cleavage, which strikes north-northeast, dipping steeply to the west (Plates 6 and 10).

Mesoscopic structures associated with D2 were not identified in the Foothills and Devil Peak area even though the rocks exposed at these localities are from the same formation (Shoo Fly) that crops out in the Robbs Peak area, where D2 mesoscopic structures are observed. In the Robbs Peak area D2 and D3 structures have a similar style and orientation (Table VI and Plate VI). In the vicinity of Colfax and Forresthill (just to the east of the Foothills area on Plate 9), upper Paleozoic rocks are intensely deformed and have structures that exhibit a similar orientation to the adjacent mildly deformed Jurassic rocks (Chandra, 1961). Thus, it is likely that D2 structures are present in the Foothills and Devil Peak areas, but are indistinguishable from D3 structures, which have a similar style and orientation.

Although structures associated with D2 were not observed in the Loon Lake area, interference patterns produced from superimposed folding suggest that D2 may have deformed rocks in this area as well. The interference pattern resulting from superimposed folding (Figure 20) indicates that D3 fold axes are coaxial with D4 fold axes (Figure 49), which plunge steeply to the northwest (Plate 10). The orientation of mesoscopic folds also indicate that D3 and D4 are coaxial (Plate 10). Steeply plunging D3 fold axes can most easily be explained by deformation of previously folded rocks. This inferred folding
Figure 49. Figure from Ramsay (1967) showing the two-dimensional interference patterns produced from two successive folding events. $a_2$ and $b_2$ are the axial surface and fold axis of the second folding event, respectively. Notice the similarity of the pattern in I (where fold axes for both events are parallel and axial planes are orthogonal) and the interference patterns in Figure 20.
$\alpha$: angle between first fold axis and $b_2$

$\phi$: angle between pole to first fold axial surface and $a_2$

$g^0$: between $0^\circ$ & $90^\circ$

$90^\circ$: between $90^\circ$ & $180^\circ$

$180^\circ$: between $180^\circ$ & $270^\circ$

$270^\circ$: between $270^\circ$ & $360^\circ$
event is thought to be related to D₂.

3. Late Jurassic Nevadan deformation (D₃)

Ample evidence for a Late Jurassic deformational event exists throughout the Sierra Nevada (Schweickert and others, 1984; Taliaferro, 1942; Bateman and Clark, 1974). Within the study area the Lower Jurassic Sailor Canyon Formation is the youngest rock unit deformed during the Nevadan orogeny (D₃). The Nevadan orogeny is the first deformational event to deform the Sailor Canyon Formation.

Nevadan deformation features are present throughout the field area (Table V and Plate 10). Nevadan deformation is characterized by tight, asymmetric, similar folds with axial plane cleavage (Table VI). In the study area Nevadan structures trend north to north-northeast (Plate 10). These trends differ markedly from north-northwest trends commonly associated with the Nevadan orogeny.

Some workers in the central portion of the northern Sierra Nevada interpret the north to north-northeast trending D₃ structures as late Nevadan (Schweickert and others, 1984), the result of a last pulse of the Nevadan orogeny. They argue that "main-phase" Nevadan structures in this area have a northwest orientation (Schweickert and others, 1984). Their support for this interpretation comes from structures in the Shoo Fly Formation (Bowman Lake area) and the Lower Jurassic Sailor Canyon Formation. Data from the present work is at odds with Schweickert and others' interpretation. Planar and linear features associated with late Paleozoic–early Mesozoic (D₂) and Nevadan (D₃) deformation for the Peale and Sailor Canyon Formations are shown in Figure 50. First-phase structures (D₂) in the Peale Formation in the Bowman Lake and Cisco Butte areas have the same orientation. The D₂ structures formed prior
Figure 50. Planar and linear features associated with late Paleozoic-early Mesozoic (D2) and Nevadan (D3) deformation for the Peale and Sailor Canyon Formations from the Bowman Lake, Cisco Butte, and North Fork of the American River areas.
to Early Jurassic time and are thus not present in the Sailor Canyon Formation. Second-phase structures (D₃, based upon overprinting relationships and orientation) in the Peale Formation have a similar orientation and style to first-phase structures in the Sailor Canyon Formation. The Nevadan (D₃) is the only event recognized in the Sailor Canyon Formation in the Cisco Butte–North Fork area. It seems reasonable to attribute this event to the "main phase" of the Nevadan orogeny, not to a late pulse of the Nevadan orogeny as Schweickert and others (1984) suggest.

Significant reorientation of Nevadan structures during the Cretaceous deformational event (D₄) is observed in the Foothills and Loon Lake areas. Poles to Nevadan planar features in both areas have a great circle distribution; the poles to these great circles coincide with D₄ fold axes for the respective areas (Plate 10). Data from the Foothills are further complicated by the similar orientation of D₂ and D₃ structures (see previous section). Most D₃ planar features in the Foothills, which are upright and strike north to northeast, have a similar orientation to D₃ structures observed elsewhere in the study area (Plate 10). Interference patterns resulting from superimposed folds in the Loon Lake area indicate that D₃ cleavage was upright and northeast-striking prior to D₄ folding (see previous section and Figure 23).

Linear features associated with D₃ generally plunge moderately to steeply. The D₃ linear features usually exhibit a more shallow plunge in Jurassic rocks and plunge more steeply in Paleozoic rocks. This is no doubt due to the fact that Paleozoic rocks experienced at least one deformational event prior to D₃. The subhorizontal D₃ fold axes in the Robbs Peak area are also a result of pre-D₃ deformation. Lineations in this area occur in the hinge zone of subhorizontal D₂ folds (Plate 10).
Only two microfabrics, both from the Cisco Butte area, are correlated to D₃ with a high degree of certainty (Plate 9). Most of the other microfabrics may be the result of D₃, but age constraints necessary to differentiate between D₃ and other events (D₂ or D₄) are lacking (Plate 9). The two microfabrics from the Cisco Butte area have markedly different extension directions. One sample (CG-5), from a zone of very intense deformation (the shear zone of Plate 2), has a subhorizontal extension direction. The other sample (CG-7) has an extension direction that plunges steeply within the XY plane of strain. In both samples the XY plane of strain is parallel to well developed D₃ cleavage. The markedly different fabrics are probably the result of non-coaxial Nevadan deformation. Because fabrics with subhorizontal and more steeply plunging extension directions are associated with Nevadan deformation, it is not possible to differentiate between Nevadan and D₂ based upon the orientation of extension directions.

4. Cretaceous deformation (D₄)

Nevadan and older structures throughout the study area are folded about moderate-to-steeply plunging D₄ folds (Plate 10 and Figure 51). A late Early to early Late Cretaceous age for D₄ is supported by regional considerations (Chapter II). Cretaceous deformation, which is the youngest event observed in the study area, is characterized by gentle, concentric folds with dextral asymmetry (Table VI). As a rule, Cretaceous deformation is not nearly as penetrative as the preceding events (D₂ and D₃).

Within the study area, D₄ linear features exhibit a variable orientation but have a relatively constant orientation at any given locality (Plate 10). The variable orientation of Cretaceous linear features is due to the varied attitude of planar surfaces prior to Cretaceous deformation. Cretaceous planar features
Figure 51. \(D_4\) folding in the Loon Lake (above) and Robbs Peak areas.
exhibit a consistent northwest trend throughout the study area.

The anomalous north to north-northeast trend of Nevadan structures in the study area is best explained by post-Nevadan oroclinal folding in the northern Sierra Nevada. The oroclinal folding occurred prior to or during the Cretaceous deformational event (D₄). The relatively consistent northwest trend of D₄ structures indicates that oroclinal folding did not postdate D₄.

Post-Nevadan oroclinal folding in the northern Sierra Nevada is supported by the following:

1) The large Z-fold defined by formation contacts and the faults of the Foothills fault system (Figure 7, 8 and 52).

2) Foliation associated with D₂ and D₃ that are folded in a similar manner to the trend of the Foothills fault system. Mesoscopic data near the hinges of the large Z-fold illustrate the reorientation that has most likely occurred (localities 1, 2, and 2a of Figure 3; localities 1, 2, and 6a of Figure 4).

3) Paleomagnetic data from the northern Sierra Nevada that support dextral rotation during Cretaceous time (Figure 9; see Chapter II for discussion).

4) Relative plate motions for central California, which indicate that oblique right-lateral convergence has occurred adjacent to the Sierra Nevada since the Early Cretaceous (120 m.y.B.P.) (Page and Engebretson, 1984).

C. Tectonic Significance of Regional Deformations

Post-Ordovician/Silurian and pre-Late Devonian deformation (D₁) within the Shoo Fly Formation most likely occurred outboard from the North American continent. Paleogeographic data for the southern Cordilleran (cf. Chapter II) favor the idea that rocks of the northern Sierra Nevada were accreted onto the margin of western North America during Permo-Triassic time (Dickinson, 1977;
Figure 52. Petrotectonic terranes of the northern Sierra Nevada. The numerous faults that occur throughout the mixed terrane make up the Foothills fault system.
Speed, 1979). If the northern Sierra Nevada was outboard from North America prior to Permo-Triassic time, the correlation Schweickert and Girty (1984) make between \( D_1 \) and Antler deformation is incorrect.

Late Paleozoic-early Mesozoic deformation \( (D_2) \) in the northern Sierra Nevada is, at least in part, the result of the Sonoma orogeny. The \( D_2 \) structures formed in association with subduction and accretion that occurred along the margin of North America during Permian through Early Jurassic time. The north-northwest strike of \( D_2 \) structures in the Sierra Nevada (Figure 3) exhibit approximately the same trend as the western boundary of the southern Cordilleran following accretion of allochthonous material during the Early to Middle Triassic (Burchfiel and Davis, 1972; Speed, 1978, 1979). Sonoman structures in Nevada have a northeast trend, similar to the strike of Paleozoic structural and geosynclinal trends for the southern Cordilleran (Burchfiel and Davis, 1972; Speed, 1979; Oldow and others, 1984). The marked difference in the configuration of the boundary of North America from northeast in the Paleozoic to northwest in the Mesozoic is apparently the result of accretion of an exotic terrane, Sonomia (Speed, 1979; cf. Chapter II).

The \( D_2 \) structures of the northern Sierra Nevada are apparently related to southwest-northeast crustal shortening which, along with paleogeographic data (Speed, 1979), indicate that a convergent plate boundary existed along the western edge of Sonomia prior to accretion onto North America. Speed (1979) believes that the convergent boundary had a large component of left-lateral strike-slip. The \( D_2 \) structures in the northern Sierra Nevada appear to be related to the normal, not the strike-slip, component of convergence since structures strike at 90 degrees to the assumed normal component of convergence (Figure 53).
Figure 53. Plate kinematic model for Early to Middle Triassic time.
EARLY TO MIDDLE TRIASSIC

Approximate (?) Convergence Direction

Calif., Nevada

Trend of D₂ Structures

Trench

Trend of Paleozoic Paleogeographic Belts

0 50 MILES

38° 39° 40°
The Late Jurassic Nevadan orogeny (D3) most likely records the effects of an arc-continent collisional event (Schweickert and Cowan, 1975; cf. Chapter II). This event is believed responsible for the present distribution of petrotectonic terranes (Figure 52) in the northern Sierra Nevada, where two Jurassic arc complexes (one exposed within the the eastern belt, the other within the western belt of Figure 52) are separated by a chaotic, mixed terrane. The numerous faults that occur throughout the mixed terrane make up the Foothills fault system.

The mixed terrane appears to be the remnant of a Permian to Late Jurassic accretionary prism. Evidence supporting this interpretation includes the occurrence of blueschist and ultramafic rocks as well as the overall chaotic structure of this terrane. Blueschists crop out along the eastern margin of the mixed terrane, near the town of Downieville (Schweickert and others, 1980). K-Ar age determinations on lawsonite-bearing blueschists from this area indicate that high-pressure metamorphism occurred prior to 174 m.y.B.P. (Schweickert and others, 1980). Schweickert and others (1980) note that blueschists from this locality "provide the first direct evidence for pre-Franciscan subduction in the Sierra Nevada region." Ultramafic rocks, which crop out throughout this terrane, generally occur as long, thin, fault-bounded slabs; larger ultramafic bodies often appear to lie at the base of stratigraphic sequences (Bateman and Clark, 1974). Hietanen (1981) notes that elongate ultramafic bodies of the Horseshoe Bend Formation, which is exposed in the northern portion of the mixed terrane, most likely represent slabs of oceanic crust and mantle that were deformed and displaced during Permian to Jurassic subduction. The occurrence of numerous faults, as well as melange, further supports a subduction complex origin for the mixed terrane. Duffield and Sharp (1975)
describe a 4 km-wide melange belt in the southern portion of this terrane. Melange, described by Hietanen (1981), also occurs in the northern portion of this terrane. The fact that melange, ophiolitic fragments and blueschists, all characteristic of subduction complexes, occur in the mixed terrane lends strong support to an accretionary-prism origin for this terrane.

The dominant trend of Nevadan structures throughout the Sierra Nevada is northwest (Figure 4), although some reorientation of structures during post-Nevadan deformation has occurred. The well-developed northwest-striking Nevadan structures are the result of northeast-southwest crustal shortening associated with the accretion of the allochthonous western belt arc-related rocks (Figure 54). Examination of extension directions for the Nevadan orogeny (see section B, this chapter) indicate that Nevadan deformation is, at least locally, non-coaxial. The subhorizontal extension direction in the shear zone in the Cisco Butte area may be the result of strike-slip deformation. This seems to imply that a tangential component of convergence accompanied Nevadan deformation. Page and Engebretson's (1984) work, in which hotspots are used to model the relative motion between the Pacific and Farallon plates, indicates that there was a significant left-lateral component of relative plate motion during the Late Jurassic.

Following the intense Nevadan deformation, a set of northeast-striking structures formed in the southern portion of the western metamorphic belt (Figure 5). The northeast-trending structures are latest Jurassic to Early Cretaceous in age. Structures associated with this post-Nevadan event most likely formed as a result of northwest-southeast crustal shortening. Northwest-southeast, post-Nevadan back-arc thrusting in western Nevada supports crustal shortening in a similar direction (Oldow and others, 1984). The abrupt change
Figure 54. Plate kinematic model for Late Jurassic time.
LATE JURASSIC
from northeast-southwest Nevadan crustal shortening to northwest-southeast post-Nevadan shortening is best explained by left-lateral convergence following the Nevadan collisional event.

During late Early to early Late Cretaceous time a northwest-striking set of structures formed in the Sierra Nevada (Figure 5) in response to northeast-southwest crustal shortening (Figure 55). Cretaceous crustal shortening (D4) is related to Franciscan subduction (Figure 55). By 120 m.y.B.P. the left-lateral component of convergence was apparently replaced with a right-lateral component (Page and Engebretson, 1984). The right-lateral component of Franciscan subduction probably caused the oroclinal Z-fold in the northern Sierra Nevada. The post-Nevadan oroclinal folding occurred prior to or during D4.
Figure 55. Plate kinematic model for the latest Cretaceous.
LATEST CRETACEOUS
VIII. SUMMARY AND CONCLUSIONS

Four regionally extensive deformational events have affected the rocks of the northern Sierra Nevada. The kinematic significance of D₁, which deforms the Ordovician-Silurian Shoo Fly Formation but not the overlying Late Devonian rocks, is obscure due to the apparent allochthonous nature of Paleozoic rocks of the northern Sierra Nevada. Structures associated with D₁ were not identified due to overprinting of later deformational events.

There is ample evidence supporting regionally extensive late Paleozoic-early Mesozoic (D₂) deformation in the Sierra Nevada. The D₂ structures are the result of compressive deformation along a northwest-striking, east-dipping convergent plate boundary along the western margin of the Sierran province (Figure 53) and may be related to the accretion of an exotic terrane, Sonomia. Although a significant component of strike-slip motion may have existed along the plate boundary, D₂ structures appear to be related to the normal component of convergence.

The intense, short-lived Nevadan orogeny (D₃) may be the result of an arc-continent collision (Figure 54). This latest Jurassic event is responsible for the present distribution of petrolectonic terranes in the northern Sierra Nevada, where two Jurassic arc complexes (an eastern Andean arc and western island arc) are separated by a chaotic, mixed terrane (Figure 52). Left-lateral oblique convergence occurred during and after the Nevadan orogeny.

The last event (D₄), of Cretaceous age, folds all previous structures. Cretaceous structures have a consistent northwest trend throughout the northern Sierra Nevada (Figure 55). Dextral oroclinal folding in the northern Sierra Nevada, which occurred prior to or during D₄, is the result of right-lateral oblique convergence.
Three main conclusions from the present work are:

1) Nevadan (D₃) structures are often indistinguishable from late Paleozoic–early Mesozoic (D₂) structures due to similarities in style and orientation.

2) Deformation associated with the Nevadan orogeny is, at least locally, non-coaxial. The non-coaxial deformation is most likely related to left-lateral oblique convergence.

3) The anomalous north to north-northeast trend of Nevadan structures in the study area is best explained by post-Nevadan oroclinal folding in the northern Sierra Nevada. The oroclinal folding occurred prior to or during the Cretaceous deformational event (D₄).
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APPENDIX

A. Classification of Fold Geometries

The excellent rock exposures in the Bowman Lake and Loon Lake areas enabled detailed examination of fold geometries to be made. Profile thickness (T) of folded layers parallel to the trace of the axial surface were measured from photos of the fold profiles. Measurements were normalized by recording the ratio (T') of the thickness on the limbs of the fold (Tα, where α is the angle that the tangent to the folded layer makes with a plane perpendicular to the axial plane of the fold) with the thickness of the hinge of the fold (T0) (for further discussion see Ramsay, 1967, p. 359).

A graph plotting values of T' versus α can be divided into fields which represent the various classes of fold geometries (Ramsay, 1962). Figure A.1 shows a T' versus α graph with the corresponding fold classes indicated. Class IB (concentric or parallel) and class II (similar) folds plot along curves which separate the less restricted fields of the other fold classes.

Graphs plotting T' versus α for folds in the Bowman Lake and Loon Lake areas, along with a sketch of the fold profiles, are shown in Figure A.2 to A.5 and in Chapter IV. For comments relating to the individual folds examined refer to the figure captions. For information concerning the nature of the deformational events (d1) that formed the given folds see Chapter IV.
Figure A.1. Graph of $T'$ versus $\alpha$ depicting various classes of fold geometries (after Ramsey, 1967).
Figure A.2. This d_2 fold occurs within a chert bed of the Taylor Formation in the Bowman Lake area. The axial plane and fold axis are oriented 057/56° and 139/52, respectively. The fold lies entirely within the class III field.
Figure A.3. This d2 fold occurs within the cherty beds of the Taylor Formation in the Bowman Lake area. The axial plane and fold axis are oriented 041/76S and 141/76, respectively. Layer A for the most part lies well within the class IC field. Layer B exhibits mostly a class II morphology. Layers C+D and layers A–D exhibit complex fold geometries.
Figure A.4. This d_2 fold occurs within siltstone and mudstone beds of the Taylor Formation in the Bowman Lake area. The axial plane and fold axes are oriented 034/68E and 140/76, respectively. Layers A-D are made up primarily of siltstone; layer E consists of less competent mudstone. Layer B, layers B-D, and layers A-E all plot within the class IC field. The mudstone bed lies entirely within the class III field.
Figure A.5. This gentle to open d3 fold occurs in the bedded chert of the Elwell Formation in the Bowman Lake area. The axial plane and fold axis are oriented 311/90 and 130/55, respectively. Layers A and B are complexly folded. Layers A+B lie in the class IC field.
B. Lattice-Preferred Orientation Data

The lattice-preferred orientation of 17 samples located within the study area have been determined (Plate 1-7 and Plate 9). Samples were prepared by cutting one thin section approximately normal to the dominant s-surfaces (cleavage and/or bedding) of the sample. Between 100 and 200 measurements were made for each sample. These data were then plotted on equal-area, lower hemisphere nets and contoured. The contouring values indicate frequency densities. The frequency density ($d$) is defined as follows:

$$d = n/NA$$

where $A$ is the area of the counting circle, $N$ is the sample size (population), and $n$ is the number of points which fall within the counting circle (Mockel, 1969). Mockel (1969, p. 97) notes that by contouring frequency densities, "contours possess absolute values indicating frequency density, being expressed in a measure that is independent both of total number of measurements and the area of the counting circle used. Therefore, ... there is no need to express contour values in percentages per percentage area."

Counting and plotting of the data were accomplished by using Warner's (1969) FORTRAN computer program. The program has been modified to plot data in an equal-area projection. The rotation subroutine (rote) was rewritten to correct for errors in the original program. It should be noted that Warner's "defilm" value, which determines the size of the counting circle, is incorrect. Correct values for defilm for counting circles of 1, 2, 3, 4, and 5% area are 0.1415, 0.2003, 0.2456, 0.2838, and 0.3176, respectively. The program was adapted to run on the IBM 360/370 computer at Rice University by Lambert (1979).
The size of the counting circle, along with the sample size and contour interval, is given for each sample (see figure captions). Fluctuations in the data that are smaller than the counting circle used to count the data are not observable. By using a small counting circle, random fluctuations may result in a biased distribution. Ideally one should choose a counting circle that is small enough to detect meaningful features and large enough to eliminate undesirable noise. As a general rule, for samples with a relatively random distribution of fabric elements and/or a small sample size (population), a larger counting circle should be used.

All the data are plotted on equal-area, lower hemisphere projections. The local lineation and foliation are plotted along with the microfabric data. In most cases two plots are shown; one with the foliation striking "north," dipping steeply and the other with the actual orientation of the fabric elements. Occasionally, when a sample lacked strong planar or linear features, only the geographic plot of the data is shown. Where possible, the orientation of the axes of the finite strain ellipsoid (X > Y > Z) have been determined from the microfabric data. Figures of most of the microfabric plots are presented in Chapter V. A few of the weaker fabrics are included here (Figures A.6-A.10).
Figure A.6. Sample TH-3: This quartzose phyllite which crops out in the Foothills east of the Melones fault zone has a well developed d_1 foliation and a d_2 lineation defined by crenulation cleavage. The orientations of 120 quartz c-axes of pre-tectonic grains that exhibit a noticeable preferred orientation have been measured. The data are contoured at 1 and 2% frequency density using a 2% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure A.7. Sample DP-44a: This quartz muscovite schist, which crops out along the Rubicon River in the Devil Peak area, has a well developed $d_1$ foliation. The $d_2$ lineation (defined by fold axes) is taken from measurements made near the sample locality. The orientations of 192 quartz $c$-axes of pre-tectonic and recrystallized grains have been measured. The data are contoured at 0.5, 1, 1.5, 2, and 2.5% frequency density using a 4% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure A.8. Sample W-37: This recrystallized nodular chert from the Loon Lake area has a weakly developed \( d_2 \) foliation parallel to bedding. The plotted lineation is parallel to a local \( d_2 \) fold axis. The orientations of 100 quartz \( c \)-axes have been measured. The data are contoured at 1, 2, and 3% frequency density using a 2% area counting circle. The data are plotted with a thin section (upper) and geographic (lower) reference.
Figure A.9. Sample EG-22: This quartzite from the Bowman Lake area has a weakly developed $d_1$ foliation parallel to bedding. The lineation is a local $d_1$ fold axis. The orientations of 199 quartz c-axes of recrystallized grains have been measured. The weakly developed small-circle girdle may have formed about the pole to XY. If so, XY is oriented 285/50S with X oriented 170/50, have approximately parallel to the regional fold axes of the area. The data are contoured at 1, 1.5, 2, and 2.5% frequency density using a counting net of 4% area. The data are plotted with a geographic reference.
Figure A.10. Sample EG-49: This quartz sandstone from the Bowman Lake area has a weakly developed fabric. The orientation of bedding is plotted on the stereogram. The orientations of 199 quartz c-axes of pre-tectonic grains which exhibit a slight dimensional-preferred orientation have been measured. The data are contoured at 0.5, 1, 1.5, and 2% frequency density using a counting circle of 4% area. The data are plotted with a geographic reference.
C. Determination of Finite Strain

The finite strain ellipsoid for one sample (TH-2) was determined from the shape of deformed amygdules in a metabasalt. The amygdules in this sample are filled with secondary calcite. Determination of the finite strain ellipsoid was accomplished by determining the state of strain on three orthogonal planes. On each plane the ratio of the major to minor axis of the strain ellipse, as well as the direction of the major axis, was determined from the mean value of 50 measurements. These data were used to construct the strain tensor for the given reference coordinate system. The principal strains and directions were obtained by calculating the eigenvalues for the strain tensor.

The results are plotted in Figure 39 (Chapter V). It should be noted that ratios and not absolute values were measured for the strain ellipses, thus making it impossible to determine the absolute values of the principal axes of the strain ellipsoid without further information. Considering the nature of the protolith, post-formational volume loss (ΔV) is most likely negligible. By assuming ΔV = 0, the absolute values of the principal axes can be calculated. It is worthwhile to note that the pole to the cleavage in this sample and the Z axis of the strain ellipsoid plot within 7° of each other. This sample has apparently experienced a relatively large amount of plane strain.
D. Mineral Parageneses

Mineral parageneses for samples collected within the study area are shown in Tables A.1-A.6. Sample localities are indicated on Plates 1-7. See Chapter VI for further discussion.
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### TABLE A.11 MINERAL PARAGENESIS—ROBB'S PEAK

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**SYNTECTONIC MINERAL PARAGENESIS**

**POST-TECTONIC MINERAL PARAGENESIS**

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LEGEND

**MESOZOIC (?)**
- **Mui** Undifferentiated Intrusive Rocks of Probable Mesozoic Age.

**MISSISSIPPIAN TO PERMIAN**
- **Mp** PEALE FORMATION: Chert, Intraformational Breccia, and Siltstone.

**DEVONIAN**
- **Dt** TAYLOR FORMATION: Volcaniclastics and Pillowed Flows.
- **De** ELWELL FORMATION: Black Phosphatic Chert.
- **Ds** SIERRA BUTTES FORMATION: Dacitic Flows and Intrusives.

**PRE-DEVONIAN**
- **Ss** SHOO FLY FORMATION: Siliciclastics with Some Chert (c) and Limestone (l) Members.

*FOR EXPLANATION OF SYMBOLS SEE PLATE 8*
**LEGEND**

- **bi**: Basic Intrusive Rocks.
- **JSc**: SAILOR CANYON FORMATION: Volcaniclastic Graywacke, Siltstone, and Mudstone.
- **Rh**: HOSELKUS LIMESTONE.
- **Mp**: PEALE FORMATION: Bedded Chert, Siltstone, and Chert Breccia.
- **Dt**: TAYLOR FORMATION: Bedded Volcaniclastics with Chert-Rich Member (c) and "Shear Zone" (SZ) Indicated.

*FOR EXPLANATION OF SYMBOLS SEE PLATE 8*
LEGEND*

- **bi** Basic Intrusive Rocks.
- **Jsc** SAILOR CANYON FORMATION: Volcaniclastic Graywacke, Siltstone, and Mudstone.
- **Trh** HOSSELSKUS LIMESTONE.
- **Mp** MISSISSIPPIAN-PERMIAN: Bedded Chert, Siltstone, and Chert Breccia.
- **Dt** DEVONIAN: Bedded Volcaniclastics with Chert-Rich Member (c) and "Shear Zone" (SZ) Indicated.

*FOR EXPLANATION OF SYMBOLS SEE PLATE 8

METERS

PLATE 2
120° 32' 55"
39° 17' 49"
**LEGEND**

- **Jmp** METAMORPHOSED PYROCLASTICS ROCKS: Mafic to Intermediate Pyroclastic Rocks.
- **Jsc** SAILOR CANYON FORMATION: Andesitic Volcanioclastics.
- **Rh** HOSSELKUS LIMESTONE: Grey Limestone.
- **Mp** PEALE FORMATION: Chert Breccia, Fine Clastics and Chert.
- **Dt** TAYLOR FORMATION: Andesitic Volcanioclastics.
- **De** ELWELL FORMATION: Black Argillites, Slates and Cherts.
- **Ds** SIERRA BUTTES FORMATION: Dacitic Volcanioclastics
- **Ss** SHOO FLY FORMATION: Grey to Black Cherts, Slates, and Wackes.

*After D'All and Harw.
**For Expla See Plate
GEOLOGY OF THE NORTH FORK AREA

*After D'Allura (1977) and Harwood (1980).

**For Explanation of Symbols, See Plate 8.
AREA*

* For Explanation of Symbols
See Plate 8. See Chapter III
For Rock Descriptions.

PLATE 4
LEGEND*

MESOZOIC


UPPER PALEOZOIC (?)

[Pcb]  Chert Breccia and Bedded Chert.


[Pt c]  Volcaniclastic Turbidites with Some Chert (c) Beds.

*FOR EXPLANATION OF SYMBOLS SEE PLATE 8

Kilometers

120° 22′ 01″
38° 59′ 17″
EXPLANATION OF SYMBOLS

--- --- Contact, Dashed where approximate Fault, Dashed where Approximate.

\( \angle ^{60} \) Strike and Dip of Bedding.

\( \angle ^{80} \) Strike and Dip of Overturned Bedding.

\( \angle \) Strike of Vertical Bedding.

\( \angle ^{70} \) Strike and Dip of \( d_1 \) Foliation.

\( \angle \) Strike of Vertical \( d_1 \) Foliation.

\( \angle ^{55} \) Strike and Dip of \( d_2 \) Foliation.

\( \angle ^{35} \) Strike and Dip of \( d_3 \) Foliation.

\( \angle ^{65} \) Trend and Plunge of \( d_1 \) Lineation.

\( \angle ^{43} \) Trend and Plunge of \( d_2 \) Lineation.

\( \angle ^{81} \) Trend and Plunge of \( d_3 \) Lineation.

\( ^{47} \) Orientation of \( d_1 \) Fold Axis and Axial Plane.

\( ^{63} \) Orientation of \( d_1 \) Fold Axis and Axial Plane Showing Clockwise (above) and Counter-Axisymmetry.

\( ^{51} \) Orientation of \( d_2 \) Fold Axis and Axial Plane.

\( ^{72} \) Orientation of \( d_3 \) Fold Axis and Axial Plane.
PLANATION OF SYMBOLS*

--- Contact, Dashed where approximate

--- Fault, Dashed where Approximate.

\[ \angle 60 \] Strike and Dip of Bedding.

\[ \angle 80 \] Strike and Dip of Overturned Bedding.

\[ \angle 70 \] Strike of Vertical Bedding.

\[ \angle 70 \] Strike and Dip of \( d_1 \) Foliation.

\[ \angle 55 \] Strike of Vertical \( d_1 \) Foliation.

\[ \angle 55 \] Strike and Dip of \( d_2 \) Foliation.

\[ \angle 35 \] Strike and Dip of \( d_3 \) Foliation.

\[ \rightarrow 65 \] Trend and Plunge of \( d_1 \) Lineation.

\[ \rightarrow 43 \] Trend and Plunge of \( d_2 \) Lineation.

\[ \rightarrow 81 \] Trend and Plunge of \( d_3 \) Lineation.

\[ 47 \rightarrow 63 \] Orientation of \( d_1 \) Fold Axis and Axial Plane.

\[ 51 \rightarrow 72 \] Orientation of \( d_1 \) Fold Axis and Axial Plane Showing Clockwise (above) and Counter-Clockwise Asymmetry.

\[ 51 \rightarrow 72 \] Orientation of \( d_2 \) Fold Axis and Axial Plane.

\[ 39 \rightarrow 48 \] Orientation of \( d_3 \) Fold Axis and Axial Plane.
Trend and Plunge of $d_2$ Lineation.
Trend and Plunge of $d_3$ Lineation.
Orientation of $d_1$ Fold Axis and Axial Plan Showing Clockwise (above) and Counter-Asymmetry.
Orientation of $d_2$ Fold Axis and Axial Plan
Orientation of $d_3$ Fold Axis and Axial Plan
Petrofabric Sample Locality with Sample Number Indicated.
Metamorphic Facies Sample Locality.
Petrofabric and Metamorphic Facies Sample Locality with Sample Number Indicated.

*For Def See Ch
Trend and Plunge of $d_2$ Lineation.

Trend and Plunge of $d_3$ Lineation.

Orientation of $d_1$ Fold Axis and Axial Plane.

Orientation of $d_1$ Fold Axis and Axial Plane Showing Clockwise (above) and Counter-Clockwise Asymmetry.

Orientation of $d_2$ Fold Axis and Axial Plane.

Orientation of $d_3$ Fold Axis and Axial Plane.

Petrofabric Sample Locality with Sample Number Indicated.

Metamorphic Facies Sample Locality.

Petrofabric and Metamorphic Facies Sample Locality with Sample Number Indicated.

*For Definition of $d_i$ See Chapter IV.

PLATE 8
Microfabric Data

EXPLANATION

Equal—Area, Lower Hemisphere Projection of the XY Plane of the Strain Ellipsoid based upon Microfabric Data. The X-Direction, sample number, and Regional Deformational Event ($D_1$) are indicated.
EXPLANATION

Equal-Area, Lower Hemisphere Projection of the XY Plane of the Strain Ellipsoid based upon Microfabric Data. The X-Direction, sample number, and Regional Deformational Event (D_i) are indicated.

PLATE 9
ORIENTATION OF REGIONAL DEFORMATIONAL FEATURES

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OTATION OF REGIONAL
RFMATIONAL FEATURES*