INFORMATION TO USERS

This reproduction was made from a copy of a manuscript sent to us for publication and microfilming. While the most advanced technology has been used to photograph and reproduce this manuscript, the quality of the reproduction is heavily dependent upon the quality of the material submitted. Pages in any manuscript may have indistinct print. In all cases the best available copy has been filmed.

The following explanation of techniques is provided to help clarify notations which may appear on this reproduction.

1. Manuscripts may not always be complete. When it is not possible to obtain missing pages, a note appears to indicate this.

2. When copyrighted materials are removed from the manuscript, a note appears to indicate this.

3. Oversize materials (maps, drawings, and charts) are photographed by sectioning the original, beginning at the upper left hand corner and continuing from left to right in equal sections with small overlaps. Each oversize page is also filmed as one exposure and is available, for an additional charge, as a standard 35mm slide or in black and white paper format.*

4. Most photographs reproduce acceptably on positive microfilm or microfiche but lack clarity on xerographic copies made from the microfilm. For an additional charge, all photographs are available in black and white standard 35mm slide format.*

*For more information about black and white slides or enlarged paper reproductions, please contact the Dissertations Customer Services Department.

UMI Dissertation Information Service
University Microfilms International
A Bell & Howell Information Company
300 N. Zeeb Road, Ann Arbor, Michigan 48106
Mirkin, Andrew Stephen

STRUCTURAL ANALYSIS OF THE EAST EAGLE CREEK AREA, SOUTHERN WALLOWA MOUNTAINS; NORTHEASTERN OREGON

Rice University M.A. 1986

University Microfilms International 300 N. Zeeb Road, Ann Arbor, MI 48106
PLEASE NOTE:

In all cases this material has been filmed in the best possible way from the available copy. Problems encountered with this document have been identified here with a check mark ✓.

1. Glossy photographs or pages ✓
2. Colored illustrations, paper or print ✓
3. Photographs with dark background ✓
4. Illustrations are poor copy •
5. Pages with black marks, not original copy •
6. Print shows through as there is text on both sides of page •
7. Indistinct, broken or small print on several pages ✓
8. Print exceeds margin requirements •
9. Tightly bound copy with print lost in spine •
10. Computer printout pages with indistinct print •
11. Page(s) lacking when material received, and not available from school or author.
12. Page(s) seem to be missing in numbering only as text follows.
13. Two pages numbered ■. Text follows.
14. Curling and wrinkled pages •
15. Dissertation contains pages with print at a slant, filmed as received •
16. Other

________________________________________

________________________________________

University
Microfilms
International
RICE UNIVERSITY

STRUCTURAL ANALYSIS OF THE EAST EAGLE CREEK AREA, SOUTHERN WALLOWA MOUNTAINS; NORTHEASTERN OREGON

by

ANDREW S. MIRKIN

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

MASTER OF ARTS

APPROVED, THESIS COMMITTEE:

[Signatures]

Dr. Hans G. Avé Lallemant,
Professor of Geology, Director

Dr. John S. Oldow,
Associate Professor of Geology

Dr. John B. Anderson,
Professor of Geology

Houston, Texas
May, 1986
Abstract

The East Eagle Creek area of the Southern Wallowa Mountains lies within the Seven Devils terrane in northeastern Oregon. This terrane comprises a Permian-Triassic volcanic arc assemblage overlain by a thick Triassic-Jurassic carbonate and clastic succession. The Seven Devils terrane is one of four lithotectonic terranes identified in the Blue Mountains region of the North American Cordillera.

Two penetrative phases of Mesozoic deformation are recorded in this portion of the Seven Devils terrane. The earliest deformational event \( (F_1) \) resulted in the formation of tight to isoclinal, northwesterly vergent folds. Associated with this deformation is the development of a close spaced axial planar fracture cleavage in nonmetamorphic and low grade metamorphic rocks and a prominent foliation in metamorphic rocks. Portions of the area were metamorphosed to the amphibolite facies during this event. Microstructural analysis indicates that northwesterly tectonic transport occurred during \( F_1 \) and that strain was coaxial. The second phase of deformation \( (F_2) \) is a cross-folding event that resulted in west-northwesterly trending folds and fracture cleavage. Locally, a well-developed crenulation cleavage is associated with this event.

Timing of these two deformational events is constrained by conformable sedimentary relationships within the Triassic-Jurassic carbonate-clastic package which is deformed by both \( F_1 \) and \( F_2 \) and cross-cut by the Jurassic-Cretaceous post-tectonic Wallowa batholith. The \( F_2 \) event occurred shortly after \( F_1 \) suggesting that the two phases represent discrete points along a continuum of a single deformational event. These events are synchronous with the regional \( D_2 \) (Nevadan) deformation which is manifested in all four Blue Mountains terranes. There is no penetrative record of the earlier Triassic-Jurassic regional \( D_1 \) event in this portion of the Seven Devils terrane.

A non-penetrative post-Cretaceous deformational event affected the region and resulted in block faulting and juxtaposition of metamorphic and non-metamorphic rocks. This deformation is thought to be related to Basin and Range extension.

Penetrative, Mesozoic deformation in the area is attributed to convergence between the Seven Devils and Huntington volcanic arcs. The resultant arc-arc collision caused the development of oppositely verging structures in each arc terrane, similar to the orientation of structures described in the present day converging arcs in the Molluca Sea. Details regarding arc polarity and convergence geometries are still incomplete. Two models are
described which account for the paleontologic, paleomagnetic, petrologic, stratigraphic and structural data gathered thus far.
ACKNOWLEDGEMENTS

I would like to thank my principal advisor, Dr. Hans G. Avé Lallemand for his support and guidance throughout this project. I am also grateful to Dr. John S. Oldow for time spent in the field and for office discussions at Rice. In addition, I want to acknowledge the technical aid and encouragement provided by my other committee member, Dr. John B. Anderson, and by department chairman, Dr. Albert W. Bally.

Others to whom gratitude is due include Howard Brooks of the Oregon Department of Geology and Mineral Industries and Michael Follo and Steve Lundblad of Harvard University for their company and assistance in the field; Richard Gottschalk and Arthur Gelber for constructive discourse; and my parents, Mr. and Mrs. Melvin J. Mirkin, who were most generous and supportive.

This study was financed in part by a grant from the National Science Foundation to H. G. Avé Lallemand and a Houston Geophysical Society scholarship awarded to the author.

Finally, I wish to thank my wife Katharine for her assistance, companionship and patience during my studies.
# TABLE OF CONTENTS

## INTRODUCTION

<table>
<thead>
<tr>
<th>Subsection</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Suspect Terranes</td>
<td>3</td>
</tr>
<tr>
<td>BlueMountains Province</td>
<td>9</td>
</tr>
<tr>
<td>Regional Deformation</td>
<td>13</td>
</tr>
<tr>
<td>Relationship Between Terranes</td>
<td>14</td>
</tr>
</tbody>
</table>

## GEOLOGY OF THE SEVEN DEVILS TERRANE

<table>
<thead>
<tr>
<th>Subsection</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stratigraphy</td>
<td>17</td>
</tr>
<tr>
<td>Permian-Triassic Volcanic Units</td>
<td>18</td>
</tr>
<tr>
<td><strong>Seven Devils Group</strong></td>
<td>18</td>
</tr>
<tr>
<td><strong>Clover Creek Greenstone</strong></td>
<td>20</td>
</tr>
<tr>
<td>Triassic-Jurassic Sedimentary Units</td>
<td>21</td>
</tr>
<tr>
<td>Pre-Tertiary Intrusives</td>
<td>23</td>
</tr>
<tr>
<td>Tertiary Rocks</td>
<td>23</td>
</tr>
<tr>
<td>Quaternary Surficial Deposits</td>
<td>24</td>
</tr>
<tr>
<td>Seven Devils Stratigraphy Compared With Other Cordilleran Terranes</td>
<td>27</td>
</tr>
<tr>
<td>Structure</td>
<td>28</td>
</tr>
</tbody>
</table>

## EAST EAGLE CREEK AREA, NORTHEASTERN OREGON

<table>
<thead>
<tr>
<th>Subsection</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stratigraphy</td>
<td>30</td>
</tr>
<tr>
<td><strong>Clover Creek Greenstone</strong></td>
<td>33</td>
</tr>
<tr>
<td><strong>Lower Sedimentary Series</strong></td>
<td>33</td>
</tr>
<tr>
<td><strong>Martin Bridge Limestone</strong></td>
<td>36</td>
</tr>
<tr>
<td><strong>Hurwal Formation</strong></td>
<td>39</td>
</tr>
<tr>
<td>Metamorphic Rocks</td>
<td>40</td>
</tr>
<tr>
<td>Greenschist Facies</td>
<td>40</td>
</tr>
<tr>
<td>Amphibolite Facies</td>
<td>40</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Subsection</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Structural Geology</td>
<td>43</td>
</tr>
<tr>
<td>Megascoopic Structure</td>
<td>43</td>
</tr>
<tr>
<td>Mesozoic Structure</td>
<td>44</td>
</tr>
</tbody>
</table>
LIST OF PLATES

PLATE I  Geologic Map of the East Eagle Creek Area Southern Wallowa Mountains: Northeastern Oregon

PLATE II  Geologic Cross Sections Through the Southern Wallowa Mountains, NE Oregon

PLATE III  Metamorphic Grade Map of the East Eagle Creek Area Southern Wallowa Mountain: Northeastern Oregon
Structural Analysis of the East Eagle Creek Area, Southern Wallowa Mountains; Northeastern Oregon

INTRODUCTION

The Blue Mountains region of northeastern Oregon is one of the least understood portions of the western Cordillera. Rocks found in the pre-Tertiary terranes of this area record a complex history of tectonic movements. Four distinct lithotectonic assemblages have been identified and designated as separate terranes (Dickinson, 1979) (Figure 1). The southern Wallowa Mountains comprise part of an island arc assemblage known as the Seven Devils terrane. Other terranes in the region are the Huntington arc terrane, Central melange terrane, and Mesozoic clastic terrane. These terranes have been described as so-called suspect or exotic terranes (Coney and others, 1980). This designation is based on geochemical, paleomagnetic, and faunal data suggesting that these rocks did not form at or near a Precambrian craton; thus, these terranes may have formed at a large distance from the North American continent.

The Seven Devils terrane consists of some of the eastern-most exposures of Permian and Upper Triassic island arc volcanic rocks in western North America (Vallier and Batiza, 1978). The Huntington arc terrane is composed of Upper Triassic island arc volcanic and volcaniclastic rocks (Brooks and Vallier, 1978). The Central melange terrane contains Devonian through Triassic age metasedimentary, metavolcanic, and ophiolitic fragments within a matrix of serpentinite and chert-argillite of Triassic to Early Jurassic(?) age (Dickinson and Thayer, 1978). Blueschists of Late Triassic age are found within the melange (Hotz and others, 1977). The Mesozoic clastic terrane consists of Upper Triassic to Middle Jurassic flysch (Dickinson 1979).

Penetrative deformation in the region took place during three periods. The first deformational event lasted from the Late Triassic to Early Jurassic. It is associated with the formation of the subduction related melange and blueschists of the Central melange terrane (Dickinson and Thayer, 1978). The second major phase of deformation occurred in the Late Jurassic and may be related to the accretion of the Blue Mountains terranes to the North American craton and/or the amalgamation of the four Blue Mountains terranes (Hamilton, 1978; Dickinson and Thayer, 1978; Dickinson, 1979; Avé Lallemand and others, 1980; Avé Lallemand and others, 1985). A third phase of deformation is recorded in metamorphic rocks of the Seven Devils terrane in the Riggins, Idaho area (Sutter and others, 1984). Non-penetrative deformation occurred during the Tertiary and may be
Figure 1. Simplified geologic map of NE Oregon from Avé Lallemand (1983); shows distribution of Blue Mountain terranes as defined by Dickinson (1979) and modified by Avé Lallemand (1983). SD = Seven Devils terrane (V-pattern), CM = Central melange terrane (dashed pattern), MC = Mesozoic clastic terrane (fine stippled pattern), HA = Huntington arc terrane (V-pattern); black areas = ophiolite fragments in SD and CM; coarse stipple pattern = Upper Jurassic and Lower Cretaceous granitic rocks; white areas = Cretaceous and younger sediments and volcanic rocks; BA = Baker, H = Huntington, JD = John Day, M = Mitchell, S = Snake River.
related to extension in the Basin and Range province of southern Oregon and Nevada (Lawrence, 1976).

The thesis area is located along East Eagle Creek in the Eagle Cap quadrangle of the southern Wallowa Mountains. This area is of particular interest for two reasons. First, none of the previous geologic studies in the region have employed structural analysis nor emphasized structural geology. Second, in other portions of the Wallowa Mountain range, many of the structural relations are obscured by the intrusion of the Wallowa Batholith. The thesis area is sufficiently remote from the batholith to make structural analysis meaningful.

With the exception of one preliminary study (Avé Lallemant and others, 1980), the relationship of these Blue Mountains terranes to one another and to other Cordilleran suspect terranes has not been based on structural geology. Detailed structural analysis has been carried out in the study area. These data are used to correlate the deformational history of the Seven Devils terrane with the other Blue Mountains terranes and an attempt is made to relate these terranes to the rest of the western Cordillera.

SUSPECT TERRANES

The western or "eugeosynclinal" portion of the North American Cordillera is a composite, tectonic collage of allochthonous and autochthonous lithologic components (Helwig, 1974) (Figure 2). Numerous Paleozoic to Mesozoic rock assemblages of widely variable paleogeographic origins have been juxtaposed by diverse tectonic processes along the evolving western edge of the North American continent (Davis and others, 1978). These distinctive assemblages are termed suspect terranes because their paleogeographic setting with respect to North America is uncertain (Coney and others, 1980).

Throughout late Precambrian and early Paleozoic time, the western margin of North America was a passive continental margin across which thick miogeoclinal and eugeoclinal deposits were draped (Stewart and Poole, 1974). The position of this passive margin has been determined on the basis of \( \text{Sr}^{87}/\text{Sr}^{86} \) initial ratios: rocks with initial ratios greater than .706 are considered to have formed on Paleozoic North America whereas those with initial ratios less than .704 are considered to have formed on oceanic basement (Armstrong and others, 1977). Subduction related to convergence consumed most of the Paleozoic Pacific ocean, and therefore all Paleozoic terranes that now occur west of the edge of the Paleozoic continental margin may have formed at some distance from the craton, and may have accreted to that margin during Phanerozoic time (Coney and others, 1980). In addition, all
Figure 2. Cordilleran Suspect Terranes from Coney and others, 1980). Northeastern Oregon and western Idaho are outlined, W = Seven Devils terrane; BL = Central melange, Mesozoic clastic, and Huntington arc terranes.
PRINCIPAL TERRANES

ALASKA
NS North Slope
Kv Kaquik
En Endicott
R Ruby
Sp Seward Peninsula
I Innoko
NF Nixon Fork
PM Pintoan McKinley
YT Yukon-Tanana
Cl Chuitna
P Peninsular
W Wrangeilla
Cq Chugach and Prince William
TA Tracy Arm
T Taku
Ax Alexander
G Goodnews

CANADA
Ch Cache Creek
St Stikine
BR Bridge River
E Eastern assemblies

WASHINGTON, OREGON & CALIFORNIA
Ca Northern Cascades
SJ San Juan
O Olympic
S Siletzia
BL Blue Mountains
Trp Western Triassic and Paleozoic
  of Klamath Mountains
KL Klamath Mountains
Fh Foothills belt
F Franciscan and Great Valley
C Calaveras
Si Northern Sierra
SG San Gabriel
Mo Mohave
Sa Salinia
Or Oroquoa

NEVADA
S Sonoma
RM Roberts Mountains
GL Golconda

MEXICO
B Baja
V Viscaino

KILOMETERS

0 600
younger terranes outside that margin are considered suspect.

More than 50 suspect terranes have been recognized in the western Cordillera, the
distribution of which is shown in Figure 2. Each terrane is characterized by a common
stratigraphic record that differs from the record in adjacent coeval packages (Monger and
others, 1982). Monger and Price (1979) note that "each terrane consists of one or more
tectonostratigraphic assemblages, each of which can be interpreted by analogy with
modern examples as having formed in a particular tectonic setting." Terranes are separated
from one another by major faults and intrusions, which often are covered by younger
rocks (Monger and others, 1982). These boundaries separate distinct temporal and/or
lithologic sequences and may juxtapose strikingly different faunas (Coney and others,
1980). Paleomagnetic records from some individual terranes vary strongly from that of the
North American craton as well as from one another, thus suggesting that large
displacements and rotations between the terranes themselves and between the terranes and
North America have occurred (Coney and others, 1980).

Jones and others (1978) described the western margin of North America as an
accretionary terrane comprised of allochthonous fragments of varying size and
composition. They cite two periods of accretion: 1) Paleozoic to Early Triassic and 2) Late
Triassic to Early Cenozoic. One such differentiated package of rocks that was accreted to
North America during the latter period is designated as "Wrangellia". Rocks described as
belonging to Wrangellia occur in the Wrangell Mountains, eastern Alaska Range, Talkeetna
Mountains, Baranoff and Chichagof Islands, Alaska, Queen Charlotte Islands and
Vancouver Island, Canada (Jones and others, 1977) (Figure 3). These rocks are correlated
to one another on the basis of each containing a thick Middle and Upper Triassic tholeiitic
basalt unit that is disconformably overlain by late Karnian or early Norian carbonates.
Older rocks, where exposed, consist of Pennsylvanian or younger sedimentary and
arc-related volcanic rocks (Jones and others, 1977).

Wrangellia is interpreted to have been a volcanic arc in Late Pennsylvanian-Permian
time, a submarine to subaerial plateau of rift-related tholeiitic flood basalt in the Late
Triassic and a marine basin and volcanic arc in the Early Jurassic (Monger and Price,
1979; Monger and others, 1982). Paleomagnetic data indicate that Wrangellian rocks
formed at low latitudes, probably within 18 ± 6 degrees of the paleo-equator (Jones and
others, 1977; Hillhouse and others, 1982). Wrangellia represents a composite terrane that
became amalgamated with another exotic terrane (Alexander terrane) prior to the accretion
of both to the North American craton (Coney and others, 1980). The Alexander terrane is
Figure 3. Distribution of rocks designated as Wrangellia, from Jones and others (1977).
Western Limit of Triassic rocks related to the North American craton.
a complex assemblage of volcaniclastic Paleozoic rocks. That Wrangellia became amalgamated with the Alexander terrane at some point during its route from the equator is demonstrated by overlapping Upper Jurassic and Cretaceous strata. Once amalgamated, the resultant super-torrane was subjected to Jurassic and Lower Cretaceous arc activity prior to its accretion to the continent. It is thought that subsequent to its accretion, Wrangellia was fragmented by major horizontal translations and rotations and is presently scattered over approximately 2000 km of the Cordillera from Oregon (?) to Alaska (Coney and others, 1980; Jones and others, 1977).

One such fragment of Wrangellia has been postulated to occur in the Seven Devils-Hells Canyon region of northeastern Oregon (Figure 3). Muller and others (1977) and Jones and others (1977) cite faunal, stratigraphic, and lithologic similarities between rocks of this area and those of Wrangellia on Vancouver Island and in the Wrangell Mountains. Recent paleontological data (Newton, 1983) support the notion that the Seven Devils terrane is a fragment of Wrangellia. A detailed examination of Norian bivalves from the Wrangell Mountains in Alaska and Hells Canyon, Oregon indicate a common low latitude zoogeographic province. Late Triassic bivalve fauna from these two terranes are reported as being more similar to one another than to any other assemblage previously reported from North America (Newton, 1983).

Paleomagnetic data from Upper Triassic volcanic rocks of the Seven Devils terrane record the same low Triassic paleolatitudes (18° ± 6°) as do Wrangellian rocks (Hillhouse and others, 1982). Structural and sedimentologic analyses indicate that these exotic rocks in Oregon were amalgamated with one another during the Late Jurassic (Dickinson and Thayer, 1978; Dickinson, 1979; Ave Lallemant and others, 1980). Accretion to the North American craton may have occurred during the Cretaceous (Ave Lallemant and others, 1985).

Mesozoic rocks of the Blue Mountains province fit the terrane concept as described by Monger and Price (1979) and as such have been so designated by Dickinson (1979). Their location west of the Sr²⁷/Sr²⁶ .704 line and island arc lithologic assemblages suggest an exotic origin. However, many details regarding the relationship between terranes in Oregon with one another and with terranes elsewhere in the Cordillera remain uncertain.
Figure 4. Physiographic map of northeastern Oregon and western Idaho showing individual ranges within the Blue Mountains Province. (After Mckee, 1972)
BLUE MOUNTAINS PROVINCE

The southern Wallowa Mountains lie in the Blue Mountains Province of northeastern Oregon and western Idaho. This province trends east from near Prineville in central Oregon to western Idaho. It contains many individual ranges including the Ochoco, Aldrich, Strawberry, Greenhorn Elkhorn, Wallowa, and Seven Devils Mountains (Figure 4) (Mckee, 1972). The Blue Mountains province separates the Columbia Plateau province on the north from the Basin and Range to the south.

Pre-Cretaceous rocks of the Blue Mountains province crop out as erosional inliers surrounded by Tertiary basalts. These inliers have been divided by Dickinson (1979) into four east-northeasterly trending petro-tektonic terranes (Figure 1). These terranes are metamorphosed to varying degrees and have been intruded locally by Mesozoic granitic plutons (Dickinson, 1979). The terranes are separated by major east-northeasterly trending faults and unconformities which often are covered by Cenozoic lavas. These boundaries converge toward the Seven Devils mountains (Brooks and Vallier, 1978). Rocks of the Wallowa Mountains belong to the Seven Devils volcanic arc terrane. The remaining three terranes are the Central melange terrane, Huntington volcanic arc terrane, and Mesozoic clastic terrane.

Recently, the Blue Mountains terranes have been renamed in accordance with USGS guidelines for nomenclature. The new names are: Wallowa terrane, Olds Ferry terrane, Baker terrane, and Izee terrane for the Seven Devils terrane, Huntington arc terrane, Central melange terrane, and Mesozoic clastic terrane, respectively (Silberling and Jones, 1984). For simplicity, the older names will be retained in this report.

In the north (Figure 1) the Seven Devils terrane consists of Permian-Triassic arc volcanic and volcaniclastic rocks. They are overlain stratigraphically by Triassic-Jurassic carbonates, shales, and slates (Dickinson, 1979). A more detailed description of this terrane will be presented in the following chapter.

The Central Melange terrane is comprised of fragments of mainly Permian and Triassic ophiolitic rocks and an argillite-chert sequence with associated limestones and volcanic tuffs and flows. Tethyan and North American fusulinid faunas occur in separate knockers of limestone which are structurally interleaved within the chert argillite (Dickinson, 1979; Nestell, 1983). Ophiolitic rocks in this terrane may be fragments of oceanic lithosphere upon which the argillite-chert was deposited (Dickinson and Thayer, 1978). However, Avé Lallemant (1976), Phelps and Avé Lallemant (1980), and Gerlach
Figure 5. Stratigraphic column of the Seven Devils terrane stratigraphy, from Vallier (1977).
## COMPOSITE STRATIGRAPHIC COLUMN

<table>
<thead>
<tr>
<th>Thickness in Meters</th>
<th>Graphic Column</th>
<th>Age</th>
<th>Stage</th>
<th>Name and Description</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Jurassic TERT.</td>
<td>Middle and Late</td>
<td>Columbia River Basalt Group: basalt flows</td>
</tr>
<tr>
<td>7500</td>
<td></td>
<td>Triassic</td>
<td>Late</td>
<td>Coon Hollow Formation: mudstone; sandstone; rare breccia and conglomerate</td>
</tr>
<tr>
<td>7000</td>
<td></td>
<td>Permian</td>
<td>Early</td>
<td>Hurwal Formation: shale and limestone</td>
</tr>
<tr>
<td>6500</td>
<td></td>
<td>Permian</td>
<td>Early</td>
<td>Martin Bridge Formation: limestone and rare dolomite</td>
</tr>
<tr>
<td>6000</td>
<td></td>
<td>Permian</td>
<td>Early</td>
<td>Lower Sedimentary Series: conglomerate; epiclastic breccia, sandstone, and shale; pyroclastics; metabasalt and keratophyre</td>
</tr>
<tr>
<td>5500</td>
<td></td>
<td>Permian</td>
<td>Early</td>
<td>Wild Sheep Creek Formation: metabasalt; meta-andesite keratophyre; tuff; conglomerate; breccia, sandstone, and argillite</td>
</tr>
<tr>
<td>5000</td>
<td></td>
<td>Permian</td>
<td>Early</td>
<td>Hunsaker Creek Formation: pyroclastic breccia; tuff; conglomerate; breccia; sandstone; argillite; rare limestone; keratophyre and spilite flows</td>
</tr>
<tr>
<td>4500</td>
<td></td>
<td>Permian</td>
<td>Early</td>
<td>Windy Ridge Formation: keratophyre flows and tuff</td>
</tr>
<tr>
<td>4000</td>
<td></td>
<td>Permian</td>
<td>Early</td>
<td>Basement rocks: amphibolite, metagabbro, metaquartz, diorite, hornblende schist, mylonite, gneissic mylonite; exposed at Oxbow of the Snake River, south of Pittsburg Landing, and near the mouth of the Immaha River</td>
</tr>
</tbody>
</table>
and others (1981) have suggested that the Canyon Mountain and Sparta ophiolites in this terrane formed in a volcanic-arc or in a back arc basin, and that the Sparta ophiolite is related to the Seven Devils volcanic arc. Blueschists of Late Triassic age are found within this terrane near Mitchell, Oregon (Hotz and others, 1977). This chaotic assemblage of dismembered oceanic crust and marine sediments is thought by Dickinson (1979) to be the result of subduction processes related to the Huntington volcanic island arc to the southeast. However, Mullen (1985) cites petrologic data suggesting that the Central melange terrane represents a forearc associated with the Seven Devils volcanic arc. Avé Lallemant and others (1985) suggest that it is related to both the Seven Devils and Huntington volcanic arc terranes.

The Huntington volcanic arc terrane lies at the southeastern margin of the Blue Mountains province. This terrane is comprised of a belt of Upper Triassic mafic to silicic volcanic and volcanoclastic rocks with a minor amount of Triassic limestone (Brooks and Vallier, 1978; Dickinson, 1979). The Upper Triassic volcanic rocks belong to the Huntington Formation (Brooks, 1979). They are submarine in origin, and range in composition from basalt to rhyolite, but are primarily andesitic. Quartz diorite and granodiorite plutons dated at 220-190 Ma intrude these volcanic rocks. The rocks of the Huntington arc terrane resemble the Upper Triassic rocks of the Seven Devils terrane, except for the absence of the thick and continuous Martin Bridge Limestone and the presence of much younger intrusives. Dickinson (1979) noted that:

Structural relations between this terrane and more interior parts of the continent are masked by Cenozoic cover. The volcanic arc either formed within an ocean basin and was later sutured to the continental margin by plate interactions, or formed in place along a submerged continental margin. In either case, internal facies relations indicated that eruptions were partly submarine within an island arc.

Very little information is available from the Huntington arc terrane due to the extensive Tertiary volcanic cover.

The Mesozoic clastic terrane consists mainly of Upper Triassic to Upper Jurassic clastic sediments interbedded locally with volcanic and volcanoclastic rocks (Brooks and Vallier, 1978; Dickinson, 1979). It has been proposed that near the Snake River the Lower Jurassic beds overlie unconformably the Upper Triassic volcanic rocks of the Huntington Formation to the southeast and are separated from the Central melange terrane to the north by the Connor Creek fault (Figure 6) (Brooks and Vallier, 1978). However,
Figure 6. Cross sections showing inferred relations of the four Blue Mountains terranes. From Brooks and Vallier (1978)
the contact between the Huntington arc and Mesozoic clastic terranes is apparently a sheared, but originally depositional contact (Avé Lallemand, 1983). Dickinson (1979) has proposed that the lower part of the sequence was deposited in an intra-arc basin and the upper part was deposited in a forearc basin.

No continental detritus is found in any of the pre-Cretaceous rocks of the Blue Mountains. This has been interpreted to indicate that all of the terranes originated from within an oceanic region and that any continental detritus that may have been present was subducted into oceanic trenches between the continent and the arc terranes (Thayer in Dickinson and Thayer, 1978). Alternatively, the lack of continental detritus may indicate the presence of a volcanic chain or other barrier to block the oceanward transport of continental clastics (Dickinson in Dickinson and Thayer, 1978).

Several episodes of Mesozoic plutonism occurred in eastern Oregon and western Idaho (Armstrong and others, 1977). These plutonic events can be divided roughly into three time frames: Middle Triassic-Early Jurassic, Late Jurassic--Early Cretaceous, and Late Cretaceous (Brooks and Vallier, 1978). The older plutons occur in the Seven Devils and Huntington arc terranes and are related to arc magmatism. The most important intrusions (by volume) are the Upper Jurassic--Lower Cretaceous compositionally diverse Bald Mountain and Wallowa batholiths in Oregon and the Upper Cretaceous Idaho batholith. Cross cutting relationships between these intrusive rocks and the country rocks provide important age constraints on deformation manifested in the terranes (Armstrong and others, 1977). Brooks and others (1976) note that some of the plutons cut contacts between several terranes. Plutonism in the area apparently post-dated the amalgamation of the accreted oceanic and island arc terranes.

REGIONAL DEFORMATION

Three major deformational events are known to have occurred in the Blue Mountain region. The first was a prolonged event that lasted from Late Triassic to Early Jurassic time. It is characterized by syndepositional deformation associated with the formation of subduction related melange and blueschists of the Central melange terrane (Dickinson and Thayer, 1978). This event resulted in north to northeast trending folds and faults in Upper Triassic and lower Jurassic volcaniclastic and marine sediments in the John Day area. It was also during this time that three mylonitic shear zones formed in the Seven Devils terrane (Avé Lallemand and others, 1985).
The second major phase of deformation occurred in the Late Jurassic and may be related to the accretion of the Blue Mountains region to the North American craton and/or the amalgamation of the four Blue Mountain terranes (Hamilton, 1978; Dickinson and Thayer, 1978; Dickinson, 1979; Avé Lallémant and others, 1980; Avé Lallémant and others, 1985). Deformation ceased prior to the intrusion of granitic plutons 160-95 million years ago into all four terranes (Armstrong and others, 1977). This event resulted in major east trending thrust faults and folds in the western portion of the region and northeasterly trending structures in the east (Brooks and Vallier, 1978), "The present-day distribution of pre-Tertiary rocks and the shape of the four terranes are mainly the result of the Late Jurassic orogeny." (Avé Lallémant and others, 1980)

A Cretaceous deformation is recorded east of the Blue Mountains in the Riggins, Idaho area. Radiometric age dates indicate that rocks in this region reached peak metamorphic temperature at 118 Ma ago (Sutter and others, 1984). This event is thought to be related to movement between the Seven Devils terrane and the continent. No penetrative evidence of this event is reported in northeastern Oregon.

Non-penetrative deformation occurred in eastern Oregon during the Tertiary. This is manifested by block faulting presumably related to the development of the Basin and Range province which resulted in the range front faults along which the present day Wallowa mountains are uplifted (Lawrence, 1976). The present distribution and structural trends of pre-Tertiary rocks suggest that oroclinal bending has resulted in clockwise rotation of these terranes (Hamilton and Myers, 1966). This is supported by paleomagnetic data which indicate the area has undergone 65 degrees of dextral rotation (Wilson and Cox, 1979, Hillhouse and others, 1982) (Figure 7).

RELATIONSHIP BETWEEN TERRANES

As discussed previously, rocks in northeastern Oregon have been described as being related to Wrangellia. Recent arguments based on geochemical, petrologic and paleomagnetic data have cast some doubt as to the relationship between Wrangellia and the Seven Devils terrane. Vallier and Batiza (1978) and Sarewitz (1983) have shown that Triassic volcanic rocks of the Seven Devils terrane are distinctly calcalkaline and formed in an island arc setting. This is contrasted with the vast quantities of coeval rift related tholeiitic basalts found in Wrangellia (Jones and others, 1977).
Figure 7. Hypothetical pre-mid-Miocene tectonic rotation of Blue Mountains province. Dashed outlines show pre-Cretaceous positions of major Mesozoic provinces. Modified from Dickinson and Thayer (1978).
In addition to the problem of the relationship of the Seven Devils terrane to other Cordilleran terranes, it is still unclear as to the relationship between the Seven Devils and Huntington arc terranes. Previous workers have noted the following possibilities:

1) The Seven Devils and Huntington arc terranes are stratigraphically and structurally continuous and therefore belong to the same arc (Brooks and Vallier, 1978),

2) The Seven Devils and Huntington arc terranes are detached fragments of the same arc (Brooks and Vallier, 1978; Avé Lallemand, 1983),

3) The Seven Devils and Huntington arc terranes are portions of different and unrelated arcs (Avé Lallemand and others 1985).

Several lines of evidence support the notion that these two terranes represent portions of the same arc. Brooks and Vallier (1978) cite the following:

1) The outcrop patterns and structural trends converge toward the Seven Devils Mountains; 2) youngest volcanic strata in both terranes are Karnian, and possibly Norian in age; 3) lithologies are similar and even color variations indicate similar changes in oxidation state during late stages of volcanism and sedimentation; 4) both sequences are in the greenschist and zeolite facies of regional metamorphism; and 5) lava chemistries show a calcalkaline differentiation trend.

Exposures of the two terranes are separated by Cenozoic lavas and by rocks of the Central melange terrane. It is possible that the Central melange terrane does not continue to the east and that the Seven Devils and Huntington arc terranes are continuous and thus portions of the same arc (Brooks and Vallier, 1978). Alternatively, the Central Melange terrane may be continuous beneath the Cenozoic lavas thus separating fragments of two different volcanic arcs. The latter possibility, if true, would indicate that the two arc terranes were juxtaposed tectonically along a major suture (Brooks and Vallier, 1978). A possible suture occurs in the Oxbow region of Hells Canyon. Mylonites from the Oxbow shear zone indicate a minimum of 65 km of left-lateral displacement (Avé Lallemand and others, 1985). This shearing has been related to left-oblique plate convergence during the Triassic. Its existence permits an interpretation that the Seven Devils and Huntington arc terranes were at one time, a continuous, single arc (Avé Lallemand, 1983).

Evidence that contradicts the single arc concept is that Upper Triassic volcanic rocks of the Seven Devils arc terrane are overlain by the Late Triassic Martin Bridge Limestone, indicating that volcanism ceased after early Late Triassic time. This is contrasted with
Upper Triassic volcanic rocks and Upper Triassic to Lower Jurassic intrusives of the Huntington arc terrane that are overlain unconformably by Lower and Middle Jurassic flysch (Brooks and Vallier, 1978). Fragmentary paleontologic data suggest that volcanism continued throughout most of Late Triassic time and possibly until the Middle Jurassic in the Huntington arc terrane (Dickinson and Thayer, 1978; Dickinson, 1979).

The continuation of volcanism in the Huntington arc terrane after volcanic cessation in the Seven Devils terrane is opposite to what one would expect if the two were originally continuous and have been juxtaposed along the left lateral Oxbow shear zone. An alternate interpretation is that the Seven Devils and Huntington arc terranes are considered to be two separate arcs. The three shear zones found in the Hells Canyon are thought to represent intra-arc strike-slip fault zones related to Late Triassic left-oblique subduction (Avé Lallemant and others, 1985). This interpretation relates the Central melange terrane to east directed subduction beneath the Huntington arc and westerly subduction beneath the Seven Devils arc. This model explains the position of the blueschists near Mitchell which occur northwest of younger radiolarian chert-bearing melange.

Paleomagnetic data that suggest a common low paleolatitude for Wrangellia, the Seven Devils terrane, and Huntington arc terrane are somewhat ambiguous with regard to the proximity of these terranes to the paleolatitude of the Hells Canyon region of continental North America (Hillhouse and others, 1982). Although the data indicate that the Seven Devils and Huntington arc terranes formed at a paleolatitude of approximately $18^\circ \pm 6^\circ$, it is not known whether this paleolatitude lies in the northern or southern hemisphere.

**GEOLOGY OF THE SEVEN DEVILS TERRANE**

**STRATIGRAPHY**

There are five major stratigraphic units within the Seven Devils terrane. Included are Permian and Triassic volcanogenic rocks, Triassic carbonates, Triassic and Jurassic epiclastics, Miocene flood basalts, and Quaternary surficial deposits. The Permian and Triassic volcanic rocks belong to the Seven Devils Group (Vallier, 1977). The Seven Devils Group is overlain by the Upper Triassic Martin Bridge Limestone which is overlain by the Upper Triassic-Lower Jurassic Hurwal Formation. The Hurwal Formation is overlain by the Upper Jurassic Coon Hollow Formation. Basalt of the Columbia River Group caps the pre-Tertiary rocks and Quaternary surficial deposits comprised of glacial
till, alluvium, and colluvium locally rest on older rocks. This stratigraphy is summarized in Figure 5.

Permian-Triassic Volcanic Units

Previous workers have referred to the chiefly metavolcanic rock sequences found in the Snake River Canyon and Seven Devils Mountains as the "Seven Devils Volcanics" after Anderson (1930) (Vallier, 1977). More recently, these rocks were elevated to Group status (Vallier, 1977). Correlative rocks found in the Wallowa Mountains include the Clover Creek Greenstone (Gilluly, 1937), Lower Sedimentary Series (Smith and Allen, 1941), and Gold Creek Greenstone (Prostka, 1962). A correlation chart showing the relationship between the stratigraphic units on a regional scale was presented by Brooks and Vallier (1978) and is shown in Figure 8. The proliferation of stratigraphic nomenclature to describe the same rocks is due to the rapid changes of facies that occur over short distances, few fossils, and the absence of regionally correlatable marker horizons (Brooks and Vallier, 1978).

Seven Devils Group

Vallier (1977) defined a thick sequence of Permian through Triassic volcanic and volcaniclastic rocks that are exposed in the Wallowa Mountains, Snake River Canyon, and Seven Devils Mountains as the Seven Devils Group. These are the oldest exposed rocks within the thesis area and are among the oldest in northeastern Oregon and western Idaho. The Seven Devils Group consists of thick basalt, andesite, and keratophyre flows intercalated with pyroclastic and volcaniclastic units, sandstones, conglomerates, shales, and rare limestones. These rocks have been locally metamorphosed to the greenschist and zeolite facies. The type section for the Seven Devils Group is in the Snake River Canyon and includes exposures of four formations. These are the Lower Permian (?) Windy Ridge Formation, Lower Permian Hunsaker Creek Formation, Middle and Upper Triassic Wild Sheep Creek Formation, and the Upper Triassic Doyle Creek Formation.

The Windy Ridge Formation is comprised predominantly of grayish-green quartz keratophyre tuff, quartz keratophyre tuff breccia, and rare quartz keratophyre flows. The base of the Windy Ridge Formation is not exposed, although shearing in the lower part of the formation suggests that it is in tectonic contact with the underlying units. The top of
Figure 8. Stratigraphic correlation chart for Mesozoic units in the Blue Mountains Region. From Brooks and Vallier (1978).
<table>
<thead>
<tr>
<th>Era</th>
<th>Southern Wallowa Mountains</th>
<th>Seven Devils Mountains/Snake River Canyon</th>
<th>Huntington Area</th>
<th>Aldrich Mountains</th>
<th>Izee-Suplee Area</th>
<th>Mitchell Area</th>
<th>Riggins Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cretaceous</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Middle</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jurassic</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower</td>
<td>Hurwal Formation</td>
<td>Lucille Slate</td>
<td>Weatherby Formation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Martin Bridge Limestone</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Triassic</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower</td>
<td>Clover Creek Greenstone</td>
<td>Seven Devils Group</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Middle</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Permian</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower Part</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

DISMEMBERED OCEANIC CRUST, INCLUDES ELKHORN RIDGE ARGILITE IN EASTERN OREGON, AND THE FIDDLE CREEK SCHIST, LIGHTNING CREEK SCHIST AND BERG CREEK AMPHIBOLITE IN THE RIGGINS AREA
the Windy Ridge Formation is either a fault or unconformity and the minimum thickness of the unit has been estimated to be 500m (Vallier, 1977). No radiometric or fossil dates are available for these rocks and they are therefore considered to be either equivalent to or older than the Hunsaker Creek Formation (Vallier, 1977).

The Hunsaker Creek Formation is a thick (estimated to be 2500m thick) sequence of metamorphosed mostly volcanogenic rocks consisting of pyroclastic breccia, agglomerate, tuff, volcanic breccia, conglomerate, sandstone, siltstone, and minor limestone, quartz keratophyre, and spilite flows. Vallier interprets the andesitic and dacitic character of these Lower Permian rocks to indicate that subduction was occurring along an extensive island arc complex of which these rocks are only a small remnant.

Overlying the Hunsaker Creek Formation unconformably is the Middle and Late Triassic Wild Sheep Creek Formation: a 2500m thick sequence of metamorphosed basalt and andesite flows, volcaniclastic rocks, argillite, graywacke, and limestone. Geochemical data on the flow rocks indicate that they formed in a volcanic arc (Vallier and Batiza, 1978).

Conformably overlying the Wild Sheep Creek Formation is the Late Triassic Doyle Creek Formation which consists of metamorphosed basalt, andesite, volcaniclastic rocks, conglomerate, sandstone, shale and limestone. The Doyle Creek Formation is estimated to be approximately 500 m thick.

Clover Creek Greenstone

The Clover Creek Greenstone (Gilluly, 1937) in the Baker quadrangle is correlative with the Seven Devils Group. It consists of quartz keratophyre, keratophyre, spilite, albite diabase, keratophyre and quartz keratophyre tuff and breccia, meta-andesite, chert, conglomerate, argillite and limestone. The formation is estimated to have a minimum thickness of 4000 feet (Gilluly, 1937) but because of structural complexity this estimate may be incorrect. Ross (1938) extended the distribution of the Clover Creek Greenstone into the Wallowa Mountains. According to Ross the Clover Creek Greenstone consists predominantly of andesite and dacite rather than keratophyre.

Overlying the Clover Creek Greenstone with apparent conformity is a group of partially metamorphosed argillites, sandstones, and conglomerates. These rocks are named "Lower Sedimentary Series" (Smith and Allen, 1941). Fossil evidence indicates that they are Upper Triassic (middle Kamin) in age. In the Sparta quadrangle, just south of the present study area, a sequence of metavolcanic rocks occur which were named the
Gold Creek Greenstone by Prostka (1962). It grades abruptly upward into the Lower Sedimentary Series, and both grade laterally into the Clover Creek Greenstone. A combined thickness of 5200 feet (1600 m) was estimated (Brooks and Vallier, 1978).

The predominance of volcanic fragments—both clasts and matrix material within the Lower Sedimentary Series suggest that it has a pyroclastic origin. Prostka (1962) cites the absence of glass shards, pumice, and bombs as evidence for phreatic eruptions being responsible for the deposit. Evidence for marine deposition is found in the siltstone sandstone in which Laudon (1956) found marine fossils and interbedded limestone.

The sedimentary structures found in the deposits suggest that they were transported by mass movement, possibly turbidity currents. These structures include graded bedding, load casts, convolute bedding, and cross bedding. According to Follo and Siever (1984) sedimentary structures and textures suggest erosion of a local Seven Devils volcanic arc and deposition in a shallow environment.

Correlation of units that occur in the Seven Devils Mountains, Snake River Canyon, and Wallowa Mountains are good in spite of the presence of intervening Cenozoic volcanics (Brooks and Vallier, 1978). The Wild Sheep Creek Formation of the Snake River Canyon area is correlated with the Gold Creek Greenstone in the southern Wallowa Mountains. Correlative rocks with the Doyle Creek Formation in the Snake River Canyon area are found in the upper part of the Clover Creek Greenstone in the northern Wallowa Mountains, and in the Lower Sedimentary Series in the southern Wallowas.

**Triassic-Jurassic Sedimentary Units**

Upper Triassic volcanic rocks of the Seven Devils Group and Clover Creek Greenstone in the Wallowa Mountains-Snake River Canyon-Seven Devils Mountains region are overlain by Upper Triassic and Lower Jurassic carbonate and clastic rocks of the Martin Bridge Limestone, the Hurwal Formation, and Coon Hollow Formation (Brooks and Vallier, 1978).

The Martin Bridge Limestone consists of interbedded massive and thin-bedded limestones which have been metamorphosed to marble in many places. These rocks are named after the bridge across Eagle Creek "... near which the best preserved fossils were found." (Ross, 1938) Formation thicknesses from undeformed sections average between 1100 and 1200 feet in the Wallowa Mountains (Nolf, 1966) and 500m in the Snake River
Canyon (Vallier, 1977). Vallier (1974) described the Martin Bridge Limestone as being predominantly a platform limestone. Nolf (1966) noted the existence of local reef facies in the northern Wallowa Mountains, and Laudon (1956) described the Martin Bridge sequence in the Snake River Canyon as a well defined reef limestone. The Martin Bridge Limestone may have formed in a type II carbonate platform which comprises knoll shaped reefs on gentle slopes (Vallier, 1977). The contact of the Martin Bridge Formation with the underlying Seven Devils Group has been described as being both conformable and unconformable by various authors. Smith and Allen (1941) described the Martin Bridge Formation in the northern Wallowa Mountains as overlying the Lower Sedimentary Series "with apparent strong unconformity". Prostka (1962) described the same contact as being conformable in the Sparta quadrangle and Vallier (1977) notes that the contact is "likely an unconformity" in the Snake River Canyon.

Abundant fossils have been collected from the Martin Bridge Formation. From the northern Wallowa Mountains upper Triassic (Karnian) nautiloids and ammonoids (Smith and Allen, 1941) and early Norian fossils (Nolf, 1966; Newton, 1983) have been reported. Karnian ammonoids and molluscs have also been reported in the southern Wallowa Mountains (Ross, 1938; Prostka, 1962, Carnahan, 1963). Thus, deposition of the Martin Bridge Formation occurred during the Late Karnian and Early Norian ages of the Late Triassic (Brooks and Vallier, 1978).

The Martin Bridge Formation is overlain conformably by the argillaceous sediments and limestones of the Hurwal Formation. Smith and Allen (1941) defined the Hurwal Formation in the northern Wallowa Mountains and named it after Hurwal divide which is located in the Wallowa Lake quadrangle. "In the Salmon River Canyon near Lucile, Idaho, equivalent strata were named the Lucile Slate (Hamilton, 1963)" (in Brooks and Vallier, 1978). Nolf (1966) describes the Hurwal as consisting predominantly of dark, thin bedded mudstone and limy mudstone with minor amounts of sedimentary breccia. Brooks and Vallier (1978) note that "With increasing stratigraphic elevation, the Hurwal Formation becomes predominantly thin-bedded argillite and siltstone, which suggests deepening of the basin; some clastic sequences are flysch like, with some sediments derived from uplifted volcanic land masses." Prostka (1962) suggested that the sediments of the Hurwal Formation record the onset of deformation.

The Hurwal Formation exceeds 2000m in thickness in the northern Wallowa Mountains and has yielded fossils ranging in age from Late Triassic (Early Norian) to Early Jurassic (Early Toarcian) (Nolf, 1966). "Most stratigraphic sections of the Hurwal
Formation are truncated at the top by granitic intrusive bodies or by the Columbia River Basalt Group. No overlying sedimentary formation was defined or identified and the Hurwal is thus the youngest pre-batholithic unit." (Nolf, 1966)

In the Snake River Canyon area the Oxfordian Coon Hollow Formation overlies the Hurwal Formation with apparent unconformity. This unit occurs only in the Snake River Canyon area and consists of mudstone, sandstone, rare breccia, and conglomerate (Morrison, 1964). These rocks may have been deposited in basins similar to those now forming off the coast of southern California (Brooks and Vallier, 1978).

Pre-Tertiary Intrusives

Three separate phases of plutonism affected the region during pre-Tertiary times (Taubeneck, 1964; Neal 1973). The oldest (Permian to Late Triassic; Walker, 1981, 1982; Avé Lallemand and others, 1980; Balcer, 1980) plutonic rocks are related to volcanic arc magmatism. The second event resulted in a Jurassic intermediate to mafic dike-sill complex emplaced prior to the intrusion of the volumetrically more significant Jurassic-Cretaceous Wallowa and Bald Mountain batholiths. Jurassic sills have been observed within the argillite of the Hurwal Formation. These intrusions predate the major phase of Jurassic deformation since they are themselves deformed. These intrusive rocks consist of metagabbro, diorite, and quartz diorite (Neal, 1973).

The compositionally diverse Wallowa batholith is a granitic body comprised of four major plutons and several smaller bodies (Taubeneck, 1964). Each pluton is zoned with borders of tonalite grading inward to cores of granodiorite. Two satellite stocks thought to be genetically related to the batholith also intrude the Seven Devils terrane. They are the Sawtooth stock in the northern Wallowa mountains and the Cornucopia stock which is present in the thesis area. The Bald Mountain batholith occurs in the western portion of the Seven Devils terrane and comprises at least eight rock types from norite to quartz monzonite, with the majority consisting of tonalite and granodiorite (in Brooks and Vallier, 1978).

Tertiary Rocks

Thick flows of the Columbia River Basalt Group overlie the pre-Tertiary rocks with angular unconformity. These extrusive deposits are distributed throughout the Columbia
Plateau in Washington and Oregon and the Blue Mountain and Snake River Canyon areas. In the Seven Devils terrane the basalt fills the valleys between mountain peaks and covers many of their flanks. Several of the higher peaks in the range are capped by basalt flow remanents (Figure 9) suggesting that the entire area was covered by basalt prior to uplift and glaciation of the range (Nolf, 1966).

Single lava flows from the Columbia River Basalt Group are generally 30-100 feet thick. Maximum exposed thickness of the Columbia River Basalt Group is approximately 3000 feet (Weiss and others, 1976). The basalt is generally grayish-brown, vesicular, and exhibits prominent columnar jointing. Most of the flows are porphyritic with phenocrysts of zoned plagioclase, zoned titanaugite, and olivine.

Contacts between the Columbia River Basalt Group and older rocks indicate that the basalt was extruded on a surface of irregular topography with local relief in excess of 1000 feet (Weiss and others, 1976; Vallier, 1977). In places the basalt is separated from the pre-Tertiary rocks by Tertiary gravel deposits. These deposits are of mixed lithology but are dominated by quartzite. No local source of quartzite has been identified, and it has been suggested that it was derived from Precambrian and lower Paleozoic quartzites that occur in southeastern Idaho (Weiss and others, 1976; Vallier, 1977).

Potassium-argon age dates for rocks of the Columbia River Group indicate a range from 15.4 ± 0.3 Ma to 13.5 ± 0.3 Ma (Baksi and Watkins, 1973). Fossil plants of late Miocene age have been found in sedimentary beds intercalated with the Columbia River Basalt (Prostka, 1962).

**Quaternary Surficial Deposits**

Intense glaciation during late Pleistocene time resulted in the carving of spectacular U-shaped valleys and cirque basins that now contain lakes (Figure 10). Glacial scouring resulted in deposits of till on the flanks and floors of valleys, cirques, and on some ridge tops. Much of the detritus was carried out of the glacial valleys away from the range front where it was deposited in large terminal and lateral moraines (Weiss and others, 1976). Alluvial deposits from streams which replaced the glaciers cover the floors of most valleys and colluvium and talus mantles many of the slopes.
Figure 9. Photograph of Aneroid Mountain, elevation 9702 feet located in the northern Wallowa Mountains. Note Tertiary Columbia River Basalt capping summit.
Figure 10. Photograph of East Eagle Creek Canyon looking north. Note U-shape of this glacial valley. Prominent peak in the foreground is Eagle Cap.
Seven Devils Stratigraphy Compared With Other Cordilleran Terranes

The pre-Tertiary stratigraphy of the Seven Devils terrane is, in a broad sense, very similar to that found in other Cordilleran terranes. Rocks from the Seven Devils Group have been correlated with coeval rocks in Nevada, southwestern Oregon, Washington, western Canada, and Alaska (in Vallier, 1977). Jones and others (1977) cite lithologic and faunal similarities between the pre-Tertiary section of the Seven Devils Terrane and rocks from the same interval found in the Queen Charlotte Islands, the Wrangell Mountains, and on Vancouver Island, which together form the terrane Wrangellia. Norian bivalve fauna from the Seven Devils Terrane are strikingly similar to those of Wrangellia with a Simpson similarity coefficient of 77 and Sorenson similarity coefficient of 0.65 (Newton, 1983). These similarities together with paleomagnetic data are the basis for considering the Seven Devils terrane to be a detached fragment of Wrangellia (Jones and others, 1977; Muller, 1977; Hillhouse and others, 1982).

However, closer comparison of the Nikolai Greenstone (Wrangellian volcanics) with the Seven Devils Group volcanics has revealed important differences that suggest the two terranes are not the same. Geochemical data from the Late Triassic meta-volcanic rocks of the Seven Devils terrane indicate that they are calc-alkaline volcanic rocks with a complete suite of flow rocks and fractionated rare-earth element distribution. The rocks were originally basalt, basaltic andesite, andesite, dacite and rhyolite and low in K2O. This composition is indicative of an intra-oceanic island arc origin (Vallier and Batiza, 1978; Sarewitz, 1983). In contrast, the Late Triassic volcanic rocks from Wrangellia consist of thick tholeiitic pillow lava, pillow breccia, aquagene tuffs, and amygdaloidal flows. These basalts have been alternatively described as having erupted as flood basalts, oceanic tholeiites, and plateau basalt fields. There are no data to suggest that the Wrangellian basalts originated in an island arc setting (Sarewitz, 1983).

The faunal similarities between Upper Triassic carbonates of Wrangellia and the Seven Devils terrane in addition to the paleomagnetic data from the Triassic volcanics suggest that the two terranes formed in close proximity with one another (Sarewitz, 1983). However, the geochemical data indicate that Triassic extrusive rocks formed in very different tectonic settings. To resolve these conflicting data it has been proposed that the Seven Devils and Wrangellian extrusive rocks represent island arc volcanics and back arc
rift basalts, respectively, and are thusly related (Sarewitz, 1983; Avé Lallemant and others, 1985).

STRUCTURE

The section on Regional Deformation (page 13) described three major phases of orogenic activity which were recorded throughout the entire Blue Mountains region. Each of these deformational events affected portions of the Seven Devils terrane to varying degrees. The first phase of deformation occurred during the Triassic-Early Jurassic and resulted in the formation of three mylonitic shear zones in the Snake River Canyon area (Avé Lallemant and others, 1985). Detailed structural analysis of two of these shear zones and reconnaissance of the third indicate that they formed by left-lateral strike-slip motion.

There has been some ambiguity regarding the existence of major structures associated with this early phase throughout the rest of the Seven Devils terrane. Ross (1938) reported an unconformity between the Triassic Martin Bridge Limestone and the underlying Lower Sedimentary Series in the southern Wallowa Mountains which suggests that the early phase of deformation affected this area. Smith and Allen (1941) apparently confirmed the existence of the intra-Triassic unconformity. Subsequent workers, however, found the Permian through Lower Jurassic section to be conformable (Laudon, 1956; Prostka, 1962; Nolf, 1966). On the other hand, Vallier (1977) suggested that the Martin Bridge Formation in the Snake River Canyon was separated from underlying rocks by an unconformity. Thus it has been unclear as to how pervasive this deformational event is in the Seven Devils terrane.

The second major phase of deformation occurred in all four Blue Mountain terranes and is clearly manifested in the Seven Devils terrane. Pre-Tertiary structures of the Wallowa mountains consist of northeasterly, easterly, and northwesterly trending folds (Ross, 1938; Prostka, 1962; Nolf, 1966) and possibly some thrust faults of similar orientations (Prostka and Bateman, 1962; Walker, 1979). Varying interpretations have been offered regarding the timing and cause of these structures. Much of the confusion is due to the presence of the large, post-tectonic Cretaceous Wallowa batholith. Many workers attributed the deformation to forceful intrusion of the batholith (Nolf, 1966; Taubeneck, 1964). Nolf (1966) described Mesozoic structures in the northern Wallowa Mountains as being geometrically and genetically related to the Wallowa Batholith. For this reason he believes that all information on the "regional structure" has been obscured due to the proximity of the batholith.
Prostka (1963) interpreted the folds in the area to be the result of fault displacements in the underlying crystalline basement. Evidence cited in support of this idea includes variable intensity of folding and deviations in the strike of bedding and orientation of fold axes. Tightly folded zones were interpreted as overlying large faults in the basement rocks and the variation in bedding and fold axis attitudes was related to drag along these basement faults.

Weiss and others (1976) suggested that this deformation occurred prior to the emplacement of the batholith. This interpretation was based on field relationships in the southern Wallowa Mountains where folding that occurs in the north continues to the south for several miles beyond the batholith.

The third phase of regional deformation is recorded only in rocks of the Seven Devils terrane located in western Idaho. This event is dated at approximately 118 Ma and resulted in penetrative deformation and metamorphism to the amphibolite facies (Sutter and others, 1984; Lund and Snee, 1985). Deformation is interpreted to be the result of the juxtaposition of the Seven Devils terrane with the North American continent along a right-lateral, transcurrent fault (Lund and Snee, 1985).

A younger episode of deformation which affected the Seven Devils terrane began in the early or middle Tertiary (Weiss and others, 1976). It began with broad arching of the region which continued during the extrusion of the Miocene Columbia River Group basalts as evidenced by offlap relations of earlier flows. Uplift of the range area occurred after the basalt had been laid down; displacement of flows indicates a minimum throw of 5000 feet. Movement along range front faults ceased prior to the late Pleistocene as evidenced by the continuous surface of glacial moraines across the fault.
EAST EAGLE CREEK AREA, NORTHEASTERN OREGON

The study area is located on both sides of East Eagle Creek in the Eagle Cap quadrangle in the southern Wallowa Mountains between 45° 00' and 45°05' north latitude and 117° 22.5' and 117° 16.5' west longitude (Figure 11). The northeastern portion of the study area lies in the federally designated Eagle Cap Wilderness with the remainder situated in the Wallowa-Whitman National Forest. Access to the field area is limited to gravel logging roads and a well-maintained network of hiking trails. No motor vehicles are permitted in the Eagle Cap Wilderness.

Weiss and others (1976) describe the Wallowa Mountains as being roughly domed shaped sloping gently toward the surrounding lowlands with the exception of the northeast flank where the range rises abruptly more than a mile above the adjacent valley. A well developed radial drainage system is centered on Eagle Cap, the prominent peak for which the wilderness was named. The northern portion of the thesis area is mountainous, while the southern part is a dissected plateau. Maximum relief in the study area is approximately 4900 feet; the maximum elevation is 9048 feet at the summit of Krag Peak.

Summer temperatures range from 75° F during the day to sub-freezing after dark. Snow that begins to fall in September accumulates to depths that restrict travel in the higher elevations until late July. Although no true glaciers exist in the Wallowa mountains, snow remains on the north facing slopes throughout many summers.

Outcrop in the thesis area (in particular, that of Mesozoic country rock) is obscured in many places by thick stands of ponderosa pine, lodgepole pine, Douglas fir, spruce, and tamarack (Figure 12). Much of the rocks of interest have been covered by thick flows of Columbia River Basalt, and much of the basalt has in turn been obscured by thick, grassy meadows. Rocks are generally well exposed along valley walls, creek beds, and selected road cuts. Spectacular and somewhat inaccessible outcrops occur on the flanks of the higher peaks.

STRATIGRAPHY

While a detailed analysis of the stratigraphy within the thesis area was beyond the scope of the present study, several aspects are important. For mapping purposes it was necessary to examine the attributes that were peculiar to each unit depicted on the geologic map. In addition, a careful examination of contacts between stratigraphic units was
Figure 11. Study Area Location Map.
Figure 12. Photograph of thesis area east of East Eagle Creek. Note thick stands of forest.
undertaken to resolve ambiguities discussed previously regarding unconformities. Each of the units described have metamorphic equivalents. The non-metamorphic units are first discussed followed by an examination of the metamorphic units.

**Clover Creek Greenstone**

No confirmed outcrops of the Clover Creek Greenstone were found within the thesis area. Rocks resembling descriptions of the Clover Creek Greenstone were found near the summit of Truax Mountain. However, the paucity of samples of this lithology precludes breaking it out as a mappable unit. An alternative interpretation may be that this lithology represents a volcanic flow within the younger Triassic Lower Sedimentary Series such as that described by Vallier (1977).

**Lower Sedimentary Series**

Extensive exposures of the Lower Sedimentary Series occur throughout the thesis area. Examples of this unit that most closely match the type description of Smith and Allen (1941) are found bordering the west flank of East Eagle Creek. Particularly good outcrops of this unit occur in the core of the northwest plunging anticline in the southwest portion of the map area along the road leading to Bradley Creek mine (Plate I). Good exposures of the contact between the Lower Sedimentary Series and the overlying Martin Bridge Formation can also be found along this road. The base of the unit is not exposed. Vast outcrops of the Lower Sedimentary Series occur on Krag Peak which led Prostka (1962) to propose that the unit be renamed "Krag Peak Formation".

Carnahan (1962) mapped in the western portion of the thesis area and on the basis of lithologies divided the Lower Sedimentary Series into two laterally correlative units: a greenstone conglomerate unit and a siltstone-sandstone unit. Both of these were observed in the thesis area but were not distinguished as separate map units. The greenstone conglomerate is poorly sorted and consists of pebbles and cobbles of predominantly volcanic rock fragments with a lesser amount of metamorphic and intrusive clasts (Figure 13). The conglomerate is clast supported with subround to round pebbles and cobbles. Carnahan (1962) quantified the clast composition as follows: Volcanic fragments—very abundant, chert—common, limestone—less common, mudstone—siltstone—less common, metamorphic fragments—rare, granitoid—rare (Figure 14).
Figure 13. Triassic Lower Sedimentary Series greenstone conglomerate unit. Location: Section 21 Township 6S Range 44E, west side of East Eagle Creek.

Figure 14. Photomicrograph of Triassic Lower Sedimentary Series greenstone conglomerate. Note large, rounded volcanic clasts and subhedral plagioclase grains. Field of view is approximately 26mm, crossed nicols.
Figure 15. Triassic Lower Sedimentary Series siltstone-sandstone unit. Note laminations and distinctive maroon color. Location: Section 32 Township 6S Range 44E on road to Bradley Creek Mine.
The most distinctive unit within the Lower Sedimentary Series is the siltstone-sandstone unit. It consists of laminated, maroon and green siltstones and sandstones (Figure 15). Abundant sedimentary structures found within this unit include cross bedding, graded bedding, and convoluted bedding. Occurrence of the unit is restricted. The northwest plunging anticline exposed in the vicinity of East Eagle mine is cored by this unit. It also outcrops on the west flank of Krag Peak and in portions of the Sparta quadrangle to the south.

Carnahan (1962) noted that both the greenstone conglomerate and siltstone-sandstone occur near the top of the formation suggesting that these two units may interfinger and thus be laterally equivalent. In the southwestern portion of the area the siltstone-sandstone is in contact with the Martin Bridge Formation while in the south central portion of the area, the greenstone conglomerate is at the contact. It should be noted, however, that north of Gold King Creek there is a dramatic increase in the metamorphic grade, and it is thus not clear what inferences can be made regarding lateral stratigraphic relations across this creek.

The contact between the Lower Sedimentary Series and the overlying Martin Bridge Limestone was examined carefully and found to be conformable. Evidence supporting this conclusion includes parallel attitudes of both units across the contact and gradational lithologic changes from one formation to the other. This contact is particularly well exposed along the road to Bradley Creek mine. Here the siltstone-sandstone of the Lower Sedimentary Series grades upward from a laminated maroon sandstone into a green to light green calcareous argillite, and finally into a gray limestone with thin, interbedded tuffaceous units which mark the base of the Martin Bridge Formation (Figure 16). The contact is also exposed in Section 21 Township 6S Range 44E between Papoose and Gold King creeks at an elevation of approximately 4975 feet MSL. Here massive, unrecrystallized limestone occurs containing clasts of volcanic fragments and chert similar to those in the subjacent Lower Sedimentary Series.

**Martin Bridge Limestone**

Perhaps the most easily recognizable unit in the study area is the limestone/marble of the Triassic Martin Bridge Limestone. This unit is a thick deposit that is found throughout the Seven Devils terrane. In the study area, good exposures occur along the road to the Bradley Creek mine, on the ridges above the west bank of East Eagle Creek, on the
Figure 16. Contact between Triassic Lower Sedimentary Series (left) and Triassic Martin Bridge Limestone (right). Note gradational, conformable nature of contact; bedding dips steeply to the right. Location: Section 32 Township 6S Range 44E on road to Bradley Creek Mine.
Figure 17. Spectacular outcrop of Triassic Martin Bridge Limestone marble. Narrow, linear brown features are Tertiary basalt feeder dikes. Location: east face of Granite Cliff.
western flank of Krag Peak and Truax Mountain, and in the valleys of Kettle Creek and Twin Canyon. Spectacular exposures of Martin Bridge marble occur on the east face of Granite Cliff (Figure 17). Present day distribution of the Martin Bridge Formation is controlled by tectonic movements subsequent to deposition as evidenced by repetition of the unit across East Eagle Creek and its truncation on the west flank of Krag Peak.

The Martin Bridge Limestone consists of argillaceous limestone and calcareous shale interbedded with massive limestone. Prostka (1962) noted that the bulk of the formation is thin bedded with massive units comprising only about one third of the sequence. However, in the study area, massive limestone and marble predominate with thinly bedded limestone occurring only in the southwest portion of the area.

Two types of massive limestones occur; one consists almost entirely of fossil fragments of coral, gastropod, and peclmatazoan stems; the other is a massive limestone breccia and conglomerate composed of limestone clasts in a calcareous matrix (Prostka, 1962). Thin bedded limestones are argillaceous with occasional calcareous shale and thin tuffaceous interbeds. They occur in the lower Martin Bridge near the contact with the Lower Sedimentary Series. Locally these thin bedded units contain large amounts of carbonaceous material with interspersed pyrite suggesting an anoxic environment. Because of tectonic disruption, the thickness of the Martin Bridge Limestone has not been determined in the area; elsewhere it has been reported to be 500 m thick (Vallier, 1977).

**Hurwal Formation**

The Hurwal Formation conformably overlies the Martin Bridge Limestone. The contact between the two is gradational, and the base of the Hurwal is taken to be the first argillite bed above the upper limestone of the Martin Bridge Limestone (Prostka, 1962). Distribution of the Hurwal Formation within the study area is limited to west of East Eagle Creek. South of Gold King Creek, the rocks of this formation are non-metamorphic or of very low grade metamorphic grade; to the north they are metamorphosed to the amphibolite facies. The Hurwal Formation consists of laminated carbonaceous and pyritic siltstones interbedded with shale and thin beds of limestone. Sedimentary structures include cross bedding and graded bedding. Since the top of the Hurwal is not exposed and truncated by the Miocene Columbia River Basalt, the thickness of the unit cannot be measured in the thesis area. Elsewhere, the estimated thickness varies from between 1500 to 6800 feet (Smith and Allen, 1941; Ross, 1938; Carnahan, 1962; Prostka, 1962).
Interbedded within the siltstones and argillites are thick beds of limestone and conglomerate. The conglomerates are massive and contain clasts that range in size from granules to cobbles and boulders. Clasts are composed of limestone, volcanic fragments, chert, argillite, and quartz granules. Good exposures of this lithology occur in sections near Okbrien Creek.

**Metamorphic Rocks**

The distribution of metamorphic rocks in the thesis area is shown in Plate II. This map indicates that most of the rocks have been metamorphosed to the greenschist facies. A smaller proportion of the area has undergone amphibolite grade metamorphism. There is evidence to suggest that locally granulite facies metamorphism has occurred. On the basis of textural evidence it is clear that metamorphism in this area is the result of regional, dynamothermal metamorphic processes; the effects of thermal overprinting due to later plutonism are minimal.

Approximately 75% of the rocks in the area are metamorphic. The remaining 25% are non-metamorphic or of very low grade. Presumably, higher grade metamorphic rocks formed at greater depths than the low grade and non-metamorphic rocks. The strikingly abrupt juxtaposition of metamorphic and non-metamorphic units in such a small area indicates the existence of significant tectonic boundaries.

**Greenschist Facies**

Greenschist facies rocks that occur throughout the thesis area typically have mineral assemblages that consist of chlorite-actinolite-albite-quartz. Biotite is found coexisting with this assemblage in several areas as indicated on Plate II. Minor constituents include calcite, magnetite, sphene, and apatite. These rocks are the metamorphic equivalent of the Lower Sedimentary Series and Hurwal Formation. Determination of to which stratigraphic unit the metamorphic rocks belong is based on relict textures and relationship to the marble units of the Martin Bridge Limestone.

**Amphibolite Facies**

Amphibolite facies rocks occur in the northwestern portion of the area (Plate II).
These rocks contain hornblende, biotite, quartz, and plagioclase of composition An$_{15-30}$ (oligoclase). Due to the difficulty in distinguishing hornblende from actinolite by petrographic means, some of the greenschists may actually be amphibolites. Thus the occurrence of amphibolite grade rocks may actually be more widespread than that shown on Plate II.

One sample from within the amphibolite terrane shown on Plate II contains a unique paragenesis. The rock has a well developed metamorphic segregation/ foliation and consists of layers of clinopyroxene (diopside?) alternating with plagioclase and quartz layers (Figure 18). A minor amount of embayed hornblende is present; biotite is conspicuously absent. This assemblage of mostly anhydrous minerals is indicative of high grade metamorphism--possibly granulite facies. However, consideration should be given to the possibility that these rocks which appear to be granulites may have started as an anhydrous protolith and actually formed under lower grade metamorphic conditions. It is also possible that this rock has been overprinted by contact metamorphism; the tectonite textures may be inherited (mimetic recrystallization). To resolve these uncertainties, far more work on field relations and detailed geochemical analyses needs to be done.

Based on these assemblages, a rough estimate of the temperature at which metamorphism occurred is possible. Winkler (1979) indicates that greenschist facies conditions span a temperature range between 350 and 500 degrees Celsius. Amphibolite facies temperatures are between 500 and 600 degrees, and granulites form at temperatures in excess of 600 degrees. Therefore, it is evident that significant movement has occurred across tectonic boundaries resulting in the juxtaposition of non-low grade metamorphic rocks with medium-high grade metamorphic rocks.
Figure 18. Photomicrograph of Granulite(?) from Triassic-Jurassic Hurwal Formation. Note tectonite texture and metamorphic segregation of pyroxene layers and quartz-plagioclase layers. Location: Cirque wall southeast of Lookingglass Lake. Field of view is approximately 3.35 mm, crossed nicols.
STRUCTURAL GEOLOGY

The structural geology of the thesis area was examined through a variety of techniques. First, the stratigraphy described in the previous section was established in order to provide recognizable form surfaces. Geometrical analysis of structures in the area was accomplished by field mapping, detailed measurement of mesoscopic structures including minor fold axes, axial planes, cleavages, metamorphic foliations and lineations. This type of analysis is typically used to produce a three dimensional picture of the structure of interest (Hobbs and others, 1976). However, the results obtained in the present study are necessarily interpretative due to the discontinuity of rock exposure throughout the area. Nevertheless, the analysis resulted in the reliable establishment of the style, number, and timing of deformations that occurred in the thesis area.

Finally, a kinematic analysis of structural movements was undertaken in order to determine transport directions. This was accomplished by determining principal strain axes orientations for selected samples. Initially the intent of the study was to apply finite strain methods (Ramsay and Huber, 1983) to deformed fossils, strained pebbles, and other objects. However, few suitable finite strain markers were found. As a result, petrofabric strain analysis was done on the recrystallized metamorphic rocks. Preferred crystallographic orientation fabrics for both calcite and quartz were used to determine principal strain axis orientations.

Two phases of Mesozoic penetrative deformation were found to exist in the area. Each phase was clearly identified from megascopic, mesoscopic, and microscopic structural analysis. Evidence of each phase of deformation is not found consistently throughout the thesis area. Therefore it is necessary to evaluate data from every scale using relationships which relate one area to the next in order to understand the deformational history of this region.

Megascopic Structure

Megascopic structural features were delineated by detailed geologic mapping. Field mapping was done at a scale of 1:21,120 on a basemap blown up from the 15 minute USGS Eagle Cap, Oregon quadrangle (Plate I). Geologic contacts were located by both compass triangulation and an aneroid altimeter. Aerial photographs proved to be of limited
use due to the great density of forest cover.

In addition to lack of outcrop, there were other problems encountered which hindered an analysis of the structure of the Mesozoic rocks in the study area. These included the disappearance of important geologic contacts beneath the widespread deposits of Tertiary basalt and Quaternary surficial deposits. Also, voluminous Cretaceous intrusive rocks truncate earlier structures and limit the area over which they can be found. Finally, as discussed below, Cenozoic faulting has resulted in the complex juxtaposition of different units making it difficult to reconstruct Mesozoic deformational events.

Mesozoic Structure

Megascopic manifestations of the two pre-Tertiary deformational events consist of 1) a first phase ($F_1$) fold set with axial plane and fold axis trending in a northeasterly direction, and 2) a second phase ($F_2$) cross-fold that trends northwesterly.

First Phase Structures The $F_1$ fold set is best seen in east-west cross sections from west of East Eagle Creek to the Cornucopia stock near Crater Lake (Plate III). These folds comprise a synclin-anticline-syncline-anticline set which have been rearranged by Tertiary normal faulting. The most prominent megascopic $F_1$ fold is a large anticline that forms Krag Peak. This structure is characterized by a northeasterly trending axial plane that dips steeply to the southeast. The fold is overturned to the northwest as indicated by the overturned contact between the Lower Sedimentary Series and the Martin Bridge Limestone on the west flank of Krag Peak (Figure 19). The map pattern suggests that the megascopic fold axis is generally sub-horizontal trending in a northeasterly direction because there is no closure of the anticline within the map area. Mapping to the north indicates that the nose of this large feature occurs approximately seven miles away (Weiss and others, 1976). Wetherell (1963) mapped the area adjacent to and east of the study area where he described the eastern, upright limb of this fold.

Structural trends are primarily the result of this first deformational phase. This is evidenced by the predominance of northeasterly striking units which dip steeply to the southeast. These trends have locally been affected by the F-2 cross folding event.

Second Phase Structures A second phase fold set has developed approximately orthogonal
Figure 19. West flank of Krag Peak. Note overturned contact between Triassic Martin Bridge Limestone marble and Triassic Lower Sedimentary Series. Note that Martin Bridge Limestone is cut-out due to faulting.
to the first phase. The most prominent F₂ structure is a northwesterly plunging anticline that occurs in the southwestern portion of the thesis area near the Bradley Creek Mine. This fold is slightly overturned and has an axial plane that strikes to the northwest and dips steeply toward the southwest. It is cored by non-metamorphic Lower Sedimentary Series and closes to the northwest. This anticline is situated north and adjacent to the Eagle Creek synclinorium described by Prostka (1962) in the Sparta quadrangle.

**Plutonic Structures** The post tectonic, Jurassic-Cretaceous Wallowa batholith and associated Cornucopia stock occur in the northwestern and southeastern portion of the thesis area, respectively. Evidence indicating that these plutons intruded the area subsequent to the Mesozoic F₁ and F₂ events can be found in areas where the contact of the pluton and country rock trends orthogonally to the dominant structural grain of the region (e.g. on the east flank of Trux Mountain). It is apparent that the pluton is discordant since it truncates both F₁ and F₂ structures. Additional evidence substantiating the post tectonic nature of the pluton is found at its margins where xenoliths exhibiting tectonite fabric are found (Figure 20). This is clear evidence that the pluton intruded subsequent to the penetrative deformation. Finally, a reconnaissance of the internal fabric within the pluton indicated that there is no consistent foliation that would indicate it was involved in a deformation.

**Post-Cretaceous Structure**

One of the most striking features of the geologic map is the distribution of metamorphic versus non-metamorphic rocks. Metamorphic rocks ranging from lower greenschist to upper amphibolite facies are found throughout most of the area. Only in the southwest portion of the area are the rocks not metamorphic. This area is bounded on the north by Gold King Creek and on the east by East Eagle Creek. As shown on the geologic map and accompanying cross sections, these boundaries are interpreted to be high angle normal faults. This conclusion is supported by the presence of prominent straight drainages.

A north-south cross section across the fault that parallels Gold King Creek indicates that the base of the Tertiary Columbia River Basalt on either side of the fault is approximately at the same level. This suggests that movement along this fault pre-dates the Miocene age of the basalt. Since the fault offsets Mesozoic units as young as Jurassic, the
Figure 20. Tectonite xenoliths within Cretaceous granodiorite. Location: Section 12 Township 6S Range 44E near Crater Lake.
age of the faulting is constrained between the Jurassic and Miocene. There is no evidence to indicate the age of the fault interpreted to underlie the East Eagle Creek other than it offsets Mesozoic units.

The western margin of the Cornucopia stock above the east bank of East Eagle Creek is interpreted to be a high angle normal fault that strikes north-northeasterly and dips steeply to the east. Its existence is supported by the repetition of the Martin Bridge Limestone along Kettle Creek and the linear contact of the Cornucopia stock with the country rock. Additional evidence for a fault is a zone of shearing and mineralization in the Kettle Creek. Since this fault truncates Cretaceous plutonic rocks it is interpreted to be a Cenozoic structure.

A northwesterly trending normal fault has been interpreted in the southeastern portion of the area. This fault forms the contact between intrusive rocks of the Cornucopia stock and Tertiary basalts downthrown to the southeast. Previous workers interpreted this fault to be a range front fault and have extended it across East Eagle Creek and out of the map area (Taubeneck, 1964; Walker, 1979). However, there is no indication that the fault extends any further than shown. None of the other faults on the map (Plate I) are exposed. Their occurrence is solely based on interpretation and they will be discussed in a subsequent chapter.

Angular Unconformity

A prominent angular unconformity is apparent from the geologic map at the contact between the Tertiary Columbia River Basalt and the Triassic Martin Bridge Formation. Marble with a steeply dipping foliation is overlain by the sub-horizontal flows of basalt with vertical columnar jointing (Figure 21). This unconformity is depicted on the cross sections of Plate III.

Mesoscopic Structure

Data on a mesoscopic scale provided the means to perform a geometrical analysis of the structural geology of the area. Mesoscopic structural data were obtained from minor folds: bedding, fold axes, axial planes, cleavages, and bedding-cleavage intersections. These data were plotted on the lower hemisphere of an equal-area stereographic projection (Schmidt net) and analyzed after methods described by Turner and Weiss (1963).
Figure 21. Angular unconformity between Triassic Martin Bridge Limestone marble and horizontal flows of Tertiary Columbia River basalt. Location: Section 15 Township 6S Range 44E, north of Kettle Creek.
F₁ Deformation

F₁ folds are found in all of the pre-Tertiary units throughout the field area (Figures 22-24). They are generally tight to isoclinal folds that have class 1C geometry (Ramsay, 1967) (Figures 22a, 22b). Chevron folds are developed in the Martin Bridge Limestone (Figure 23a, 23b) while flattened isoclinal folds can be found in quartz-plagioclase layers in the metamorphic Lower Sedimentary Series (Figure 25).

Close examination of cleavage associated with both folding events (F₁ and F₂) and fold geometries demonstrated that cleavage from both phases can be considered axial planar (Figure 24). Cleavages found in the absence of minor fold closures were discriminated on the basis of cleavage style, orientation, and cross-cutting relationships with other cleavage.

In general, first phase cleavage (S₁) is characterized as being a close spaced fracture cleavage in sedimentary units (Figure 26) or penetrative, foliation in metamorphic rocks. Metamorphic foliation includes a prominent schistosity in phyllosilicate rich rocks and gneissic layering in marbles and higher grade metamorphic rocks (Figure 24). S₁ is found throughout the area.

The development of cleavage appears to be controlled in part by lithology. In particular, those units that have an abundance of phyllosilicate constituents exhibit cleavage more often than phyllosilicate poor ones and coarse-grained rocks (Figure 27). Cleavage is also more widespread in the more highly strained metamorphic rocks than in their sedimentary counterparts.

F₂ Deformation

F₂ folds are rare in the field area. They are generally open to close folds of class 1B and 1C geometry of Ramsay (1967). In the southwestern portion of the area, F₂ folds demonstrably refold F₁ folds and S₁ surfaces in the Martin Bridge Limestone.

Second phase cleavage (S₂) is less prominent and widespread than S₁ cleavage. S₂ occurs primarily as a fracture cleavage in both metamorphic and non-metamorphic rocks (Figure 27). Only locally is S₂ developed as a weak crenulation cleavage.
Figure 22a. \( F_1 \) isoclinal fold in Triassic Martin Bridge Limestone marble. Location: West flank of Krag Peak, south bank of The Box.

Figure 22b. \( F_1 \) tight fold in Triassic-Jurassic Hurwal Formation. Location: Intersection of Snow Creek and East Eagle Creek, north of thesis area.
Figure 23a. F₁ fold in Triassic Martin Bridge Limestone. Location: Section 29 Township 6S Range 44E on road to Bradley Creek Mine.

Figure 23b. F₁ fold in Triassic Martin Bridge Limestone. Location: Section 29 Township 6S Range 44E on road to Bradley Creek Mine.
Figure 24. Flattened $F_1$ isoclinal folds (circled) in metamorphic Triassic Lower Sedimentary Series. Location: Section 28 Township 6S Range 44E north bank of Twin Canyon.
Figure 25. $F_1$ folds in calcite layers within Triassic Martin Bridge Limestone marble. Note axial planar $S_1$ surfaces.
Figure 26. Well-developed, close spaced fracture cleavage in Triassic Martin Bridge Limestone. Location: Section 29 Township 6S Range 44E, north of Bradley Creek Mine.
Figure 27. $S_2$ fracture cleavage in limestone interbed between conglomeratic units. Note absence of cleavage in coarser grained conglomerates. Location: Section 31 Township 6S Range 44E, east bank of Bradley Creek.
Fabric Analysis

Examination of the data indicated the existence of three structural domains (Figure 28). Domain 1 is located west of East Eagle Creek—an area that clearly shows the effects of the \( F_2 \) event. Domains 2 and 3 are east of East Eagle Creek, north and south of Kettle Creek, respectively. Structural trends in these areas appear to be related to \( F_1 \).

In the southwestern portion of the thesis area (Domain 1), poles to bedding planes form a diffuse south-southwesterly plunging maximum (Figure 29a). \( L_1 \) lineations and fold axes form a diffuse east-west trending girdle containing a steep elongate north-south trending maximum (Figure 29b). A well-defined steep northeasterly striking \( S_1 \) cleavage is developed (Figure 29c), which is cross cut by a steeply southerly dipping \( S_2 \) cleavage (Figure 29d).

The intersections of the mean bedding plane with the mean \( S_1 \) and \( S_2 \) cleavages are assumed to be the mean orientations of \( \beta_1 \) and \( \beta_2 \), respectively (Figure 29a) although the orientation of \( \beta_1 \) has no kinematic significance due to rotations during the second phase. The elongate north-south trending maximum of \( L_1 \) lineations may be the result of the first deformation, or it is caused by rotation about \( \beta_2 \). The east-west trending \( L_1 \) lineations are not understood; it is possible that these have been misinterpreted in the field and are in reality \( L_2 \) lineations.

In the northeast (Domain 2) poles to bedding indicate a fairly uniform northeasterly strike with steep dips to the northwest and southeast. There is, however, a hint of a girdle that trends northwesterly with a moderate dip to the southwest (Figure 30a). The pole to this girdle distribution plunges moderately to the northeast. It is considered to be a \( \beta_1 \) axis since the domain is on the limb of an \( F_1 \) fold. There is also a hint of a small-circle girdle with a steep (\( \beta_2 \)?) axis. Lineations in this region are distributed in a vertical northeasterly trending girdle (Figure 30b). The orientation of both \( S_1 \) and \( S_2 \) cleavages are similar to that in Domain 1 (Figure 30c, 30d). The trend of \( S_2 \) is more west-northwesterly and dips steeply to the southwest and northeast. Note that the orientation of \( S_1 \) is essentially parallel to \( S_0 \) indicating that \( F_1 \) is isoclinal.
Figure 28. Map showing location of domains defined for structural analysis.
Figure 29a. Domain 1, filled circles = poles to $S_0$. Dashed line = interpreted great circle girdle. Open Circle = $\beta_2$ fold axis. (Equal-area, lower-hemisphere projection).

Figure 29b. Domain 1, open squares = $L_1$ intersection lineation; filled squares = $L_2$ intersection lineation; open circles = measured $\beta_1$ axes. (Equal-area, lower-hemisphere projection).
Figure 29c. Domain 1, poles to $S_1$. (Equal-area, lower-hemisphere projection).

Figure 29d. Domain 1, poles to $S_2$. (Equal-area, lower-hemisphere projection).
Figure 30a. Domain 2; filled circles = poles to $S_0$, dashed line = interpreted great circle distribution of $S_0$ poles, open circle = interpreted $\beta_1$ axis. (Equal-area, lower-hemisphere projection).

Figure 30b. Domain 2; open squares = $L_1$ intersection lineations. (Equal-area, lower-hemisphere projection).
Figure 30c. Domain 2; poles to $S_1$. (Equal-area, lower-hemisphere projection).

Figure 30d. Domain 2; poles to $S_2$. (Equal-area, lower-hemisphere projection).
The intersection of the mean $S_0$ and $S_1$ with $S_2$ is subvertical; this intersection is interpreted as the $\beta_2$ axis. The $\beta_2$ axis can explain the small circle configurations of $S_0$ and is not inconsistent with the $S_1$ distribution.

In the southeast (Domain 3) the distribution of $S_0$ and $S_1$ (Figures 31a, 31c) indicate that the $F_1$ folds are not isoclinal; most of the bedding planes dip westerly whereas the cleavage is subvertical or dips steeply to the southeast. The $L_1$ lineations (Figure 31b) occur in a north-south trending girdle, subparallel to the mean bedding plane. The intersection of the mean $S_0$ and $S_2$ (Figures 31c and 31d) is $\beta_2$ plunging moderately to the northwest. The $\beta_1$ and $\beta_2$ axes are plotted in Figure 31a. Although the number of $S_0$ measurements is insufficient for statistical analysis, the data are not inconsistent with rotation about $\beta_1$ and $\beta_2$.

$S_2$ surfaces form a point maximum in all three domains; they strike northwesterly and dip steeply. $S_2$ is approximately orthogonal to $S_1$. These regional data do not indicate any obvious refolding of the $S_1$ surfaces by the $F_2$ event. However, locally the $\beta_2$ axis is parallel to $S_1$ as shown in Figure 32, where structural data are plotted from one outcrop in which a mesoscopic $F_2$ fold is observed to refold $F_1$ folds. The figure indicates that $S_1$ surfaces are folded into a great circle girdle about a moderately plunging, west-northwesterly trending $\beta_2$ axis. $S_2$ surfaces form a point maximum that falls on the girdle of $S_1$ surfaces. $\beta_1$ axes have been rotated in a small circle girdle about the $\beta_2$ axis.

Microscopic Structure

Structural analysis of microscopic features included textural and microfabric studies. Careful examination of textural features provided important information on the development of metamorphic tectonite fabric and its relationship to the different folding phases. Textural analyses also provide the basis for selection of samples for petrofabric analysis. Petrofabric analyses of selected specimens enable the establishment of principal
Figure 31a. Domain 3; poles to $S_0$.

Figure 31b. Domain 3; $L_1$ intersection lineations.
Figure 31c. Domain 3; poles to $S_1$.

Figure 31d. Domain 3; poles to $S_2$. 
Figure 32. Mesoscopic fabric data from F_1 folds refolded by F_2 fold. open triangles = poles to S_1, filled triangles = poles to S_2, open circles = \( \beta_1 \) axes, filled circle = pole to great circle (\( \beta_2 \)) of S_1 poles.
strain orientations.

Textural Analysis

The thesis area contains metamorphic tectonites of varying composition, grade, and texture. In general, four types of tectonites were observed: 1) low grade metasedimentary and metavolcanic rocks with relict texture and cleavage, 2) foliated marble of the Martin Bridge Limestone, 3) medium grade metamorphic rocks with a prominent foliation and sometimes an incipient crenulation cleavage--lacking relict textures, 4) medium to high grade metamorphic rocks with well-developed metamorphic segregation.

The low grade metamorphic rocks are characterized by an incipient first phase cleavage (Figure 33). In the volcanic rocks, a reticulate pattern of cleavage has developed within the fine-grained matrix. It is lined with opaque materials and in some cases bends around porphyroclasts of plagioclase, calcite, and quartz. In other cases, the cleavage is straight and cuts off the porphyroblasts. This type of cleavage is termed anastamosing and is common in low grade rocks with an abundance of silt and sand sized constituents (Borradaile and others, 1982). This type of cleavage can be interpreted as having formed by dissolution. The anastamosing cleavage is axial planar to micro-folds developed in $S_0$ surfaces (Figures 34a, 34b). As metamorphic grade increases, biotite grains oriented parallel to $S_1$ are observed. This type of texture marks the incipient development of a well-defined metamorphic foliation.

Marble found throughout the field area is the result of regional metamorphism of the Martin Bridge Limestone. The marble is composed predominantly of granoblastic-elongate calcite grains with minor amounts of muscovite and opaques (Figures 35a, 35b). Many of the samples examined exhibit granoblastic-polygonal texture, indicating that post tectonic recrystallization has occurred (Spry, 1969). Foliation in the marble is defined by the elongate calcite grains and is always parallel to $S_1$. None of the marble samples examined exhibit evidence of a penetrative $S_2$ surface.

As the metamorphic grade increases the textural manifestation of the two deformational events becomes more evident. $S_1$ foliation is defined by granoblastic elongate grains of quartz, plagioclase, amphibole, and phyllosylicates such as chlorite and biotite. In at least one sample, flattened calcite augen define the foliation (Figure 36).
Figure 33. Solution cleavage ($S_1$) in low-grade metamorphic Triassic Lower Sedimentary Series. Field of view approximately 3.35 mm, nicols uncrossed.
Figure 34a. $S_1$ solution cleavage axial planar to folded $S_0$ layer. Field of view is approximately 26 mm, uncrossed nicols.

Figure 34b. $S_1$ solution cleavage axial planar to folded $S_0$ layer. Field of view is approximately 3.35 mm, crossed nicols.
Figure 35a. Calcite marble from Triassic Martin Bridge Limestone in contact with
greenschist from Triassic Lower Sedimentary Series. Field of view is approximately 26
mm, crossed nicols.

Figure 35a. Calcite marble from Triassic Martin Bridge Limestone. Field of view is
approximately 3.35 mm, crossed nicols.
Figure 36. Flattened calcite augen parallel to $S_1$ in metamorphic Triassic Lower Sedimentary Series. Plane of thin section is parallel to $Y$-$Z$ principal strain plane. Field of view is approximately 3.35 mm, crossed nicols.
Figure 37a. Euhedral pyrite crystals with straight fibers of quartz, biotite, and chlorite in pressure shadow. Same sample as Figure 35. Plane of thin section parallel to X-Z strain plane. Field of view approximately 3.35 mm, nicols crossed.

Figure 37b. Close up of pressure fringe from same thin section as Figure 36a. Field of view is approximately 0.88 mm, uncrossed nicols.
Chlorite fringes in the pressure shadows adjacent to euhedral pyrite cubes parallel the X principal strain axis within the \( S_1 \) foliation plane (Figures 37a, 37b).

Textural evidence for the second phase of deformation is rarely found. When present, \( S_2 \) surfaces are defined by a crenulation cleavage. As shown in Figures 38a and 38b, \( S_2 \) is parallel to the axial planes of kinked \( S_1 \) surfaces. \( S_2 \) appears to be an incipient differentiated crenulation cleavage with biotite rich-quartz poor domains developed with the biotite grains rotated slightly towards the direction of the second phase foliation.

Several examples of interkinematic mineral growth occur. These consist of plagioclase poikiloblasts containing helicitic trails of the \( S_1 \) foliation (Figure 39). Flattening of the \( S_1 \) foliation around the poikiloblast was in response to the second phase of deformation. A second, non-regional metamorphic event is suggested by other textural evidence. Post tectonic sheaves of amphibole oriented with their long axes oblique to the dominant foliation are found in several samples (Figure 40). Annealing textures are found in a few samples from locations proximal to the Wallowa batholith (Figure 41). These textures indicate that the rocks have been heated sufficiently to allow ionic mobility in the absence of stress. This resulted in recrystallization of the minerals to a strain free configuration (Spry, 1969).

Textural evidence indicates that two metamorphic events affected the area. The first (\( M_1 \)) is contemporaneous with \( F_1 \), and the second is a thermal event related to the intrusion of the Wallowa batholith. The peak of \( M_1 \) metamorphism occurred during the \( F_1 \) folding event and resulted in the development of \( S_1 \) foliation. \( F_2 \) deformation resulted in the flattening of the \( S_1 \) foliation around interkinematic porphyroblasts and rotation of pre-existing first phase minerals into an \( S_2 \) foliation. The subsequent thermal metamorphic event affected rocks proximal to the Wallowa batholith and Cornucopia stock. It resulted in slight recrystallization of pre- and syn-tectonic minerals and crystallization of post tectonic minerals.

Petrofabric Analysis

It has long been known that certain minerals in metamorphic tectonites display strong preferred crystallographic orientations. Researchers have shown that these
Figure 38a. Incipient differentiated $S_2$ crenulation cleavage in biotite-quartz-plagioclase schist unit of the Triassic Lower Sedimentary Series. Field of view is approximately 3.35 mm, uncrossed nicols.

Figure 38b. Close up of $S_2$ crenulation cleavage. Note kinked biotite. Field of view is approximately 0.88 mm, crossed nicols.
Figure 39. Plagioclase poikiloblast with helicitic trails parallel to $S_1$. Note flattening of $S_1$ around poikiloblast. Field of view is approximately 3.35 mm, crossed nicols.
Figure 40. Photomicrograph of post tectonic amphibole "sheaves" in metamorphic Lower Sedimentary Series. Field of view is approximately 3.35 mm, crossed nicols.

Figure 41. Photomicrograph of annealed quartz grains in metamorphic Jurassic Hurwal Formation close to Wallowa batholith. Field of view is approximately 3.35 mm, crossed nicols.
preferred orientations form distinctive patterns that relate to the strain history of the sample. Hobbs and others (1976) cited two primary mechanisms for the development of preferred orientations. The first occurs at low temperatures or high strain rates without recrystallization. These conditions may result in a preferred orientation due to rotation of inequant grains or to intracrystalline slip. The second mechanism occurs under conditions of widespread recrystallization and may be associated with that process. Preferred orientations of minerals syntectonically recrystallized have been related to the principal stresses, strains, and to the kinematics of deformation.

Petrofabric analysis was performed on selected samples. Thin sections were prepared by cutting the sample perpendicular to the tectonic foliation and parallel to the strike of the foliation. Crystallographic axis orientations of quartz and calcite were measured using the universal stage microscope. All c-axes were plotted on a lower-hemisphere, equal-area projection and counted as percent per one percent area using a Kalsbeek counting net. These values were then contoured and the results are shown in Figures 43 through 48.

Figure 42 shows the location of samples chosen for petrofabric analysis of lattice preferred orientations. Ten samples were chosen—seven with quartz and three with calcite. Samples were chosen on the basis of exhibiting tectonite textures and containing a sufficient number of either quartz or calcite crystals.

**Quartz Preferred Orientations.** Seven samples from all three domains were selected for quartz fabric analysis: four from Domain 1, one from Domain 2, and two from Domain 3. All of the samples except one are biotite-quartz-plagioclase schists with varying amounts of amphibole; one sample is a pyroxene-plagioclase-quartz gneiss. Each has a prominent S₁ foliation and lacks relict textures due to extensive recrystallization. Only one sample has a well-developed lineation which is due to a crenulation cleavage.

Three different quartz c-axis fabric types were observed. The most common fabric is characterized as a peripheral girdle (Figures 43a-d). In addition, three of the samples with this fabric also have a point maximum developed approximately normal to the S₁ foliation (Figures 43a, b, d). In general, these fabrics exhibit orthorhombic symmetry with S₁ as a symmetry plane. The down dip direction of S₁ is also the center of a large c-axis pole free area.

Two samples display a diffuse crossed-girdle pattern with orthorhombic symmetry
Figure 42. Map showing location of samples chosen for petrofabric analysis.
Figure 43a. Sample 83-38. Equal-area, lower-hemisphere projection of 200 quartz c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; $S_1 = $ cleavage; great circle is the horizontal with $W = $ west and $N = $ north.

Figure 43b. Sample 83-44. Equal-area, lower-hemisphere projection of 200 quartz c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; $S_1 = $ cleavage; great circle is the horizontal with $E = $ east and $N = $ north.
Figure 43c. Sample 84-23-1. Equal-area, lower-hemisphere projection of 200 quartz c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; $S_1 =$ cleavage; great circle is the horizontal with $E =$ east and $S =$ south.

Figure 43d. Sample 84-131. Equal-area, lower-hemisphere projection of 200 quartz c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; $S_1 =$ cleavage; great circle is the horizontal with $E =$ east and $S =$ south.
Figure 44a. Sample 83-63a. Equal-area, lower-hemisphere projection of 200 quartz c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; \( S_1 \) = cleavage; great circle is the horizontal with E = east and S = south.

Figure 44b. Sample 84-136. Equal-area, lower-hemisphere projection of 200 quartz c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; \( S_1 \) = cleavage; great circle is the horizontal with E = east and S = south.
Figure 45. Sample 84-133-1. Equal-area, lower-hemisphere projection of 200 quartz c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; $S_1$ = cleavage; great circle is the horizontal with E = east and S = south.
(Figures 44a, 44b). Like the single-girdle fabrics, $S_1$ is a symmetry plane and intersects the center of a large pole free area. The remaining sample does not exhibit a particularly strong c-axis preferred orientation (Figure 45). However, there is a weakly defined pole free area centered over the $S_1$ foliation, and a weak small circle girdle is centered about the pole to $S_1$.

**Interpretation of Quartz Fabrics** Much work has been done on quartz c-axes preferred orientations. The results of experimental studies, theoretical computer simulations, and analyses of naturally deformed rocks can be compared with the microfabrics observed in this study. Most of the preferred orientation fabrics discussed in the literature are from metamorphic tectonites which have been completely recrystallized (Green and others, 1970). Preferred orientations can also form by translation glide (Tullis and others, 1973). All seven samples examined were comprised of inequant, polygonal grains with straight boundaries; this texture is indicative of syntectonic recrystallization (Green and others, 1970).

Quartz fabrics produced experimentally by syntectonic recrystallization are similar to those formed by translation glide. At low temperatures and high strain rates, a c-axis maximum develops parallel to the compression direction, and at high temperatures and low strain rates, crossed girdles of c-axes form intersecting at the intermediate principal strain or stress axis (Green and others, 1970). It is generally assumed that fabrics produced by syntectonic recrystallization are also related to the finite strain axes although other interpretations have been made (Kunze and Avé Lallemant, 1981).

Naturally formed c-axis patterns have been studied in prograde metamorphic terrane where it was observed that microfabrics change "...successively from random to peripheral girdles, to crossed girdles, to random to either very strong girdles or maxima" (Wilson, 1973). A crossed-girdle fabric is one of the most common fabric types observed. It consists of two great-circle girdles arranged in orthorhombic symmetry such that the $X$-$Y$ strain plane and $X$ strain axis are symmetry plane and axis, respectively (Sylvester and Christie, 1968), where $X \geq Y \geq Z$.

The relationship between patterns of preferred crystallographic orientations in quartz and principal strain axes has been examined in both experimentally and naturally deformed rocks (Sylvester and Christie, 1968; Green and others, 1970; Hara and others, 1973; Wilson, 1973; Tullis and others, 1973; Tullis 1977). In addition, preferred orientations
have been predicted by computer modelling for various strain states and deformation paths (Lister and others, 1977). In general, the center of a c-axis pole free area on a stereoplot is parallel to the orientation of the X axis of the finite strain ellipsoid. Since it is commonly accepted that metamorphic foliations form perpendicular to the Z principal strain axis, it is a simple matter to determine the orientation of the third principal strain axis, Y.

Preferred orientations may also be used to deduce deformation kinematics: "...symmetry principles form the soundest basis for correlating the tectonic fabric with the physical factors—stress, strain, movement picture, and so on—concerned in their evolution" (Turner and Weiss, 1963). Thus genetic links exist between symmetry of the final fabric, componental displacement, strain state, and stress state (Friedman and Sowers, 1970). Theoretical studies involving computer simulation of quartz c-axes support the symmetry principal. Lister and Hobbs (1980) simulated various combinations of glide systems and deformation paths and found characteristic fabrics associated with different strain states. In general, fabrics with axial and orthorhombic symmetry were associated with coaxial plane strain and/or flattening deformation. Under conditions of progressive simple shear, the fabric became monoclinic.

Three distinct quartz c-axis fabric types were observed in this study. Four of the samples displayed peripheral girdles centered about a large pole free area (Figures 43a-d). Three of the samples exhibit peripheral point maxima normal to the S₁ foliation plane (Figures 43a and d). A similar fabric was described by Wilson (1973) and has been interpreted to indicate that the X strain axis lies in the center of the pole free area and is contained in the S₁ foliation. A second fabric type, characterized as a crossed-girdle pattern has been observed in two samples (Figures 44a, 44b). This type of fabric has been described as occurring often in nature (Sylvester and Christie, 1968; Green and others, 1970; Wilson, 1973). Based on the results of these previous studies, it appears that the X strain axis lies within the S₁ foliation and parallels the center of the pole free area. The X axis is a symmetry axis and is orthogonal to the intersection of the crossed girdles. The third fabric type is found in only one sample (Figure 45). It consists of a weakly developed small-circle girdle about the pole to the S₁ foliation and a large relatively pole free area. The X strain axis is interpreted to lie within S₁ and is parallel to the center of the pole free area.

Analysis of all of the quartz fabrics indicates that the major principal compressive
Figure 46. Sample 84-63. Equal-area, lower-hemisphere projection of 200 calcite c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; $S_1$ = cleavage; great circle is the horizontal with W = west and N = north.
strain (Z) is normal to the cleavage plane and the major principal extensile strain (X) trends in the downdip direction of the S₁ foliation. The high degree of symmetry exhibited by all of the quartz fabrics indicates that they developed as a result of coaxial strain. The weakest quartz fabrics (Figures 44b and 45) are from samples that occur in the northwest portion of the map area. These rocks are in close proximity to the Wallowa batholith and the relative weakness in the preferred orientation fabric may be due to contact metamorphic effects. The influence of temperature on preferred orientation fabrics was described by Wilson (1973). He noted that in a region of prograde regional metamorphism, petrofabric pattern strength increased until high grade metamorphic conditions. At those high temperatures the fabrics became almost random similar to that in Figure 45.

**Calcite Preferred Orientations** Three samples of marble—one from each domain were chosen for analysis of calcite c-axis preferred orientation. Sample 84-63 is a calcite marble with a well-developed foliation. The distribution of calcite c-axes is characterized as a weak peripheral girdle (Figure 46).

Sample 84-98 from domain 2 is a fine-grained biotite schist with a prominent S₁ foliation that is partially defined by flattened calcite augen (Figure 47). A distinct mineral lineation is evident and trends in the down dip direction of the foliation plane. Euhedral pyrite grains are disseminated throughout the rock. These pyrites exhibit quartz-chlorite-biotite fibers growing in pressure shadows. The fibers exhibit no evidence of rotation. The orientation of the fibers within the pressure shadow is parallel to the X principal strain axis (Ramsay and Huber, 1983). C-axes of the augen display a strong preferred orientation. They are distributed in a peripheral girdle with two point maxima developed obliquely to the pole to S₁. A large pole free area is evident, the center of which is parallel to the X strain direction as determined from the pressure shadow fibers.

Sample 84-25 is a foliated calcite marble from Domain 3. The calcite c-axis fabric displays a large area that contains less than one percent per one percent area centered on the S₁ foliation (Figure 48). There is a diffuse crossed-girdle distribution with S₁ as a symmetry plane.

**Interpretation of Calcite Fabric** Calcite grains in experimentally deformed limestone and marble have distinctive preferred orientations. Calcite fabrics from experimental shear
Figure 47. Sample 84-98. Equal-area, lower-hemisphere projection of 200 calcite c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; $S_1 =$ cleavage; great circle is the horizontal with E = east and S = south.
Figure 48. Sample 84-25. Equal-area, lower-hemisphere projection of 200 calcite c-axes; contours at intervals of 1% per 1% area; shaded area is less than 1% per 1% area; \( S_1 \) = cleavage; great circle is the horizontal with \( E = \) east and \( N = \) north.
zones have been studied extensively. Experimental studies by Rutter and Rusbridge (1977) resulted in the observation that calcite grain-shape fabrics rotated more slowly than did the c-axis fabric during a two stage, coaxial deformational process. They concluded from these results that the grain-shape orientation reflected the finite strain orientation, while the lattice preferred orientation responded to rapid variations in the orientation of the principal stress or incremental strain axes. Similar results were found by Kern and Wenk (1982). Friedman and Higgs (1981) reported that during experimental simple shear deformation the calcite c-axis fabrics track the shortening axis. They also observed that the principal extension axis was perpendicular to a girdle of c-axes.

Two studies on naturally produced calcite fabrics were done on limestones from the Helvetic nappes of Switzerland which were deformed at temperatures of 400° C or less (Schmid and others, 1981). They found a strong tendency for axial symmetry around the position of a c-axis point maximum. In two cases the c-axis point maximum coincided with the pole to the tectonic foliation (X-Y strain plane). However, in the other two cases, the c-axis maximum was 30° and 40° from the pole to the foliation. In addition, it was found that the short axes of the individual calcite grains were parallel to the c-axis maximum, and not perpendicular to the foliation. This obliquity between the grain-shape fabric and the macroscopic foliation was attributed to a rotational strain path.

Similar results were obtained by Dietrich and Song (1984) from limestones also found in the Helvetic nappes. They found a regionally constant c-axis preferred orientation pattern typified by a point maximum subperpendicular to the strike of the mountain chain and parallel to the axis of shortening of the grain-shape fabric. An obliquity between the orientation of the macroscopic cleavage and the microfabric was also observed.

Two explanations are offered for the observed fabric. Rutter and Rusbridge's (1977) experimental results suggest that the c-axis maxima coincide with the orientation of the principal compressive stress during the final strain increments. The fabric is in effect attributed to a continuous shear deformation with noncoincidence between the incremental strain axes with the finite strain axes (Schmid and others, 1981). An alternative explanation is that the grain-shape fabric and the c-axis fabric reflect a finite deformation that postdates the formation of the cleavage. The grain-shape fabric would therefore be interpreted as an incipient second phase cleavage that cuts the macroscopic first cleavage. The grain-shape fabric would thus be related to the strain of the later event.

Interpretation of the calcite fabrics observed in this study is based on the foregoing and in one case, strain data from finite strain markers (Sample 84-98). Sample 84-63 has a
peripheral girdle distribution and pole free area centered over the $S_1$ foliation. The X principal strain axis is interpreted to parallel the pole to the girdle (Friedman and Higgs, 1981). Sample 84-98 displays a peripheral girdle with a large pole free area centered over the down dip direction of the foliation and lineation. The lineation is parallel to pressure fringes and is interpreted to be parallel to the X principal strain axis (Ramsay and Huber, 1983). Sample 84-25 displays a crossed girdle fabric. This fabric is similar to that described for quartz c-axes and can be interpreted to intersect in the Y strain axis. Thus the X strain axis is orthogonal to the Y axis and trends downdip within the $S_1$ foliation plane. As with the quartz fabrics discussed previously, the calcite fabric symmetry indicated that deformation occurred as irrotational strain.
DISCUSSION OF STRUCTURES

$F_1$ Deformation

Detailed analysis of structural features at all scales reveals the style, timing, and kinematics of deformation associated with the first phase ($F_1$) event. Manifestations of this deformation are found in megascopic map-scale structures, mesoscopic structural fabric elements, microscopic textures, and lattice preferred orientations. These data indicate that $F_1$ resulted in the formation of tight to isoclinal folds that trend northeasterly with northwesterly vergence. Shortening axes associated with these folds trend northwesterly.

Petrographic analysis of selected samples indicates that preferred orientations of quartz and calcite crystallographic axes developed as a result of syntectonic recrystallization during $F_1$. These fabrics can be used to determine the orientation of principal strain axes. Figure 49 shows for each sample the direction of the principal strain axes reoriented from the previous diagrams so that geographic coordinates are horizontal. The distribution of $X$ strain axes forms a northwesterly trending girdle with the majority plunging moderately-steeply to the southeast. These strain axis orientations combined with the northwesterly fold vergence observed in the field indicate that tectonic transport directions are to the northwest.

Few finite strain markers were found suitable for analysis. However, the fold morphologies and the symmetry of preferred orientation fabrics reveal information about the strain state. Class 1C folds described earlier are thought to form by flexural slip and progressive flattening (Ramsay, 1967). The quartz and calcite fabrics have orthorhombic symmetry suggesting that deformation is the result of coaxial strain, although it has been shown experimentally that quartz c-axis fabrics may have axial or orthorhombic symmetry even if the deformation was non-coaxial (Tullis, 1977). However, the coaxial nature of the deformation is demonstrated by the quartz- chlorite-biotite fibers found in pyrite pressure shadows.

There are no thrust faults exposed in the thesis area. However, the nature of the folding and penetrative structures throughout the area suggest that thrusting occurred during folding. It is assumed that thrust faults are present at depth. The coaxial nature of
Figure 49 Equal-area, lower-hemisphere projection of principal strain axes ($X\geq Y \geq Z$) determined from petrofabric analysis.
strain suggests that the structures exposed at the surface in this area are removed from any shear plane that may exist at depth. Non-coaxial strain is expected to dominate in regions proximal to zones of major thrusting.

It is difficult to determine the geometrical relationships of the megascopic $F_1$ folds due to the extensive metamorphism that has obscured many of the $S_1$ surfaces. Tertiary faulting has further complicated the original relationships. However, an interpretation that is internally consistent is shown in the cross sections on Plate III. These figures depict northeasterly plunging folds with northwesterly vergence disrupted by steep, younger normal faults.

Timing of the $F_1$ deformation is constrained by the Lower Jurassic (Toarcian) age of the youngest sediments involved (Nolf, 1966) and the age of post-tectonic plutons, the oldest of which in the vicinity of the area are dated at 160 ± 5 Ma old (Armstrong and others, 1976). Designation of this deformation as being first phase ($F_1$) is based on scrutiny of stratigraphic relationships within the thesis area which yielded no evidence of an earlier deformational event. However, the metamorphic and igneous clasts in the Lower Sedimentary Series indicate that a previously deformed area was exposed somewhere in the Seven Devils terrane.

$F_2$ Deformation

The second deformational phase is a cross-folding event that occurred shortly after $F_1$. Stratigraphic relationships within the thesis area indicate that both phases of deformation occurred between deposition of the youngest sediments and intrusion of the post-tectonic plutons. The narrow time span of approximately 30 million years during which both $F_1$ and $F_2$ structures formed suggests that the two phases may in fact be the discrete manifestation of a single but continuous deformational event. The time span is even more narrow when one considers that the youngest rocks in the Seven Devils terrane which were deformed by $F_1$ are of Early Oxfordian age (Morrison, 1964).

$F_2$ structures indicate that they are the result of north-northeasterly shortening. Fold axes plunge gently to steeply to the west-northwest. No metamorphism is associated with $F_2$, although an incipient differentiation associated with crenulated $S_1$ surfaces was
observed in several samples.

Cross folds have been described in many areas and are thought to represent the last phase of deformation in multiply deformed areas. Folds indicating shortening parallel to the dominant structural grain have been described in the Sierra Nevada, Caledonides of Ireland, and Precambrian of India, Canada, and South Africa (in Tobisch and Fiske, 1976). This widespread phenomenon suggests that it may represent a fundamental process in orogenesis.

Several models have been proposed to explain the development of cross folds. Tobisch and Fiske (1976) suggest that upon cessation of normal compression, "elastic recovery" becomes active. This induces shortening parallel to the structural grain and the formation of crenulation cleavage and cross folds. Avé Lallemant and others (1980) call upon local convergent flow in a low-angle thrust zone or in a subduction complex to explain the development of cross folds. Another possibility is that cross folding is the result of oblique convergence. The shortening axes of folds developed in these areas are nearly perpendicular to the orientation of the convergent plate boundary. This direction represents the normal component of convergence. The strike-slip component of the oblique convergence is expressed as strike-slip faulting and associated folds in or behind the arc (Oldow and others, 1984). Thus, shortening axes in cross folds may represent the strike-slip component in a zone of oblique subduction (Avé Lallemant, per. comm.). It has been suggested that pre-Nevadan cross folds developed in the Blue Mountains region are related to left-lateral displacements on faults within the volcanic arc terranes (Oldow and others, 1984). Finally, a fourth model calls on rotation of first phase fold axes into the tectonic transport direction (Escher and Watterson, 1974) and subsequent refolding orthogonally to initial folds. There are no data to indicate that this model is applicable to the present situation. One aspect common to all of these models is that deformation is continuous. There is little time lapse between the formation of first phase folds and second phase folds.

$F_3$ Deformation

A third and final phase of non-penetrative deformation is superimposed on the two Mesozoic deformations. This $F_3$ event resulted in the formation of west-northwesterly and north-northeasterly trending high angle normal faults that cut Cretaceous intrusive rocks and in some instances also displace Tertiary basalt flows. As discussed previously, these
faults have brought into juxtaposition rocks of medium to high-grade metamorphism and non-metamorphic rocks.

Most of these Tertiary faults are not exposed. Their existence is proposed to explain the repetition and sudden disappearance of many of the geologic units that occur throughout the map area. Prominent straight drainages coincide with the interpreted trace of these faults thus providing geomorphologic support for the existence of faulting.

A north-northeasterly trending normal fault is interpreted to lie beneath the alluvium in East Eagle Creek (Plate I). This fault displaces Mesozoic rocks down to the east and has resulted in the repetition of these units across East Eagle Creek. A second fault is interpreted to trend approximately parallel to the East Eagle Creek fault along the west flank of Krag Peak and Trux Mountain. This fault displaces rocks as young as the Cretaceous intrusives down to the east and is responsible for the abrupt "cut-out" of the Martin Bridge Limestone on the west flank of Krag Peak.

As previously mentioned, a fault is interpreted to lie in the Gold King Creek alluvial valley where it juxtaposes non-metamorphic rocks on the south with metamorphic ones to the north. This fault is thought to be a high angle normal fault that has displaced previously folded and thrust metamorphic rocks down to the north. The earlier deformation is interpreted to have resulted in a stack of thrust imbrications that brought high-grade metamorphic rocks into structural superposition with lower grade rocks. The younger normal faulting described in Gold King Creek resulted in the present distribution of these rocks.

**STRUCTURAL RELATIONS**

The following is an attempt to relate the structural geology of the thesis area to that of the rest of the Seven Devils terrane and other Blue Mountains terranes. In doing so, the deformational history of this region is placed in a plate-tectonic context and is related to the tectonic development of other portions of the Cordillera.

**SEVEN DEVILS TERRANE**

The results of this study indicate that two penetrative phases of deformation are re-
corded in the southwestern portion of the Seven Devils terrane. Structural and age relationships suggest that the two phases represent discrete points along a continuum of a single deformational event. The shortening axis associated with these structures trends northwesterly, while a second phase cross folding has a shortening axis trending north-northeasterly. Strain analysis indicates that tectonic transport directions are to the northwest. Timing of this deformation is constrained to have occurred between the upper Lower Jurassic (Toarcian) to lower Late Jurassic (Oxfordian).

The deformation observed in the Seven Devils terrane is interpreted to represent the regional D2 (Nevadan) event which is recorded in all four Blue Mountains terranes (Avé Lallemant and others, 1980; Oldow and others, 1984). In the western portion of the Central melange and Mesozoic clastic terranes, D2 folds trend easterly and are south vergent. These folds trend northeasterly in the eastern portion of the Central melange terrane (Avé Lallemant and others, 1980).

The earliest phase of deformation in the region (D1) is a Middle Triassic-Early Jurassic event. Manifestations of this long-lived event are found throughout the Central melange and Mesozoic clastic terranes. There is no penetrative record of this early event in the Seven Devils terrane. However, Vallier (1977) described three mylonitic shear zones within the terrane that yielded 40Ar/39Ar ages between 235 and 215 Ma B.P. (Avé Lallemant and others, 1980; Balcer, 1980). Detailed study of the southernmost Oxbow shear zone indicates that it is a major left-lateral strike-slip zone (Avé Lallemant, 1983; Avé Lallemant and others, 1985). Lund and others (1985) describe a pre-225 Ma. metamorphic event in the Seven Devils terrane in the Salmon River Canyon, Idaho which may be related to the shear zones in the Snake River Canyon.

TECTONIC SETTING

Several models have been proposed for the structural development of the Blue Mountains terranes. All of the models recognize the Central melange terrane and Mesozoic clastic terranes as portions of a forearc associated with either the Huntington arc terrane, Seven Devils terrane, or both (Dickinson and Thayer, 1978; Dickinson, 1979; Mullen, 1985, Avé Lallemant and others, 1985). In addition, various opinions exist regarding the relationship of the two arcs with one another, to Wrangellia, and to the North American craton. These models were discussed briefly on pages 12-16. Two of them will be further
Figure 50. Schematic hypothetical Model 1 Late Triassic and Late Jurassic tectonic configurations, after Brooks and Vallier (1978) and Avé Lallemant (1983).
Model 1
Late Triassic Configuration
examined, and a third suggested.

In the first model, the Huntington and Seven Devils volcanic arcs are interpreted to be fragments of a single, continuous arc (Figure 50). Brooks and Vallier (1978) cite similarities in the volcanic lithologies of both terranes and convergence of outcrop patterns and structural trends as evidence of the single arc theory. Avé Lallemand (1983) noted that removal of approximately 200 km of left-lateral slip along the Oxbow shear zone would bring the two terranes into on-strike alignment. The Central melange terrane is thought to be related to easterly subduction beneath the continuous Seven Devils-Huntington volcanic arc. The Mesozoic clastic terrane represents the deposits of a forearc basin related to the Huntington arc terrane.

Paleomagnetic data that indicate both arc terranes were at low latitudes during the Late Triassic (Hillhouse and others, 1982) support the single arc hypothesis. However, problems with this model have been raised on the basis of stratigraphic and structural relationships. Silberling (1983) objected to the correlation of the Huntington arc terrane with the Seven Devils terrane because of the absence of the thick Martin Bridge Limestone. Furthermore, the continuation of volcanism into the Jurassic in the Huntington area cannot be explained. Avé Lallemand and others (1985) discounted this model on the supposition that if the Seven Devils terrane was displaced outboard of an originally continuous arc system, volcanism would have continued in the Seven Devils and ceased in the Huntington arc—the opposite of which is found.

A second model has been proposed in which the two arc terranes are separate, distinct arcs (Avé Lallemand and others, 1985) (Figure 51). In this model, the Central melange terrane is related to easterly subduction beneath the Huntington arc and westerly subduction beneath the Seven Devils arc. Westerly subduction is interpreted on the basis of 223 Ma blueschists near Mitchell lying northwest of melange containing Lower Jurassic radiolarian chert. The Oxbow and related shear zones are interpreted to be interarc transcurrent fault zones related to left-oblique plate convergence during the Late Triassic (Avé Lallemand and others, 1985).

The data do not exclude the possibility of a third model (Figure 52). As in the previous model, the Huntington and Seven Devils arcs are considered to be separate arcs, and the Central melange and Mesozoic clastic terranes are related to eastward subduction beneath the Huntington arc terrane. This model differs from model 2 in that eastward subduction beneath the Seven Devils arc is hypothesized. Given this geometry, the blueschists near Mitchell would be related to subduction beneath the Huntington arc rather
Figure 51. Schematic hypothetical Model 2 Late Triassic and Late Jurassic tectonic configurations, after Avé Lallémant and others (1985).
Model 2
Late Triassic Configuration

Model 2
Late Jurassic Configuration
Figure 52. Schematic hypothetical Model 3 Late Triassic and Late Jurassic tectonic configurations.
Model 3
Late Triassic Configuration

Model 3
Late Jurassic Configuration
than the Seven Devils arc in the previous model. While not impossible, given the age of
the blueschists which coincides with cessation of volcanism in the Seven Devils arc, it
seems unlikely that the blueschists are related to the Huntington arc.

CORDILLERAN RELATIONS

Structural trends in the Seven Devils and adjacent Blue Mountains terranes can be
used to relate these areas with other portions of the Cordillera. In doing so, inferences are
made regarding the affinity of these terranes to known suspect terranes and/or North
America. These relations combined with paleomagnetic data enable tectonic
reconstructions for the region to be made.

Most studies have examined the tectonic framework of western North America on
the basis of stratigraphic relations (Burchfiel and Davis, 1972, 1975; Davis and others,
1978; Monger and others, 1982). One recent paper (Oldow and others, 1984) has
assessed the relationship of Cordilleran terranes to one another using structural
information. They point out that structures in the Mesozoic arc terranes of the northern
Sierra Nevada and of northeastern Oregon are remarkably uniform in terms of timing,
style, and orientation if the Late Cretaceous to Cenozoic rotations are restored. It is on the
basis of these similarities that the Mesozoic rocks of the Blue Mountains region are
interpreted to have formed in an along-strike segment of the Sierran arc system (Oldow and
others, 1984). They proposed that the Huntington arc terrane is not an exotic terrane but
rather represents a fringing arc system to North America. The lack of D₁ structures
combined with the nature of the fauna in the Seven Devils terrane lend support to their
hypothesis that this volcanic arc is indeed exotic to North America.

The regional D₁ deformation is a prolonged event lasting from the Late Triassic
through Early Jurassic. It has been interpreted to be related to subduction processes
associated with the formation of the Central melange terrane, blueschists, and tectonic
interleaving of Upper Triassic flysch with the melange. It is also during this interval that
the three shear zones in the Seven Devils terrane formed (Avé Lallemand and others, 1985).

The second event (D₂) occurred over a very short time interval within the Oxfordian
of the Late Jurassic (Nevadan). This deformation has been interpreted to be the result of
accretion of the Blue Mountains terranes to the North American continent (Dickinson and
Thayer, 1978; Dickinson, 1979; Avé Lallemand and others, 1980). This is perhaps not
strictly correct. $D_2$ structures are most probably the result of a collisional orogeny. Depending on which model one chooses, the colliding components may vary. The single arc model requires $D_2$ structures to be the result of an arc-continent collision. On the other hand, both "double arc" models attribute $D_2$ structures to arc-arc collisions. The resultant amalgamated arc assemblage would still be some distance outboard of the North American craton, presumably at the leading edge of attached oceanic or transitional lithosphere.

Paleomagnetic data support the arc-arc collision model. These data indicate that the Blue Mountains region has undergone approximately 65° since the Cretaceous (Magill and Cox, 1981; Simpson and Cox, 1977; Beck and others, 1978; Wilson and Cox, 1980; Schultz and Levi, 1981; Hillhouse and others, 1982). The first 30-35 degrees of rotation may be related to the closing of the marginal basin between the amalgamated arcs and the continental margin (Oldow and others, 1984). No penetrative deformational features associated with the impingement of the arcs to the continent are found in the Blue Mountains region (Avé Lallemand and others, 1985). However, low amplitude folds, flexures and faults may be the result of this Cenozoic rotation (Wilson and Cox, 1980; Hillhouse and others, 1982). The collision of the amalgamated Blue Mountains terranes with cratonic North America may have caused the intense late Cretaceous structures in the eastern part of the Seven Devils terrane and the Riggins Group in western Idaho (Sutter and others, 1984). The remaining rotation of about 30 degrees has been attributed to Basin and Range extension.

The single arc hypothesis must call on an arc-continent collision to explain the development of the penetrative Late Jurassic structures. In view of the paleomagnetic data which suggest that the arc did not dock with the continent until the Cretaceous or Cenozoic, this model seems inadequate.

**ARC-ARC COLLISION**

The Late Jurassic collision between the Seven Devils and Huntington island arc terranes resulted in the formation of oppositely verging structures in the respective arcs. When the effects of Cenozoic rotation are removed, the structures from both terranes are approximately parallel to the western coast of North America. However, folds and thrusts in the Huntington arc terrane verge towards North America, whereas those in the Seven
Figure 53. Arc-Arc collision between Sangihe and Halmahera island arcs in the Molucca Sea, from Hamilton (1979). A possible analogue to Model 2.
Devils terrane verge in the opposite direction. A somewhat analagous situation is presently occurring in the area between Sulawesi, Halmahera, and the Philippines (Hamilton, 1979). Here the eastward facing Sangihe island arc and western facing Halmahera island arc are converging (Figure 53). They are separated by the Molucca Sea melange wedge. Westward directed thrusting occurs in the Sangihe arc, while eastward directed thrusting occurs in the Halmahera arc.

SEVEN DEVILS VS. WRANGELLIA; HUNTINGTON ARC VS. NORTH AMERICA

The unique structural history of the Seven Devils terrane with respect to the other Blue Mountains terranes as well as the argument against a single arc-continent collision are compelling reasons for considering this terrane to be unrelated to the Huntington arc and therefore, North America. This combined with the paleontologic evidence of Newton (1983) are the basis for correlating the Seven Devils terrane with Wrangellia. Geochemical arguments by Sarewitz (1983) suggest that the two terranes represent a volcanic arc/backarc rift system.

Paleomagnetic data indicate that Wrangellia displays a complex polar wandering path from the Late Paleozoic through the Mesozoic (Stone and others, 1982). Two different travel paths have been suggested (Figure 54). In one model, Wrangellia has drifted since the Permian from northern to southern latitudes until the Late Jurassic when it began to drift north again. In the second model Wrangellia remains in the northern hemisphere throughout its travels. From Permian through Late Triassic time it is interpreted to have moved from intermediate to low northerly latitudes. It then moved northward from the Late Triassic until the Late Jurassic. Subsequently it moved southward until the Early Cretaceous when it moved northward again.

Avé Lallemant and others (1985) note that the Late Triassic left-lateral shear zones in the Seven Devils terrane are consistent with the southward translation of Wrangellia which was proposed in both models. In addition, both models indicate that Wrangellia/Seven Devils was moving northerly during the Late Jurassic, at which time it amalgamated with the Huntington-Mesozoic clastic-Central melange terranes. The second model is favored, however, because structures in northwestern Nevada that formed between 160 and 100 m.y. ago are suggestive of left-oblique convergence (Oldow and others, 1984). This convergence is consistent with the coeval southward drift interpreted for Wrangellia (Avé Lallemant and others, 1985).
Figure 54. Paleolatitude versus time for Cordilleran suspect terranes based on paleomagnetic data. Curve A interprets southern hemisphere drift, Curve B interprets only northern hemisphere drift. From Stone and others (1982).
Paleolatitude versus Time Curves
After (Stone and others, 1982)
Implicit in the above discussion is the assumption that the Huntington arc terrane remained essentially at the same latitude as the craton while Wrangellia was wandering. This is based on the assertion that the Huntington arc, Mesozoic clastic, and Central melange terranes are "...elements of a common arc system that formed the western margin of North America in the Mesozoic." (Oldow and others, 1984).

The correlation between Wrangellia and the Seven Devils terrane notwithstanding, it still is uncertain as to how much "traveling" the island arc itself did. If one accepts the model favored by Avé Lallemant and others (1985) in which the Seven Devils is the result of westward, left-oblique subduction, it may be that only the portion of Wrangellia behind the left-lateral, strike-slip shear zones travelled south as indicated by paleomagnetic data. On the other hand, if the model which relates the Seven Devils arc to east dipping subduction is employed, the arc itself may have translated to the south. If this possibility occurred, then Nevadan deformation may be the result of right-oblique convergence between the Seven Devils arc which is drifting northward relative to the Huntington arc terrane. However, no structures to substantiate this interpretation have been reported. Perhaps these discrepancies could be resolved by collecting paleomagnetic data from the Mesozoic section exposed in the Seven Devils terrane and comparing them with the data of Stone and others (1982).
SUMMARY AND CONCLUSIONS

The East Eagle Creek portion of the Seven Devils terrane in northeastern Oregon has undergone two phases of penetrative deformation during the Late Jurassic. Field relations indicate that the first phase resulted in northeasterly trending structures with northwesterly oriented shortening axes. Mesoscopic and microscopic analyses indicate that tectonic transport during this event was to the northwest. This deformation is associated with a regional metamorphic event that metamorphosed the volcanic and sedimentary rocks to the greenschist and amphibolite facies.

The second deformation occurred shortly after the first and is a cross-folding event. Both occurred subsequent to the deposition of the Lower to Upper? Jurassic Hurwal Formation and prior to the intrusion of Upper Jurassic-Cretaceous plutons. Because of the geometric and chronologic relationship between the two phases, they are considered to be discrete manifestations of a deformational continuum related to the regional, D₂ (Nevadan) deformation. No evidence was found of the regional Late Triassic-Early Jurassic D₁ deformation which is observed elsewhere in the Blue Mountains Province.

A third phase of non-penetrative deformation resulted in high angle normal faults. Field evidence suggests that most of these structures developed subsequent to the intrusion of Cretaceous plutons and prior to the extrusion of Tertiary basalts. Several normal faults have also displaced the Tertiary basalts.

The absence of D₁ structures in the Seven Devils terrane together with additional structural, stratigraphic, and paleontologic data provide the basis for considering the Seven Devils arc to be separate and distinct from the Huntington arc. In addition, these data indicate that the Seven Devils terrane is an exotic terrane and is correlatable with Wrangellia, and that the Huntington arc terrane is an original inhabitant of North America.

Mesozoic structures in the Seven Devils terrane are interpreted to be the result of an arc-arc collision between the quiescent Seven Devils arc terrane and active Huntington arc. The collision resulted in oppositely verging folds and thrust faults in each terrane. Details regarding the geometry of convergence between these terranes are still incomplete. Two models are permissible. Both situate the Huntington arc as a west facing arc on the leading edge of oceanic lithosphere attached to North America. The first model considers the Seven Devils arc to be an east facing arc above a left-oblique subduction zone (Avé
Lallemand and others, 1985). The Central melange terrane is thus related to both Seven Devils and Huntington arc subduction, and the Late Triassic blueschists near Mitchell are related to subduction beneath the Seven Devils arc. Model two considers the Seven Devils arc to be a west facing arc also situated above a left-oblique subduction zone. In this model, the Central melange terrane and the blueschists are unrelated to the Seven Devils terrane.

Paleomagnetic data that indicate Wrangellia had a complex drift history throughout the Mesozoic (Stone and others, 1982) have important implications on the travel path of the Seven Devils terrane. Depending on the model chosen, the movements of the volcanic arc varies. The first model permits Wrangellia to follow the polar wander curve without the company of the Seven Devils arc. The second model allows for the Seven Devils arc to have undergone the same complex translational history as Wrangellia. The correct differentiation between these two extremes will help to constrain the convergence configuration between the Huntington and Seven Devils arcs in the Jurassic.
References Cited

geochronometry of Mesozoic granitic rocks and their Sr isotopic
composition, Oregon, Washington, and Idaho: Geological Society of

late-Triassic strike-slip displacement in the Seven Devils terrane, Oregon
and Idaho: a result of left-oblique plate convergence?: Tectonophysics, v. 119, p.
299-328.

Avé Lallemant, H. G., 1976, Structure of the Canyon Mountain (Oregon)
ophiolite complex and its implication for sea floor spreading: Geological

Avé Lallemant, H. G., Phelps, D. W., and Sutter, J. F., 1980, \(^{40}\text{Ar}^{39}\text{Ar}\) ages of
some pre-Tertiary plutonic and metamorphic rocks of eastern Oregon and
their geologic relationships: Geology, v. 8, p. 371-374.

region, NE Oregon: Geological Society of America Abstracts with
Programs, v. 15, p. 372.

comparison of oceanic ridges, islands, and the Columbia plateau

Balcer, D. E., 1980, \(^{40}\text{Ar}^{39}\text{Ar}\) ages and REE geochemistry of selected
basement terranes, Snake River Canyon, Oregon-Idaho: Ohio
State University, MS Thesis, 111p.

W., Paleomagnetism of the middle Tertiary Clarno Formation,
north-central Oregon: Constraint on models for tectonic rotation: EoS,
v. 59, p. 1058.


Brooks, H. C., 1979, Plate tectonic and geologic history of the Blue Mountains:
Oregon Geology, v. 41, p. 71-80.


Gerlach, D. C., Leeman, W. P., and Avé Lallemant, H. G., 1981, Petrology and
geochemistry of plagiogranite in the Canyon Mountain ophiolite, Oregon: Contributions to Mineralogy and Petrology, v. 77, p. 82-92.


accretion and the origin of two major metamorphic and plutonic welts in the Canadian Cordillera: Geology, v. 10, p. 70-75.


Smith, W. D., and Allen, J. E., 1941, Geology and physiography of the northern Wallowa Mountains, Oregon: Oregon Department of Geology and Mineral Industries, Bull. 12, 64 p.


Sutter, J. F., Snee, L. W., and Lund, K., 1984, Metamorphic, plutonic, and


Wilson, D. and Cox, A., 1979, Paleomagnetic evidence for the rotation of Jurassic plutons in eastern Oregon (abs.): EOS (American Geophysical Union Transactions) v. 60, p. 240.


C MAP OF THE EAST EAGLE CREEK AREA
SOUTHERN WALLOWA MOUNTAINS:
NORTHEASTERN OREGON

BY
ANDREW S. MIRKIN
1986
CREEK AREA
ROCK UNITS

- QUATERNARY SURFICIAL DE
- TERTIARY COLUMBIA RIVER E
- CRETACEOUS INTRUSIVES
- TRIASSIC JURASSIC HURWAL
MAGNETIC NORTH DEC 1954

SCALE 1: 31,680

MILES
BASE MAP FROM U.S.G.S. EAGLE CAP QUADRANGLE, OREGON
15 MINUTE SERIES (1954)

GEO
(DAS)

NOR
(DAS)

BED
OVE
S₁ F
S₂ F

OFFICIAL DEPOSITS

A RIVER BASALT

PUSIVES

NC HURWAL FORMATION
GEOLOGIC CONTACT
(DASHED WHERE APPROXIMATELY LOCATED)

NORMAL FAULT
(DASHED WHERE APPROXIMATELY LOCATED)

BEDDING STRIKE AND DIP

OVERTURNED BEDDING STRIKE AND DIP

S₁ FOLIATION STRIKE AND DIP

S₂ FOLIATION STRIKE AND DIP
GEOLOGIC CONTACT
(DASHED WHERE APPROXIMATELY LOCATED)

NORMAL FAULT
(DASHED WHERE APPROXIMATELY LOCATED)

BEDDING STRIKE AND DIP

OVER TurnED BEDDING STRIKE AND DIP

S₁ FOLIATION STRIKE AND DIP

S₂ FOLIATION STRIKE AND DIP
SIDE MAP OF THE EAST EAGLE CREEK AREA
SOUTHERN WALLOWA MOUNTAINS:
NORTHEASTERN OREGON

UNMAPPED AR
QUATERNARY SURFICIAL DEPOSITS

TERTIARY EXTRUSIVES

CRETACEOUS INTRUSIVES

NON-METAMORPHIC /
LOW GRADE METAMORPHIC ROCKS

GREENSCHISTS

MARBLE

BIOTITE SCHISTS

AMPHIBOLITES &
POSSIBLE GRANULITES
TECTONIC BOUNDARY
(DASHED WHERE APPROXIMATELY LOCATED)

STRATIGRAPHIC CONTACT
(DASHED WHERE APPROXIMATELY LOCATED)
GEOLOGIC CROSS SECTIONS THROUGH Ti

WEST

A

ELEVATION IN FEET (MSL)

7600
6800
6000
5200
4400
3800
3000
2200
1400
400

WEST

B

8000
THROUGH THE SOUTHERN WALLOWA MOUNTAINS, NE OREGON

EAST

A'

Qs

7800
6800
6000
5200
4400
3600
2800
2000
1200
400

EAS

B
LEGEND

Quaternary Surficial Deposits, includes alluvium

Tertiary Columbia River Basalt

Cretaceous intrusives—granodioritic plutonic rocks—Wallowa batholith and Cornucopia stock

Triassic Jurassic Hurwal Formation

Triassic Martin Bridge Formation

Triassic Lower Sedimentary Series
Illuvium, colluvium and till

Ionic rocks of the

NO VERTICAL EXAGGERATION