AN ABSTRACT OF THE THESIS OF

Phyllis A. Camilleri for the degree of Master of Science in Geology presented on December 15, 1988.

Title: Superposed Compressional and Extensional Strain in Lower Paleozoic Rocks in the Northwestern Grant Range, Nevada

Redacted for Privacy

Abstract approved: ______________________
Karen Lund

The Grant Range, in east-central Nevada, is a north-east trending range bounded on the west by a west-dipping normal fault system. Rocks within the range record a complex polyphase Mesozoic ductile compressional and Cenozoic brittle extensional deformational history. The northwestern Grant Range exposes deformed, regionally metamorphosed and unmetamorphosed, Cambrian to Mississippian carbonate and clastic strata, and minor Tertiary granitic and andesitic dikes. Cambrian and Ordovician rocks are ductilely strained and metamorphosed. Metamorphic grade decreases stratigraphically upwards, generally commensurate with the degree of ductile strain.

Two Mesozoic compressional events are recorded in the rocks of the northwestern Grant Range. The first event produced mesoscopic, east-vergent folds with spaced axial-planar cleavage. These folds were overprinted by small-scale, west-vergent thrust faults and folds of the second event. Regional metamorphism began during the first folding event, but outlasted deformation. Static metamorphism was followed by west-vergent deformation, which marked the end of

metamorphism. The compressional structures may have been part of an east-vergent anticline or the hanging wall of an east-vergent thrust fault.

Ductile Mesozoic compressional structures and fabrics are cut by an arched, imbricate stack of Cenozoic low-angle normal faults of a more brittle character. The low-angle normal faults omit stratigraphic section, and each successively structurally higher fault is generally younger than the one below it. Most faults appear to be west-directed. The age of the low-angle normal fault system is constrained only as late Oligocene to Pleistocene, but could be largely Miocene to Pleistocene. Some granitic and andesitic dikes cross-cut or are cut by low-angle normal faults, indicating that igneous activity is at least in part synchronous with extension.

The geometry of the low-angle normal faults suggest that these faults could be rotated, extinct fault segments formed as a result of arching of the upper reaches of the high- to moderate-angle west-dipping normal fault system responsible for the uplift of the Grant Range. This would suggest that the low-angle normal faults are products of progressive Basin and Range extension and do not represent a distinct extensional event.

SUPERPOSED COMPRESSIONAL AND EXTENSIONAL STRAIN IN LOWER PALEOZOIC ROCKS OF THE NORTHWESTERN GRANT RANGE, NEVADA

by

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SUPERPOSED COMPRESSIONAL AND EXTENSIONAL STRAIN IN LOWER PALEOZOIC ROCKS OF THE NORTHWESTERN GRANT RANGE, NEVADA

INTRODUCTION

This paper focuses on polyphase deformation in Paleozoic rocks in the northwestern part of the Grant Range, east-central Nevada (Fig. 1). The Grant Range lies to the west of the Cretaceous to early Tertiary (Heller et al., 1986; Wiltschko and Dorr, 1983) Sevier fold and thrust front and lies within a region where Mesozoic compressional structure has been severely obscured by younger, Tertiary extensional deformation. The Grant range exposes deformed Paleozoic carbonate and clastic sedimentary rocks and Tertiary carbonate, clastic, and volcanic rocks.

Rocks within the Grant Range record a complex polyphase compressional and extensional deformational history. In the southern Grant Range, regionally metamorphosed Cambrian and Ordovician rocks compose a Mesozoic map-scale, east-vergent overturned anticline with associated mesoscopic folds (Cebull, 1970; Fryxell, 1987). Regional metamorphism began during folding, but outlasted deformation (Fryxell, 1987). Subsequent to regional metamorphism, the anticline was intruded by a Mesozoic pluton; this pluton and Paleozoic country rocks are cut by Tertiary low-angle faults that omit stratigraphic section (Fryxell, 1987). Fryxell (1987) interpreted the low-angle faults to be normal faults that formed at low-angles. In the central Grant Range, Hyde and Huttrer (1970), working in similar rocks, also noted that low-angle faulting post-dated the metamorphism, however, they suggested that all low-angle faults were Mesozoic

Quaternary Alluvium

Quaternary basalt

Tertiary basalt

Oligocene and Miocene silicic to intermediate volcanic and volcaniclastic rocks

Miocene-Pliocene "basin-fill"

Cretaceous (?) granite (in the southern White Pine Range)

Cretaceous granite (in the southern Grant Range)

Mississippian to Permian carbonate and clastic rocks

Cambrian to Devonian carbonate and clastic rocks

low-angle normal fault

"range-front fault": arrow indicates dip direction

Figure 1. Simplified geologic map of east-central Nevada. Data from Kleinhampl and Ziony (1985), Stewart (1977), Moores et al. (1968), Fryxell (1984) and this study. Cross-section X-X′ is shown in Figure 2. The polygon outlined in bold in the northwestern Grant Range is the study area.

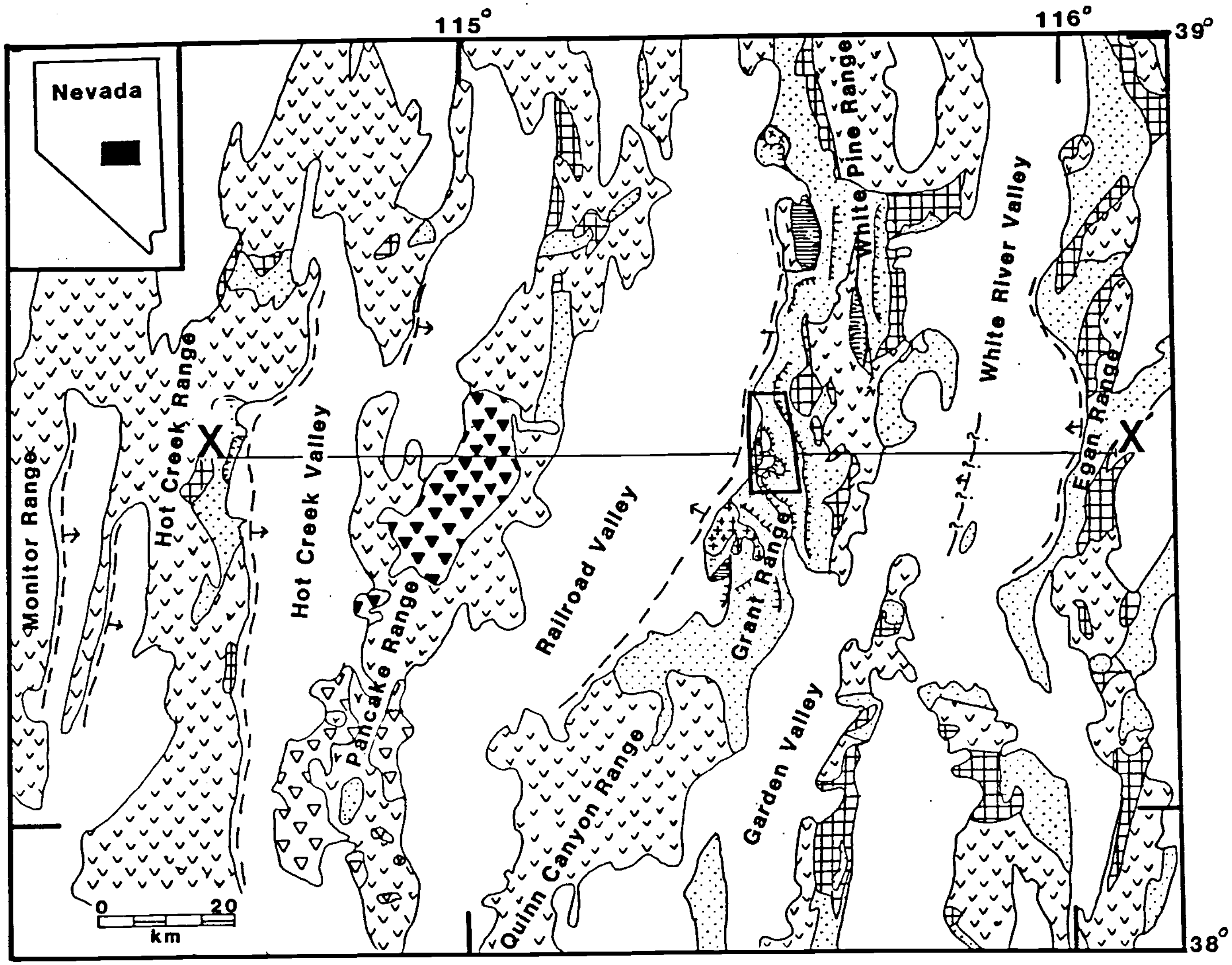


Figure 1 continued.

thrust faults and that "overturned to the east" mesoscopic folds in proximity of the low-angle faults formed at the same time as the faults and thus were an indication that the low-angle faults were east-vergent. In contrast, in the northernmost Grant Range and southern White Pine Range (Fig. 1) low-angle faults that omit stratigraphic section are interpreted to be mainly Tertiary surficial gravity slides which were superimposed on gentle, large wavelength Mesozoic folds (Moores et al., 1968). Thus there exists a variety of interpretations to explain similar structural relations in the Grant Range.

This study is a detailed part of a U. S. Geological Survey study of a proposed wilderness area in the northern Grant Range. The purpose of this study was to map in detail a small part of the northwestern Grant Range (Fig. 1) previously mapped by Hyde and Huttrer (1970) and to determine the nature of the metamorphism, its relation to folding and faulting, and the kinematics and geometry of folding and low-angle faulting. This area was brought to my attention by Karen Lund who, through the course of her USGS mapping project in the northern Grant Range, recognized that the low-angle faults in the study area had a geometry that was more compatible with an extensional origin and that some of the low-angle faults overprinted rocks that record a complex ductile deformational-metamorphic history (Lund et al., 1987).

Detailed mapping and structural evaluation of a part of the northwestern Grant Range suggests: (1) that the rocks record two Mesozoic compressional events; an earlier east-vergent deformation that is overprinted by west-vergent deformation, (2) the major low-angle faults are Tertiary (crudely bracketed as Oligocene to Pleistocene) normal faults, (3) most of the low-angle faults may be west-directed and could represent rotated, extinct faults related to the modern west-dipping normal fault system responsible for the uplift of the Grant Range.

REGIONAL GEOLOGIC SETTING

Paleozoic rocks exposed in east-central Nevada (Fig. 1) compose a thick (minimum 9 km; Kellogg, 1963) sequence of Cambrian to Permian carbonate and siliciclastic rocks that represent marine and nonmarine deposition along the Paleozoic continental margin of north America. Permian carbonate rocks represent the last vestige of marine depostion in east-central Nevada. Mesozoic sedimentary rocks are generally absent in east-central Nevada (Fig. 1), and Tertiary rocks were deposited on relatively flat-lying upper Paleozoic strata (Moores et al., 1968). Although there is no record of major surficial Mesozoic tectonic events, Paleozoic rocks in the southern Grant Range record compressional deformation, metamorphism, and plutonism at depth during this time (Fryxell, 1987).

Tertiary rocks unconformably overlie upper Paleozoic rocks and formed mainly during extensional deformation. The oldest Tertiary rocks compose the Paleocene-Eocene Sheep Pass Formation (Fouch, 1979), which consists of alluvial clastic and lacustrine carbonate strata. Kellogg (1964) suggests that the Sheep Pass Formation was deposited in a basin created by a normal fault and thus may represent the first record of extension in east-central Nevada. The Sheep Pass Formation is overlain by a widespread sequence of Oligocene to Early Miocene (Moores et al., 1968; Ekren et al., 1974) silicic to intermediate volcanic and volcaniclastic rocks. The source of some of these volcanic rocks is a caldera complex in the central Pancake Range (Ekren et al., 1974; Fig.1). In the southern Grant Range, Bartley et al. (1988) have recognized normal faults that are synchronous with this period of volcanism. However, little regional work has been done on the sequence of volcanic and volcaniclastic strata, thus the nature and extent of supracrustal extension during this time are poorly known. However, apparently no major normal fault-bounded basins (such as the modern ones; Fig. 2) developed because there are no such thick basin-fill deposits preserved within the volcanic and volcaniclastic sequence. The Sheep Pass Formation and the younger volcanic and volcaniclastic sequence are cut by faults that formed the modern ranges and basins (Fig. 2; the distribution of Sheep Pass Formation is too small to be shown in Figs. 1 and 2).

The volcanic and volcaniclastic sequence is overlain, in places, by fluvio-lacustrine Miocene-Pliocene to Quaternary basin-fill strata, which presumably are products of uplift and denudation of the modern ranges. The age of these strata provides a minimum estimate of time of inception of "Basin and Range extension". Basin-fill strata consist of breccia, conglomerate, sandstone, shale, and minor tuffaceous and volcaniclastic strata (Moores et al., 1968; Fryxell, 1984). Miocene-Pliocene basin-fill strata are exposed in the northernmost Grant Range-southern White Pine Range and in the southern Grant Range (the Horse Camp Formation: Moores et al., 1968; Fryxell, 1984; Fig. 1); these strata are tilted to the east and are in low- to moderate-angle normal-fault contact with lower Paleozoic rocks. In the modern basins (e.g. Railroad Valley), basin-fill strata are Miocene to Recent in age and are relatively flat lying (Effimoff and Pinezich, 1979; Fig. 2).

In places, generally well indurated, relatively flat-lying Pleistocene (?) pluvial deposits ("old alluvium or fanglomerate") mantle the east and west flanks of the southern White Pine and Grant Ranges (Kleinhampl and Ziony, 1985); these deposits are now

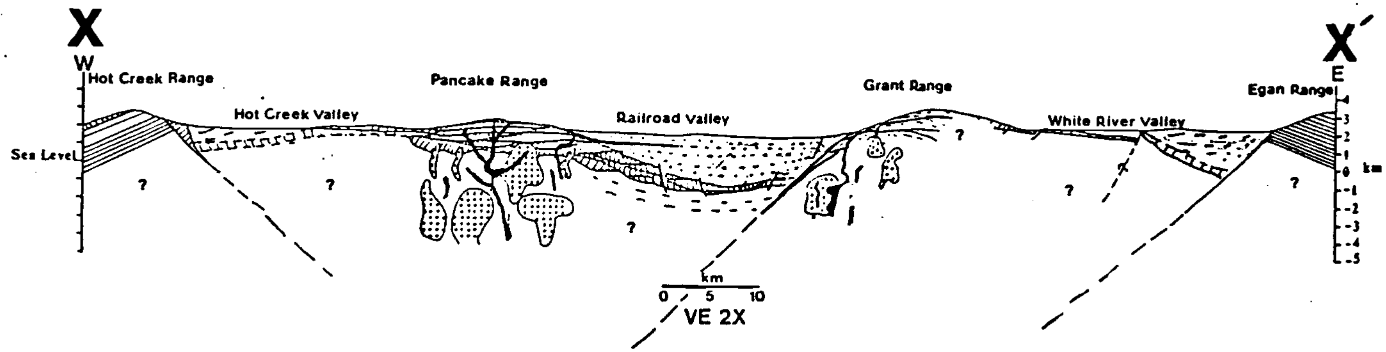


Figure 2. Interpretive east-west geologic cross-section through the Hot Creek, Pancake, Grant, and southern Egan Ranges. Vertically ruled pattern represents Oligocene to early Miocene silicic to intermediate volcanic rocks. Stippled hemispherical masses represent inferred silicic to intermediate intrusives and solid black bodies represents inferred basaltic intrusions. The lined pattern represents Paleozoic strata and shaded regions are Miocene to Recent basin-fill strata. The distribution of the Paleocene-Eocene Sheep Pass Formation is too small to show on this cross-section. Unpatterened areas that are queried represent areas where the structure of the Paleozoic section is not known. Curved lines in the Grant Range represent low-angle younger-over-older faults in the Paleozoic section. The geometry of Railroad Valley was based on a published seismic reflection profile (Effimoff and Pinezich, 1981). The geometry of Hot Creek and White River Valleys is inferred because there is little published data on these basins. Data used in construction of this cross-section is from this study, Kellogg (1963), Bortz and Murray (1979), Effimoff and Pinezich (1981), Ekren <u>et</u> <u>al</u>. (1974), Newman (1979), Scott (1965), and Kleinhampl and Ziony (1985).

deeply disected by the modern streams. The youngest rocks in the region are Quaternary basalt. Basalt occurs in the Pancake Range (Fig. 1) and in a small area in the southern Grant Range where it rests on old alluvium (Huttrer, 1966).

GEOLOGIC SETTING OF THE STUDY AREA

The study area exposes Cambrian to Mississippian carbonate and clastic strata (Fig. 3 [more detailed descriptions are given in Appendix B]) and minor granitic and andesitic dikes (Plate 1; these were not studied in detail). These rocks are cut by an arched, imbricate stack of low-angle faults that omit stratigraphic section (shown diagramatically in Fig. 2).

Cambrian and Ordovician rocks are metamorphosed to low-grades and ductilely strained. The highest grade rocks are greenschist facies and metamorphic grade decreases stratigraphically upwards (Fig. 4). Upper to Middle Cambrian pelitic rocks (Fig. 3) are biotite-, chlorite-, white mica-phyllite or schist (Fig. 4). Diagnostic metamorphic minerals in carbonate rocks with a pelitic component are amphibole and phlogopite in Middle Cambrian rocks and white mica and chlorite (rare) in Upper Cambrian to Middle Ordovician rocks (Figs. 3 and 4). Metamorphism in Upper Ordovician rocks is manifested as sericite phyllite or argillite (Fig. 4) or not manifested where rocks contain no pelitic component. Upper Ordovician rocks are fossiliferous and tend to retain a sedimentary character except where deformed. Silurian and Lower Devonian rocks are everywhere in low-angle fault contact with older rocks (Plate 1) and appear to be unmetamorphosed, however, these rocks are dolostone (Fig. 3) and therefore lack a mineralogy that would reflect metamorphism. Upper Devonian and Mississippian rocks are unmetamorphosed (Fig. 3). Metamorphism in the study area can be described as gradually decreasing stratigraphically upwards. However, because Silurian and Lower Devonian rocks lack a mineralogy that would reflect very low-grade metamorphism (Fig. 3) and they lie in fault

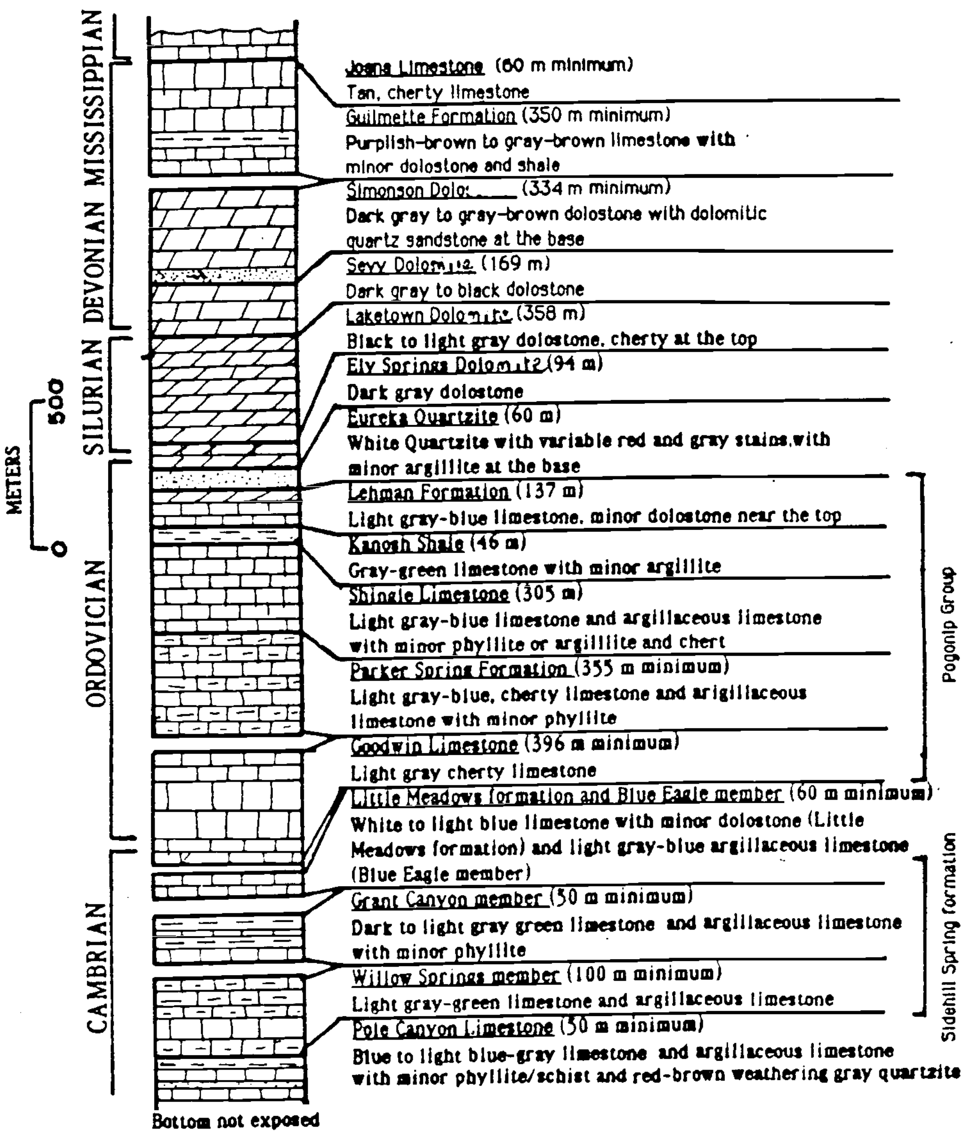


Figure 3. Stratigraphic column of the northwestern Grant Range. Breaks in the stratigraphic column represent faulted sections. Thicknesses of the Ely Springs, Sevy and Simonson Dolomites, Guilmette Formation and Joana Limestone were obtained by trigonometric methods off of an unpublished geologic map by Lund and Beard (1984-1987) from undeformed sections a few kilometers south of the study area. Thicknesses of stratigraphically lower units were obtained from Plate 1. "Minimum thicknesses" represent tectonic thicknesses.

		PELITE	PELITIC CARBONATE
ORDOVICIAN	Ely Springs Dolomite	N/A	N/A
	Eureka Quartzite	sericite, pyrite: phyllite or argillite	N/A
	Antelope Valley Limestone	N/A	N/A
	Kanosh Shale	sericite: argillite	sericite: limestone
	Shingle Limestone	sericite, pyrite: phyllite	sericite, pyrite: limestone
	Parker Spring Formation	" "	sericite, pyrite, chlorite (rare): limestone
	Goodwin Limestone	N/A	N/A
CAMBRIAN	Little Meadows formation	N/A	N/A
	Blue Eagle member	N/A	white mica, tourmaline: limestone/fine-grained marble
	Grant Canyon member	white mica, chlorite, ilmenite (?, altered to leucoxene): phyllite	no data
	Willow Spring member	N/A	phlogopite, amphibole (after phlogopite; pargasite?), chlorite, white mica: fine-grained marble
	Pole Canyon Limestone	biotite, white mica, chlorite: phyllite or schist	phlogopite, white mica, chlorite: fine-grained marble

Figure 4. Diagnostic metamorphic minerals and rock types in Cambrian and Ordovician rocks in the study area.

contact with older, demonstrably metamorphosed rocks, albeit low-grade, it is not clear whether they have undergone the same thermal history as the older rocks. This problem awaits further study.

COMPRESSIONAL STRUCTURES

INTRODUCTION

Compressional structures in the northwestern Grant Range comprise mesoscopic folds and small-scale thrust faults. Compressional structures abound in thin-bedded Cambrian and Ordovician limestone, argillaceous limestone, and pelite but are comparatively rare in younger rocks (dominantly thick-bedded dolostone). The distribution of compressional structures is illustrated in Plate 2. The Cambrian and Ordovician rocks can be divided into two domains which will be referred to as the northern and southern domain. The boundary between the two domains is the ridge that separates Beaty and Blair Canyons (Plate 1): reference to rocks in the southern or northern domains refers to rocks south or north of this ridge respectively.

EAST-VERGENT FOLDS

The oldest compressional structures, which are found in the northern and southern domains, are mesoscopic, open to isoclinal, east-vergent folds (F_1; Fig. 5) with spaced axial-planar cleavage (S_1; Plates 1 and 2). Rare parasitic folds are associated with some of these folds at lower stratigraphic levels.

F_1 fold axes and intersection lineations (S_1 X S_0) plunge gently in north to north-northwest and south to south-southeast directions (Fig. 6). In rocks of the southern domain, axial surfaces and S_1, where not overprinted, dip west (Plates 1 and 2). Axial surfaces and S_1 in rocks of the northern domain, however, dip west on the west side of the map area and dip east on the east side (Plate 2). The poles to S_1 and axial

W E

Figure 5. East-vergent F_1 fold in the Willow Springs member. Pencil for scale.

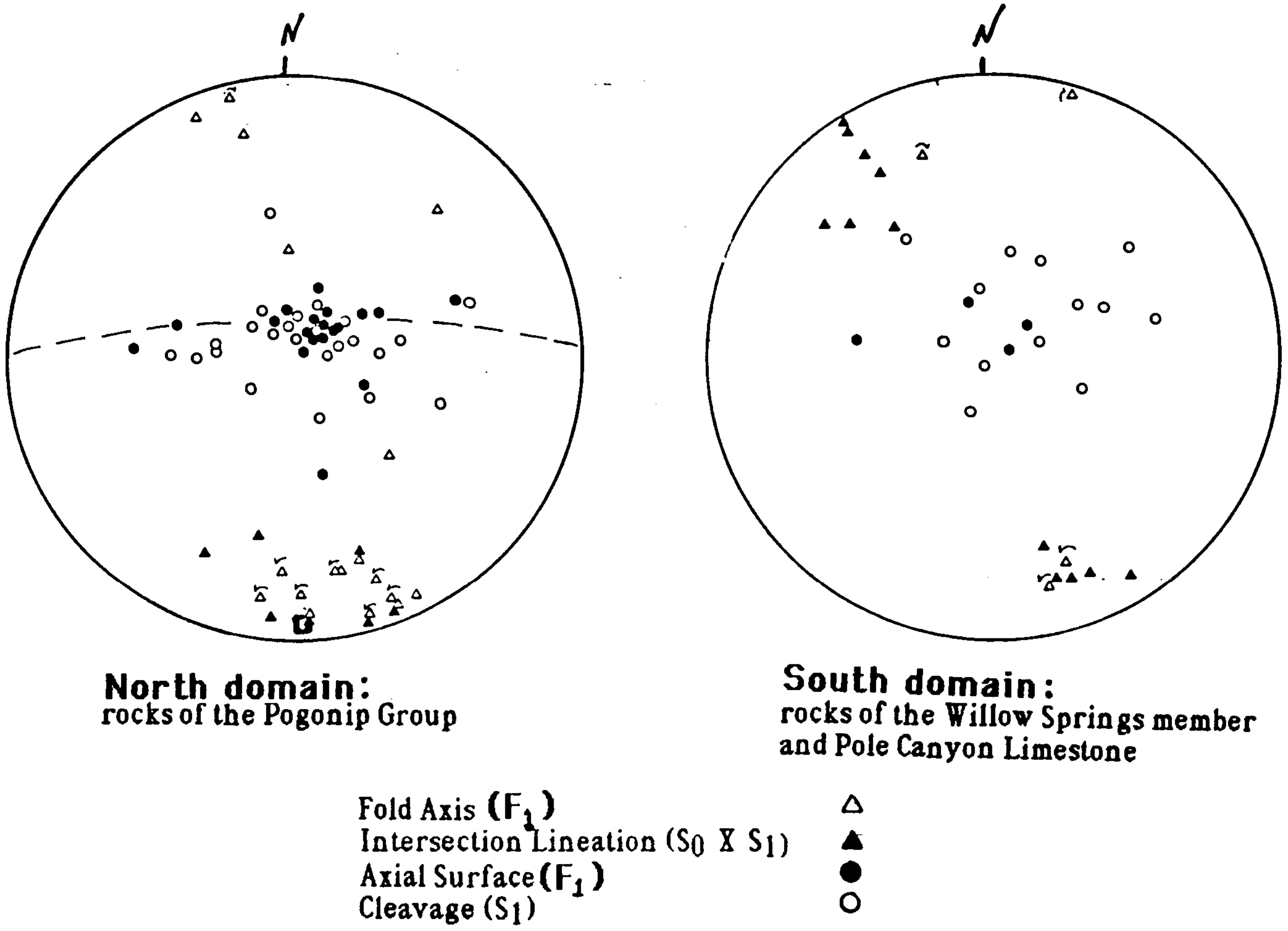


Figure 6. Lower hemisphere projection of F_1 fold axes, intersection lineations, axial surfaces and poles to S_1. Large open square is the pole to the great circle formed by the poles to S_1 and axial surfaces in the Pogonip Group in the northern domain.

surfaces in the northern domain crudely fall on a great circle: the pole to this great circle suggests subsequent refolding of the rocks of the northern domain about a north-northwest trending axis (Fig. 6).

Although local F_1 fold morphology varies with lithology and bedding character within and between stratigraphic units, the degree of ductile strain recorded in the rocks generally increases with stratigraphic and structural depth. At higher stratigraphic levels, fold geometry is roughly concentric. At lower stratigraphic levels, particularly within thin-bedded argillaceous carbonate, folds exhibit significant attenuation or boudinage on the limbs and thickening in the hinge regions; in certain parts of the deepest stratigraphic levels, in thin-bedded argillaceous carbonate, the presence of rootless folds gives evidence that bedding is completely transposed. At all stratigraphic levels, however, cleavage tends to be strong in the hinge regions of folds and bedding is locally transposed.

WEST-VERGENT STRUCTURES

In the northern and southern domains, east-vergent folds are overprinted by west-vergent, small-scale thrust faults and open to tight folds (F_{2x}; Plates 1 and 2), which are approximately coaxial with F_1 folds (Fig. 7). Carbonate layers folded by F_{2x} folds can contain a rare, weak, mesoscopic axial-planar cleavage. Phyllitic layers folded by F_{2x} folds contain a strong crenulation lineation formed by the intersection of S_1 and S_2; this lineation is coaxial with F_{2x} fold axes (Fig. 8). F_{2x} folds occur most commonly in the upper plates of minor west-vergent thrust faults (Plate 1 and 2). In the upper plate of these thrust faults, S_1 is invariably

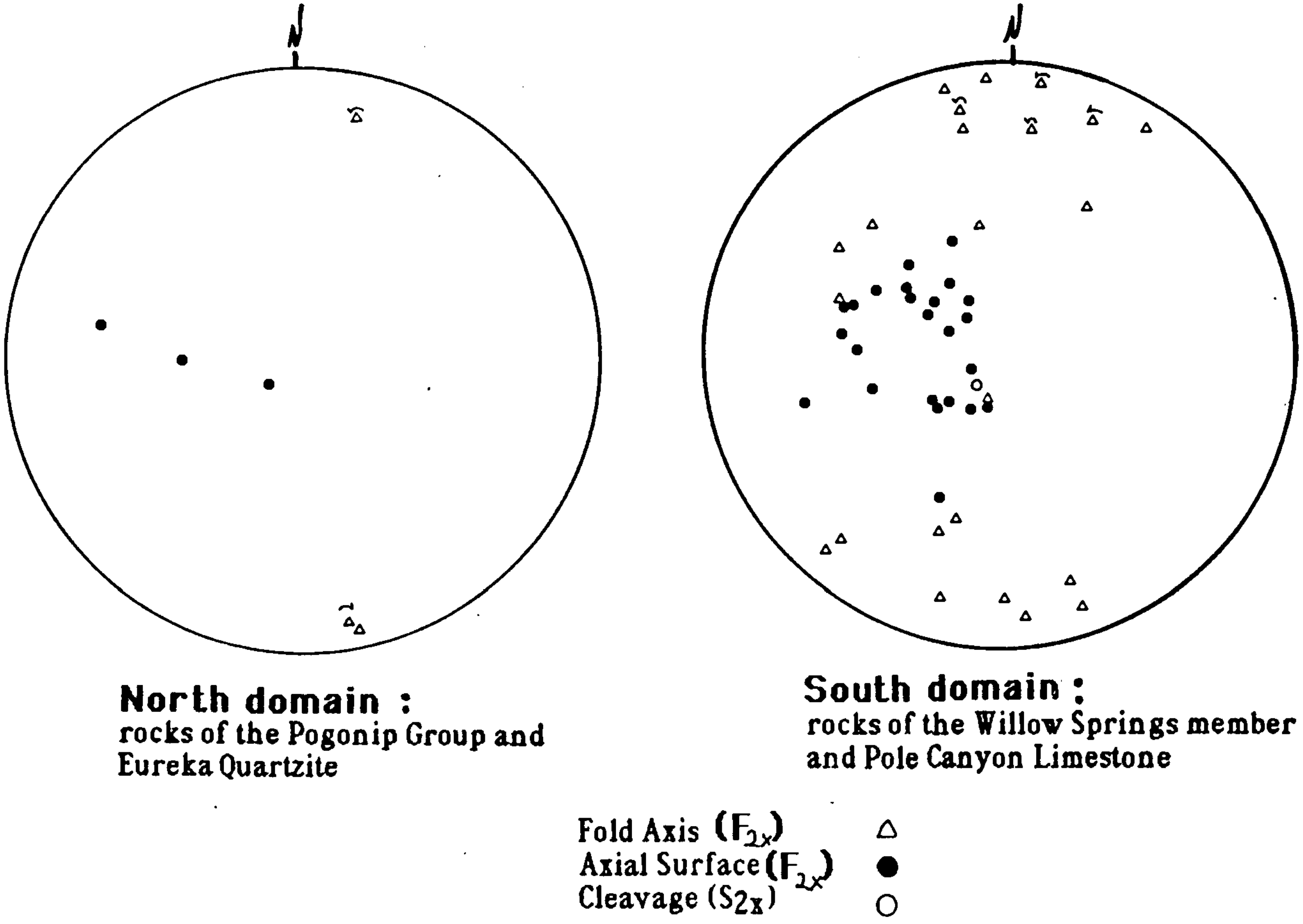
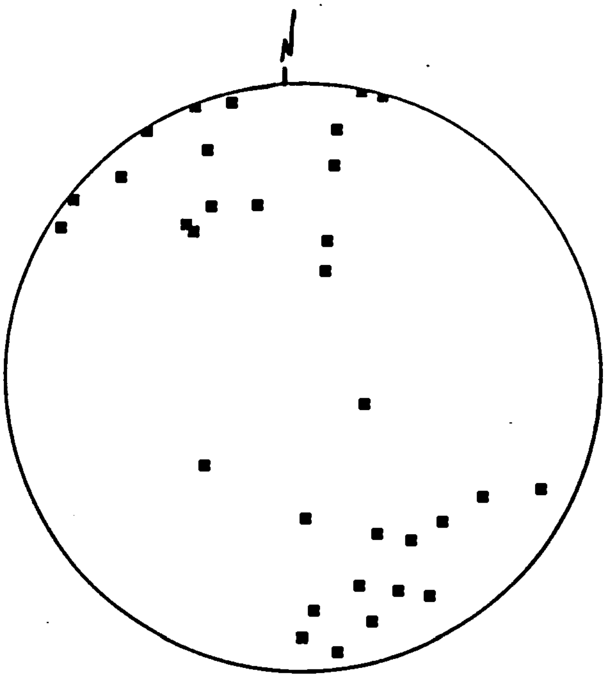


Figure 7. Lower hemisphere projection of fold axes and poles to S_{2x} and axial surfaces of F_{2x} folds.



South domain:
rocks of the Pole Canyon Limestone

Intersection Lineation (S_1 X S_2) ■

Figure 8. Lower hemisphere projection of intersection lineations (S_1 X S_2).

deformed (Fig. 9) and in a few places, F_2 folds can be seen to refold F_1 folds.

In the northern domain, small-scale, west-vergent fault-bend kink folds occur in the Eureka Quartzite. West-vergent thrust faults in the Pogonip Group (stratigraphically below the Eureka Quartzite) in this same area, however, are more complex. These faults cut F_1 folds, emplace younger rocks over older rocks, and where displacement across the faults dies out (the faults become "blind") F_1 folds are broadly refolded (shown diagrammatically in Plate 2, cross-sections A-A'and B-B'). No fault-bend folds, such as those in the Eureka Quartzite, occur in the Pogonip Group and no thrust faults like those in the Pogonip Group occur in the Eureka Quartzite. The reasons for the differing thrust fault geometries in these units may be related to the pre-thrust faulting geometry; the Pogonip Group is tightly folded (F_1 folds) near the Shingle Limestone-Parker Spring Formation contact and the amplitude of these folds dies out stratigraphically upwards. It is possible that during west-vergent thrust faulting, anistropy caused by the strong S_1 cleavage may have been oriented in such a way that strain was accommodated along S_1 in the Pogonip Group.

In the southern domain, the Willow Springs member and Pole Canyon Limestone are cut by a west-vergent thrust fault (Plate 2). This fault discordantly cuts west-dipping rocks that contain F_1 folds (Fig. 10; this relationship is also shown diagrammatically in Plate 2, cross-section E-E') and emplaces older rocks on younger rocks. F_{2x} folds occur in the upper plate of this fault as well as other areas which may be related to blind thrusts.

Figure 9. S_1 (conspicuous layering) in the Pole Canyon Limestone is folded by small F_{2x} folds in the upper plate of the west-vergent thrust fault on the north side of Heath Canyon (Plate 1). Note folded boudins (arrows show locations). Pencil for scale.

Figure 10. Photograph and sketch of a west-vergent thrust fault on the north side of Heath Canyon (see Plate 1 for location). "Cpc" is the Pole Canyon Limestone (stippled pattern); "Csw" is the Willow Springs member (cross pattern); and "Coblg" is the Blue Eagle member, Little Meadows formation and the Goodwin Limestone undifferentiated (hatchured pattern). The thrust fault juxtaposes the Pole Canyon Limestone over the Willow Springs member, and cuts earlier east-vergent folds (shown diagrammatically in the sketch). Note the west-vergent fold in the upper plate of the thrust fault. The thrust fault is cut by a low-angle normal fault. The low-angle normal fault juxtaposes "Coblg" on top of "Cpc" and "Csw".

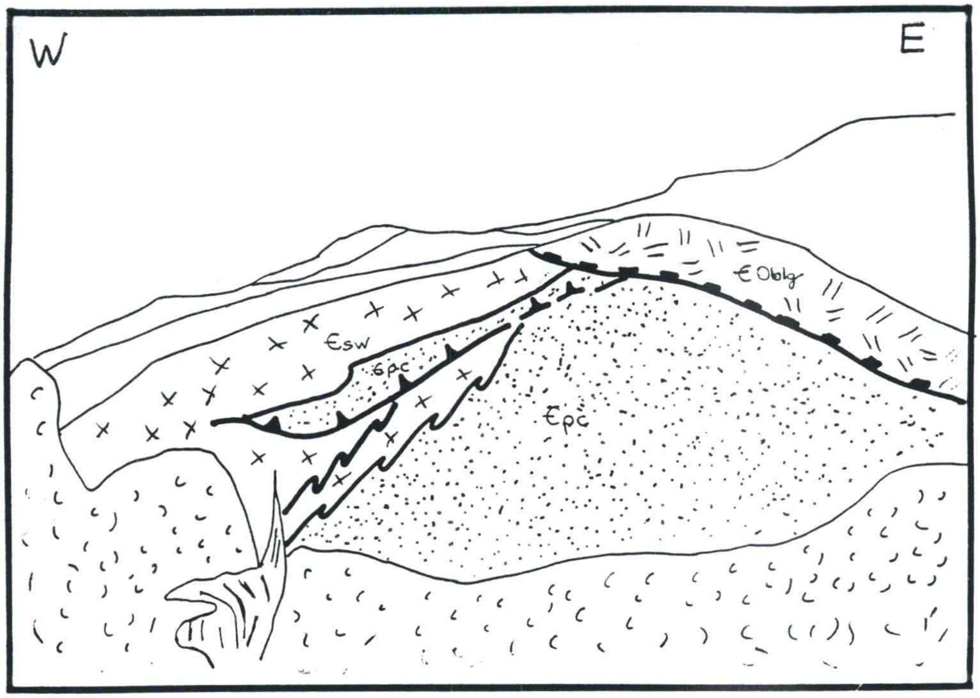


Figure 10 continued.

UPRIGHT FOLDS

In the northern domain, on the northwestern side of Beaty Canyon, small-scale upright symmetrical folds (F_{2y}) refold F_1 folds (Fig. 11), but their relationship to F_{2x} folds is not known. F_{2y} folds are small in amplitude and wavelength and have poorly developed axial-planar cleavage. The folds are approximately coaxial with F_1 and F_{2x} folds. Axial surfaces dip moderately to steeply west to southwest (Fig. 11).

METAMORPHISM

Regional metamorphism began during the formation of F_1 folds as evidenced by the growth of white mica synchronous with S_1 (Fig. 12). Metamorphism continued after the cessation of east-vergent deformation as indicated by randomly oriented mica and amphibole poikiloblasts. This is especially evident in the Middle Cambrian Willow Springs member wherein phlogopite is randomly oriented within the S_1 cleavage and in places crosscuts the cleavage boundaries (Fig. 13). Metamorphic minerals with static textures are most abundant in stratigraphically lower rocks and are lacking in stratigraphically higher rocks. Conversely, synkinematic textures associated with F_1 folds are best preserved in stratigraphically higher rocks and are less apparent in stratigraphically lower rocks where the effects of static metamorphism are most apparent.

West-vergent deformation followed static metamorphism. White mica and chlorite (rare) are the only minerals oriented in S_2. Metamorphic phlogopite or biotite in rocks affected by west-vergent deformation are kinked, sheared, and in places exhibit polygonization or less commonly recrystallization (Fig.

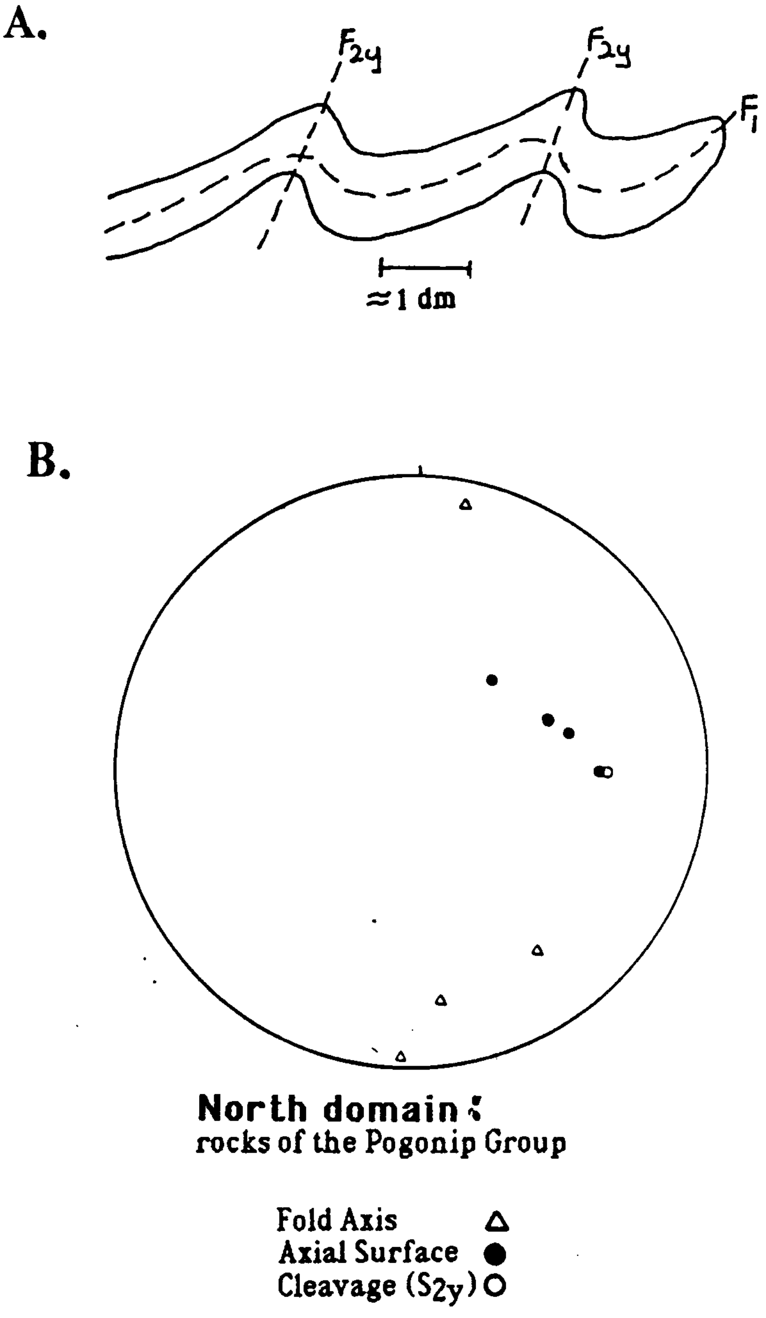


Figure 11. A. Sketch of F_{2y} folds. B. Lower hemisphere projection of fold axes and poles to S_{2y} and axial surfaces of F_{2y} folds.

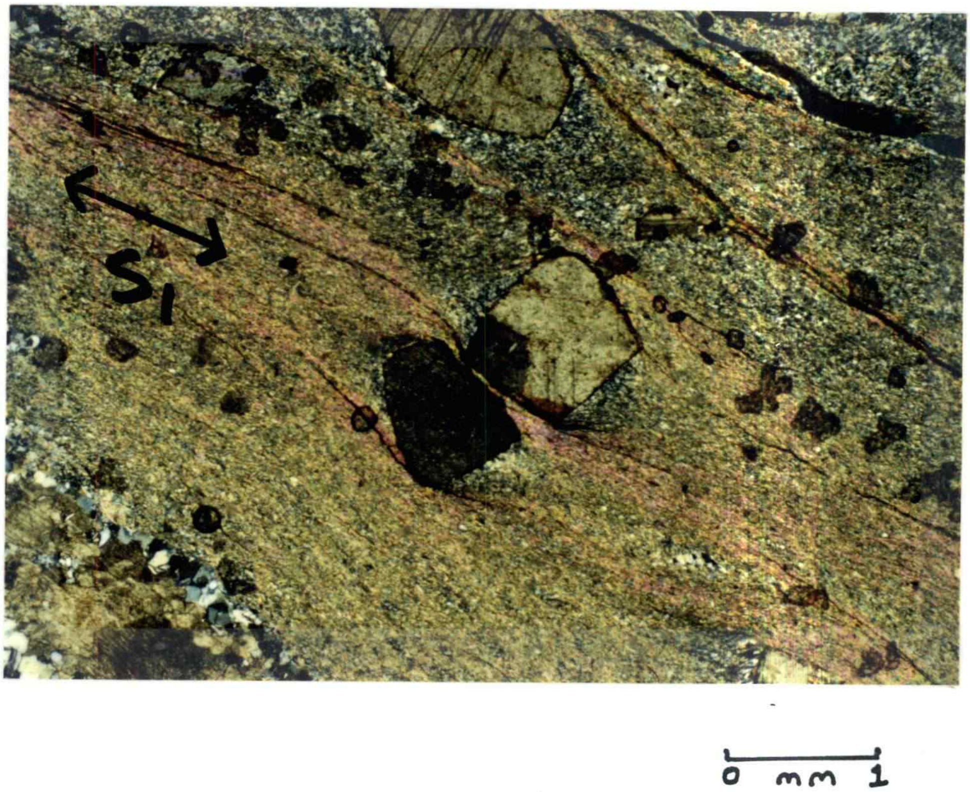
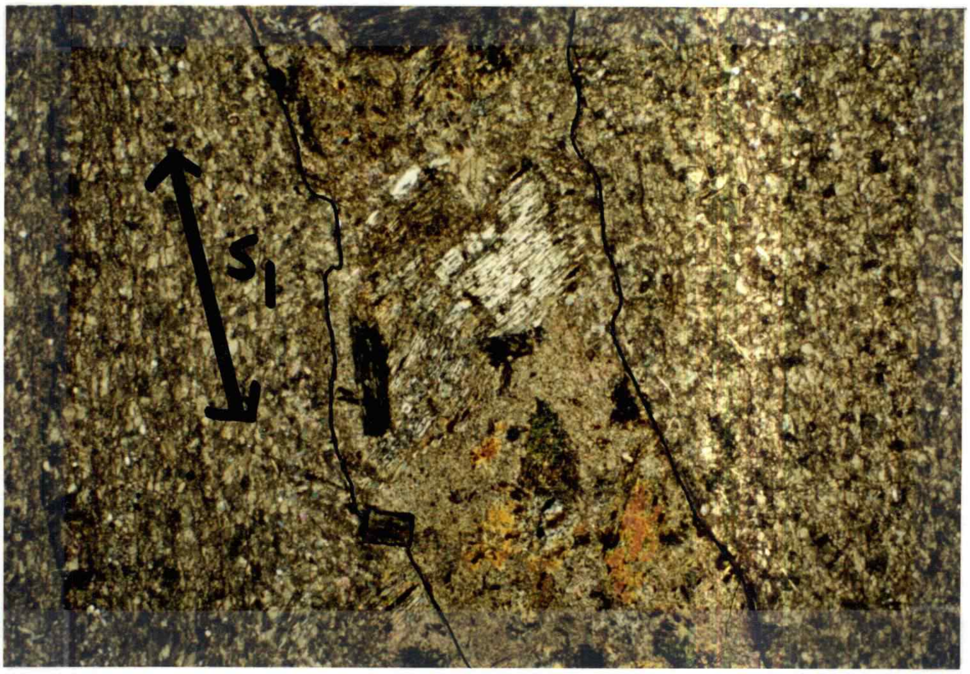


Figure 12. Photomicrograph of S_1 in the Parker Spring Formation in Beaty Canyon. Mineralogy is calcite, quartz, white mica, and dolomite. Mineral with pressure shadows is dolomite. White mica forms the foliation. Section cut perpendicular to intersection lineation. Crossed polars.

Figure 13. A. S_1 cleavage and bedding in the Willow Springs member. Figure 12 C shows an enlarged line drawing of this photo. Pencil for scale. Cleavage weathers out into relief and is composed of coarse-grained micas. From the upper plate of the thrust fault on the north side of Heath Canyon. B. Photomicrograph of the S_1 cleavage in photo "A". Figure 12 D shows a line drawing of this photomicrograph. Mineralogy is Phlogopite and calcite. This photomicrograph shows a fine penetrative cleavage that is parallel to the large cleavage in the center. In the large cleavage, micas tend to be randomly oriented and in places crosscut the cleavage boundary. Section cut perpendicular to intersection lineation
Crossed polars.

a

b



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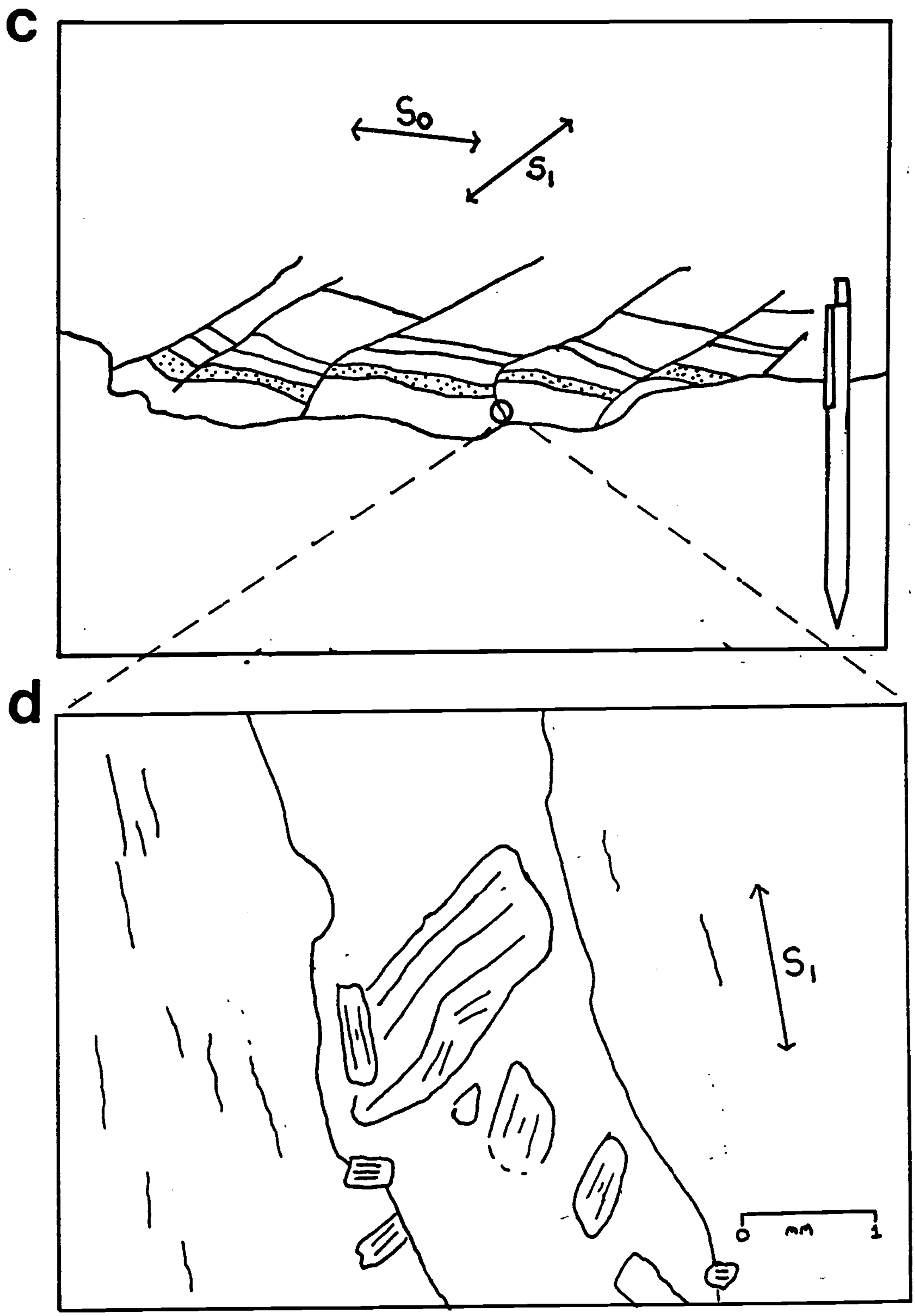


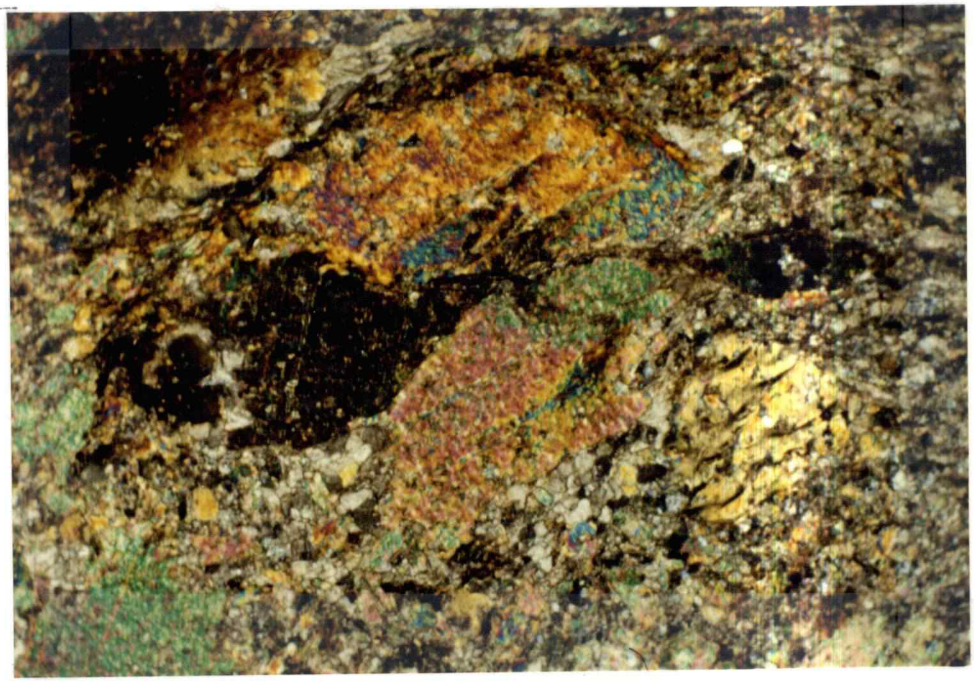
Figure 13 continued.

14). Thus subsequent to west-vergent deformation there was insufficent temperature to allow for complete annealing of deformed phlogopite/biotite.

GEOMETRY AND TIMING OF COMPRESSION

The oldest structures in the study area are mesoscopic east-vergent folds (F_1). Regional metamorphism began during the development of these folds, and after the cessation of deformation, was followed by static metamorphism. Following static metamorphism, the east-vergent folds and metamorphic fabrics were overprinted by small-scale west-vergent folds (F_{2x}), which are coaxial with earlier east-vergent folds, and minor thrust faults. The development of the upright folds (F_{2v}) occurred after east-vergent deformation. Undated granitic and andesitic dikes discordantly cut compressional fabrics, mesoscopic and map-scale structures (Plate 1), indicating that igneous activity post-dates compressional deformation.

At the southernmost extension of the metamorphosed and ductilely deformed rocks in the Grant Range, in the southernmost Grant Range, Cebull (1970) and Fryxell (1984) suggested that metamorphosed upright to overturned Cambrian strata form an east-vergent overturned anticline. The rocks that compose the anticline contain mesoscopic folds that are coaxial with those in the study area (Fryxell, 1984); however, from descriptions of these folds it is not clear whether the mesoscopic folding in the southern Grant Range occurred before or during the generation of the map-scale anticline. This anticline is cut by a 70 Ma pluton (Fryxell, 1984). Fryxell (1984) mapped a small-scale, east-dipping, west-vergent thrust fault that discordantly cuts the map-scale anticline and that, in



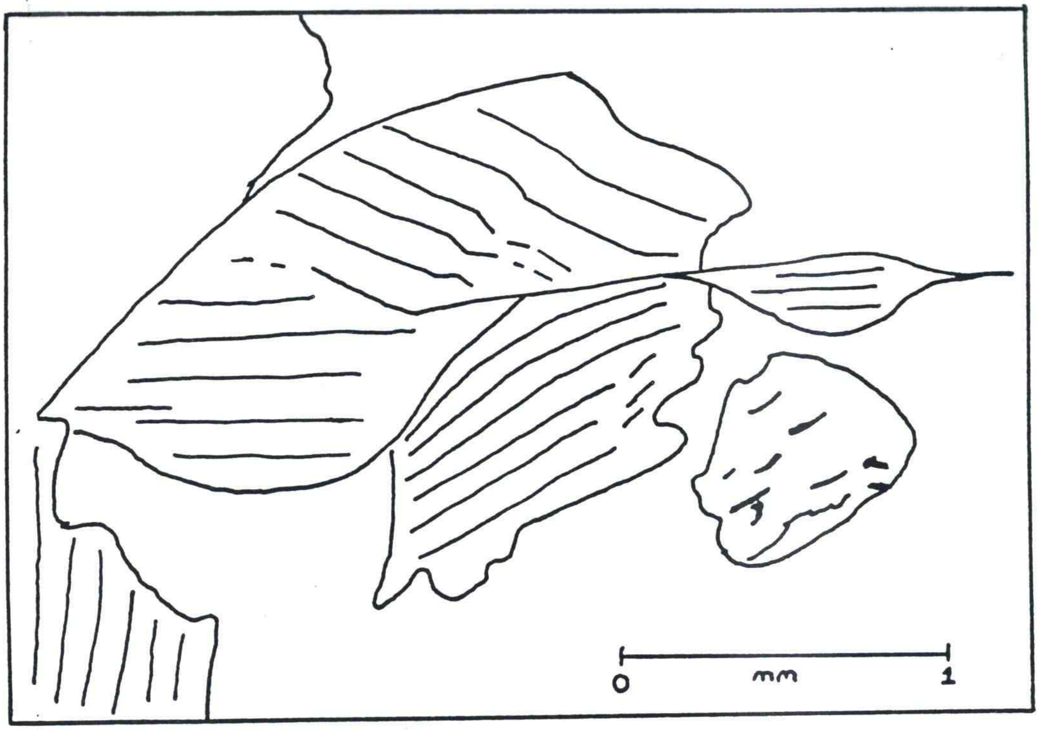


Figure 14. Deformed phlogopite poikiloblasts in the Willow Springs member. From an F_{2x} fold (a refolded F_1 fold) in the upper plate of the thrust fault on the north side of Heath Canyon. Large grains in the center were formerly one grain. Mineralogy is phlogopite and calcite. Crossed polars.

turn is cut by the 70 Ma pluton. Fryxell (1987) suggests that metamorphism began during folding and that static metamorphism followed folding, but that most of the metamorphism took place prior to the intrusion of the 70 Ma pluton. The sequence of deformation, metamorphism, and intrusion in the southern Grant Range (Fryxell, 1987) is similar to that discussed in this paper. Thus, based on correlation of the sequence of events in the study area with those in the southern Grant Range, compressional deformation and regional metamorphism in the study area probably pre-date 70 Ma. Although there are no direct dates on compressional deformation in the study area, a Mesozoic age can be inferred from regional evidence. Regionally, the ages of compressional deformation in the eastern Great Basin have been determined in extensional "blocks" which expose Paleozoic metasedimentary rocks that were metamorphosed during compressional deformation. Structures and metamorphic fabrics in metasedimentary rocks in these areas are generally associated with syn- to post- to pre-kinematic intrusives. Radiometric dating of minerals in these intrusive rocks as well as metamorphic minerals in metasedimentary rocks reveals Mesozoic ages of compressional deformation (Miller et al., 1988; Smith and Wright, 1988; Miller et al., 1987; Dallmeyer et al., 1986).

Rocks in the study area may be part of the upper limb of the east-vergent overturned anticline recognized in the southern Grant Range, but there is no compelling evidence for this. In the Quinn Canyon Range (Fig. 1), Bartley and Martin (1986) and Bartley et al. (1987) have recognized an east-vergent thrust fault that emplaces Cambrian and Ordovician strata over Devonian strata. This fault cuts up-section in the hanging wall and footwall. The hanging wall of this thrust fault contains west-vergent back-folds and back-thrust faults, which in places overprint earlier east-vergent structures (Bartley and Martin, 1986). Rocks in the study area record a similar compressional history and may represent the hanging wall of a geometrically similar (or the same ?) thrust fault that was subsequently extensionally dismembered. Figure 15 shows one possible pre-extension compressional geometry.

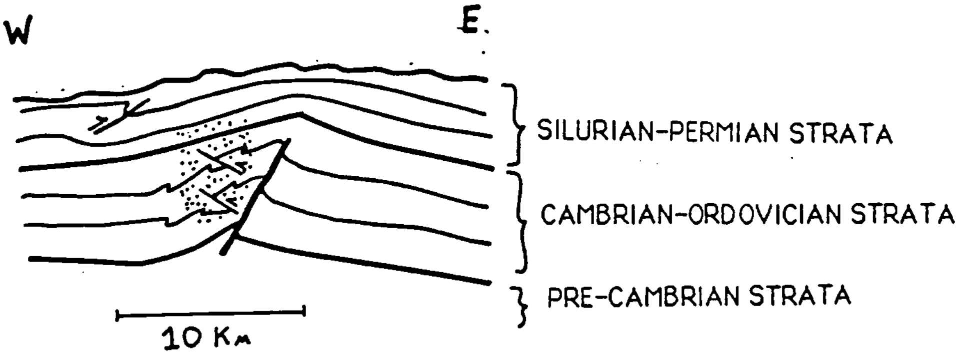


Figure 15. Sketch of one possible pre-extension compressional geometry of rocks in the study area. The stippled area represents rocks that are now exposed in the study area. This sketch shows mesoscopic east-vergent folds overprinted by west-vergent back-thrusts.

EXTENSIONAL STRUCTURES

INTRODUCTION

Mesozoic ductile compressional structures are cut by low- to high-angle normal faults of a more brittle character (see Plates 1 and 2 for cross-cutting relationships). Thus low-angle normal faults post-date compressional deformation and regional metamorphism. The intrusion of granitic and andesitic dikes post-dates compressional deformation, however, some dikes cross-cut various low-angle normal faults or are cut by a low-angle normal fault (Plate 1). This indicates that igneous activity in the study area is post-compressional and at least in part synchronous with low-angle normal faulting.

The pre-extension geometry of the rocks in the study area is not known; however, the rocks could have composed the upright, west-dipping limb of an east-vergent overturned anticline or could have composed the hanging wall of an east-vergent thrust fault (see discussion in section entitled "geometry and timing of compression").

Six major low-angle normal faults are recognized within the study area (Figs. 16 and 17 and Plates 1 and 2). The low-angle normal faults are broadly arched about a north trending axis. These faults omit stratigraphic section and most of the faults cut down section to the west. As a generalization, low-angle normal faults young structurally and stratigraphically upwards, that is, any particular low-angle normal fault cuts the one beneath it and emplaces younger rocks on top of older rocks. Because metamorphic grade, and in general the degree of ductile strain, dies out stratigraphically upwards, the older faults juxtapose

Guilmette Formation (Devonian) and Joana Limestone (Mississippian)

Ely Springs Dolomite (Ordovician), Laketown Dolomite (Silurian), and Sevy Dolomite and Simonson Dolomite (Devonian)

Eureka Quartzite, Lehman Formation, Kanosh Shale, Shingle Limestone, and Parker Spring Formation (Ordovician)

Goodwin Limestone (Cambrian to Ordovician), Little Meadows formation, and Blue Eagle member (Cambrian)

Grant Canyon member (Cambrian)

Willow Spring member and Pole Canyon Limestone (Cambrian)

Granitic or andesitic dike

Fault "A"

Fault "B"

Fault "C"

Fault "D"

Fault "E"

Fault "F"

Thrust fault

High-angle normal fault

"Range-front fault"

General orientation of bedding

Figure 16. Tectonic map of the study area. Unpatterned areas are Quaternary alluvium. This map extends to the west of the area shown mapped in Plate 1: the additional mapped area represented in this figure is from Kleinhampl and Ziony (1985).

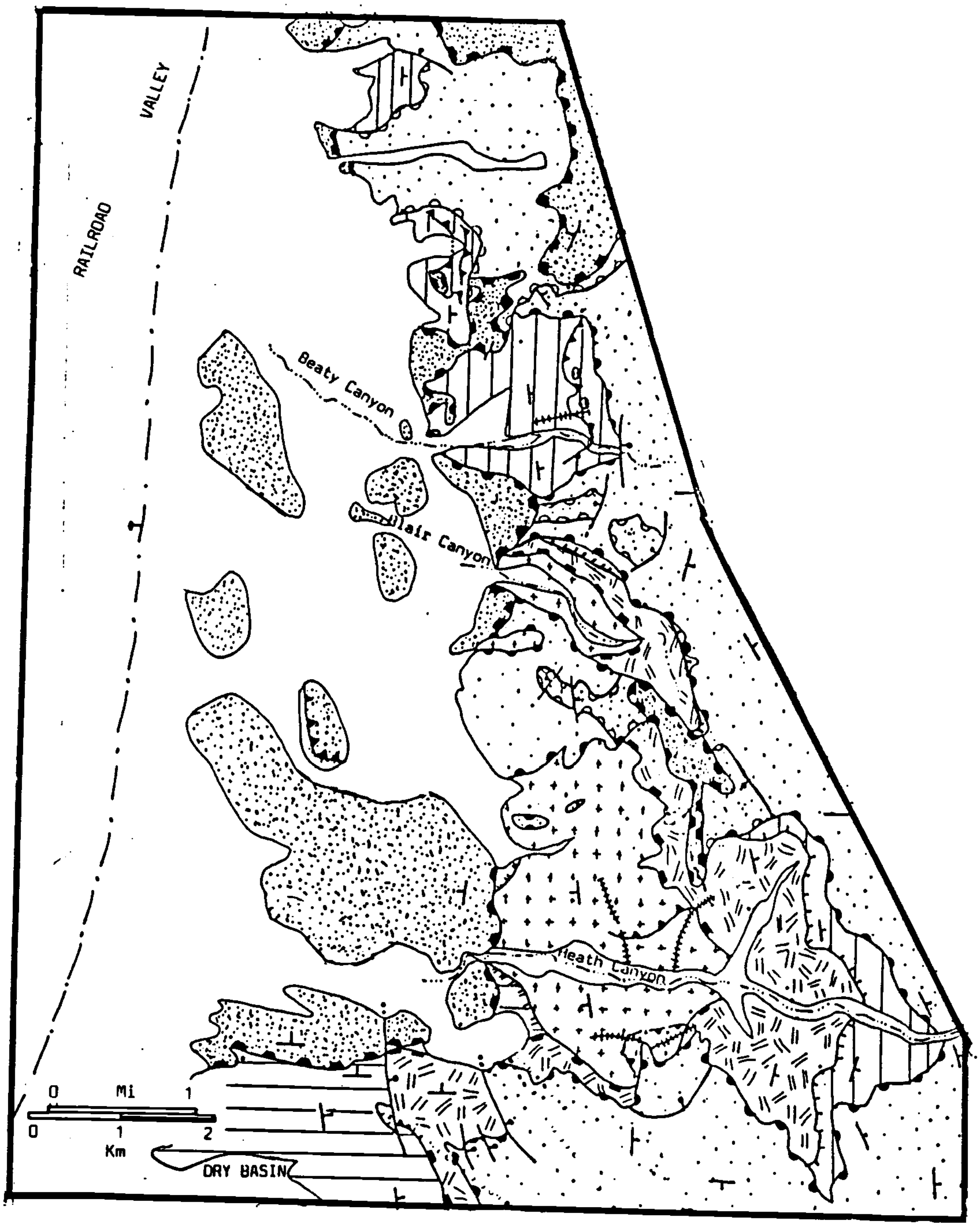
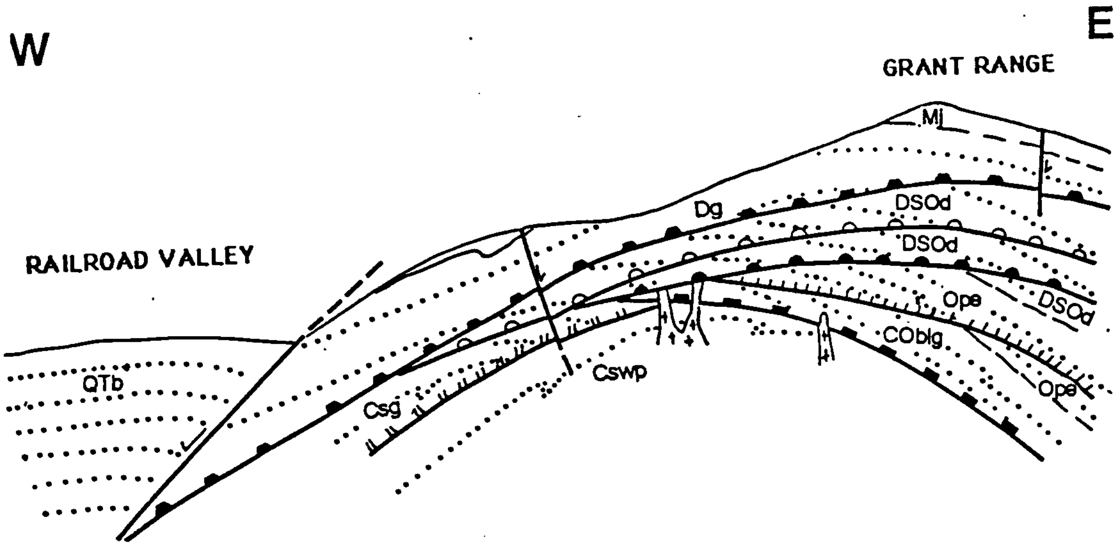


Figure 16 continued.



QTb "Basin-fill" strata

Dike (Tertiary ?)

Mj Joana Limestone (Mississippian)

Dg Guilmette Formation (Devonian)

DSOd Ely Springs Dolomite (Ordovician), Laketown Dolomite (Silurian), and Sevy Dolomite and Simonson Dolomite (Devonian)

Ope Eureka Quartzite, Lehman Formation, Kanosh Shale, Shingle Limestone, and Parker Spring Formation (Ordovician)

CObLg Goodwin Limestone (Cambrian to Ordovician), Little Meadows formation, and Blue Eagle member (Cambrian)

Csg Grant Canyon member (Cambrian)

Cswp Willow Spring member and Pole Canyon Limestone (Cambrian)

Fault "A"

Fault "B"

Fault "C"

Fault "D"

Fault "E"

Fault "F"

Figure 17. Schematic cross-section of the extensional geometry of the study area. This cross-section extends to the east and west of the study area; this was done to show geometric relationships at depth to the east and west of the study area. Data used in construction of this cross-section are from this study, Effimoff and Pinezich (1981), and published geologic maps by Kleinhampl and Ziony (1985) and Lund _et al_. (1987).

less metamorphosed/deformed rocks over more metamorphosed/deformed rocks, and the youngest fault juxtaposes younger, brittlely deformed unmetamorphosed rocks over older, ductilely deformed metamorphosed rocks (Fig. 17). Where rocks above a low-angle normal fault are dolomite or thick-bedded massive limestone, they tend to be pervasively brecciated and where these rocks are thin-bedded limestone or argillaceous limestone, little brecciation is evident. Near where low-angle faults intersect, intervening rocks tend to be thoroughly brecciated or cut by mesoscopic low- to high-angle normal faults; where low-angle normal faults diverge, intervening rocks lack such pervasive deformation.

LOW-ANGLE NORMAL FAULTS

Fault "A"

Fault "A" is the oldest fault in the study area. Fault "A" is west-dipping and juxtaposes the Grant Canyon member atop the older Willow Springs member (Plate 1). Beneath fault "A", the Willow Springs member (an "argillaceous limestone") is pervasively folded, recrystallized and contains amphibole and phlogopite. In contrast, the structurally overlying Grant Canyon member contains only a few mesoscopic folds and is composed of generally unrecrystallized fossiliferous argillaceous limestone and minor phyllite. Thus fault "A" juxtaposes less metamorphosed/deformed younger rocks over more strongly metamorphosed/deformed older rocks. Fault "A" is cut by faults "B" and "F" (Plates 1 and 2). Fault "A" omits stratigraphic section, however, because of the limited areal extent of this fault in the study area, the geometry and kinematics of this structure are not constrained.

Fault "B"

Fault "B" juxtaposes the Upper Cambrian Blue Eagle member, Little Meadows formation and Upper Cambrian to Lower Ordovician Goodwin Limestone on top of the middle Cambrian Pole Canyon Limestone and Willow Springs member. Near fault "B", the Blue Eagle member, Little Meadows formation and Goodwin Limestone are cut by several low-angle faults (Plate 1), which are hereafter collectively refered to as the "fault zone" (FZ). Just north of Heath Canyon, a small slice of the underlying Pole Canyon Limestone has been incorporated into the FZ (Plate 1). Above the FZ, strata in the Goodwin Limestone are not deformed.

In general, rocks in the FZ lack overt signs of brittle deformation, except in the few places where rocks of the FZ are brecciated near where fault "B" is cut by younger faults. For example, northeast of Dry Basin where fault "B" is in proximity of faults "D" and "F" (see cross-section E-E′ on Plate 2) the rocks in the upper plate of fault "B" tend to be pervasively brecciated.

Fault "B" is a gently undulating surface and is broadly arched about a north-northwest trending axis (Fig. 18). The maximum dip on the fault is 19° (derived geometrically from Fig. 18). Where fault "B" dips east (see Fig. 18), the Goodwin Limestone, which is stratigraphically above the network of fault slices at the base of fault "B", dips an average of about 25° east. This suggests that strata in the upper plate of fault "B" dip more steeply than the fault and that the fault cuts up section in its upper plate. However, fault "B" cuts down section to the east in its lower

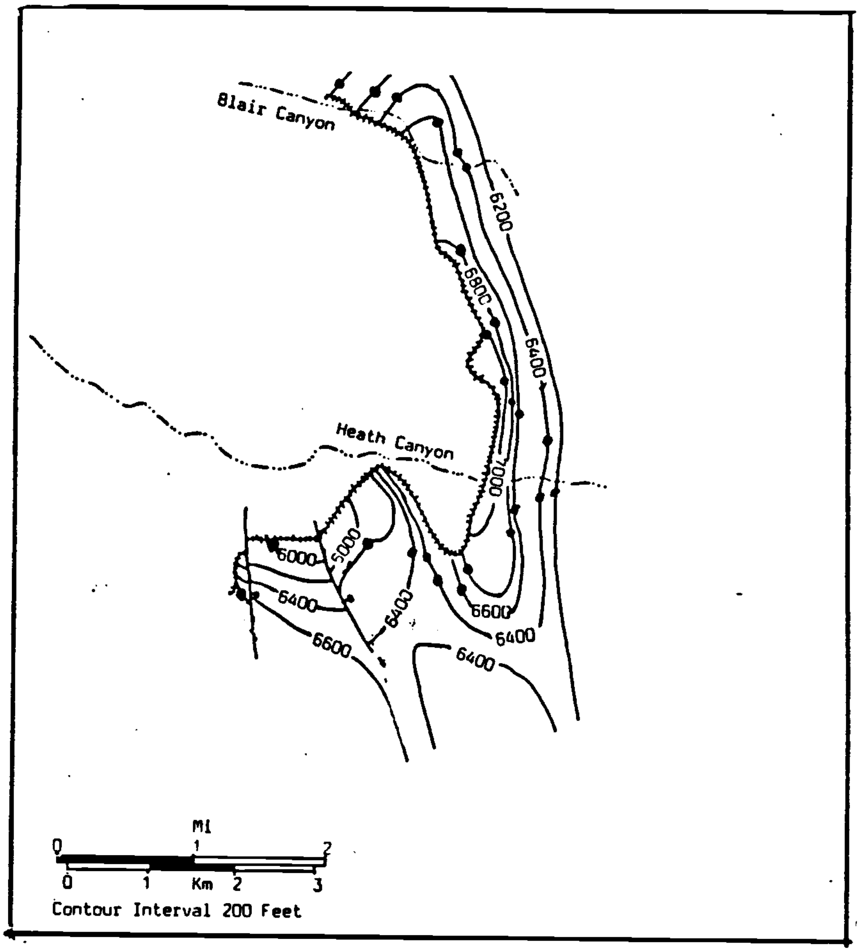


Figure 18. Structure contour map of fault "B". Sea level datum. Solid circles represent data points used in contouring: these data were taken from Plate 1. The hatchured line represents the approximate points at which fault "B" is cut by younger low-angle normal faults.

plate (Plate 2). This kinematics of this fault are not clear. In addition, where the fault that juxtaposes the Little Meadows formation atop the Blue Eagle member (this fault forms part of the FZ) intersects Blair Canyon it is well exposed. Here, this fault contains fault striae that have an average plunge of 25° in a southwest direction; this lineation, however, can not be used as a reliable kinematic indicator of fault movement because it is nearly perpendicular to the north to northwest trending axis of the flexure in the fault surface (Fig. 18) and therefore may be related to flexing rather than faulting.

Fault "B" cuts fault "A". Several granitic dikes cut fault "B" near Heath Canyon (Plate 1), one of these dikes is cut by fault "D" indicating that fault "D" is younger.

Fault "C"

Fault "C" is exposed from Blair Canyon to the south side of Heath Canyon. Fault "C" lies structurally above fault "B" and juxtaposes Ordovician Parker Spring Formation over Cambrian to Ordovician Goodwin Limestone. There are no exposed cross-cutting relationships between this fault and fault "B" so relative age relationships are unknown. Fault "C" is cut by and thus older than fault "D".

Fault "C" is broadly arched. In Heath Canyon it dips towards the east, whereas to the north in the Blair Canyon area, the dip of the fault shallows and begins to dip northward (see Plate 2, cross-section C-C'). The overall geometry of fault "C" appears to be similar to that of Fault "B" (Fig. 18). Fault "C" dips approximately 24° northeast in Heath Canyon (attitude derived from contouring the fault plane). In this area, bedding in the upper plate dips an average of about 35° east (Plate 1); dips of bedding in the underlying Goodwin Limestone ranges from 17° to 38°, but in proximity of fault "C" dips generally exceed that of the fault. In addition, fault "C" is well exposed approximately 1.5 km south of Heath Canyon. Here, it can be seen that strata above and below the fault dip more steeply eastward than the fault. These geometric relationships would suggest that fault "C" cuts down section to the west in both the upper and lower plates, and that the sense of stratigraphic offset is top-to-the-west; this is displayed in Plate 2 cross-section E-E′ and shown diagrammatically in Figure 17.

Fault "D"

The next youngest and structurally higher fault is fault "D". Fault "D" juxtaposes the Ordovician Ely Springs Dolomite and Silurian Laketown Dolomite on top of older Cambrian and Ordovician rocks (Plate 1; Fig. 19). From east to west, successively older Ordovician to Cambrian rocks are exposed in the lower plate (Plates 1 and 2).

Fault "D" is broadly arched about a northwest trending axis (Fig. 20). The maximum dip on the fault is approximately 17° east (derived geometrically from Fig. 20) in the eastern part of the study area (Fig. 20). In this area, bedding in the upper plate of the fault dips an average of 25° east and in the lower plate dips an average of 35° east suggesting that the fault cuts down-section to the west in its lower and upper plates. Although offset on fault "D" is not exposed, the sense of stratigraphic offset across the fault is top-to-the-west. This relationship is best observed

Figure 19. Photograph of fault "D" on the south side of Blair Canyon. Upper plate rocks are thoroughly brecciated Silurian Laketown Dolomite and lower plate rocks are sheared Cambrian-Ordovician Goodwin Limestone. Notebook is 7 inches long.

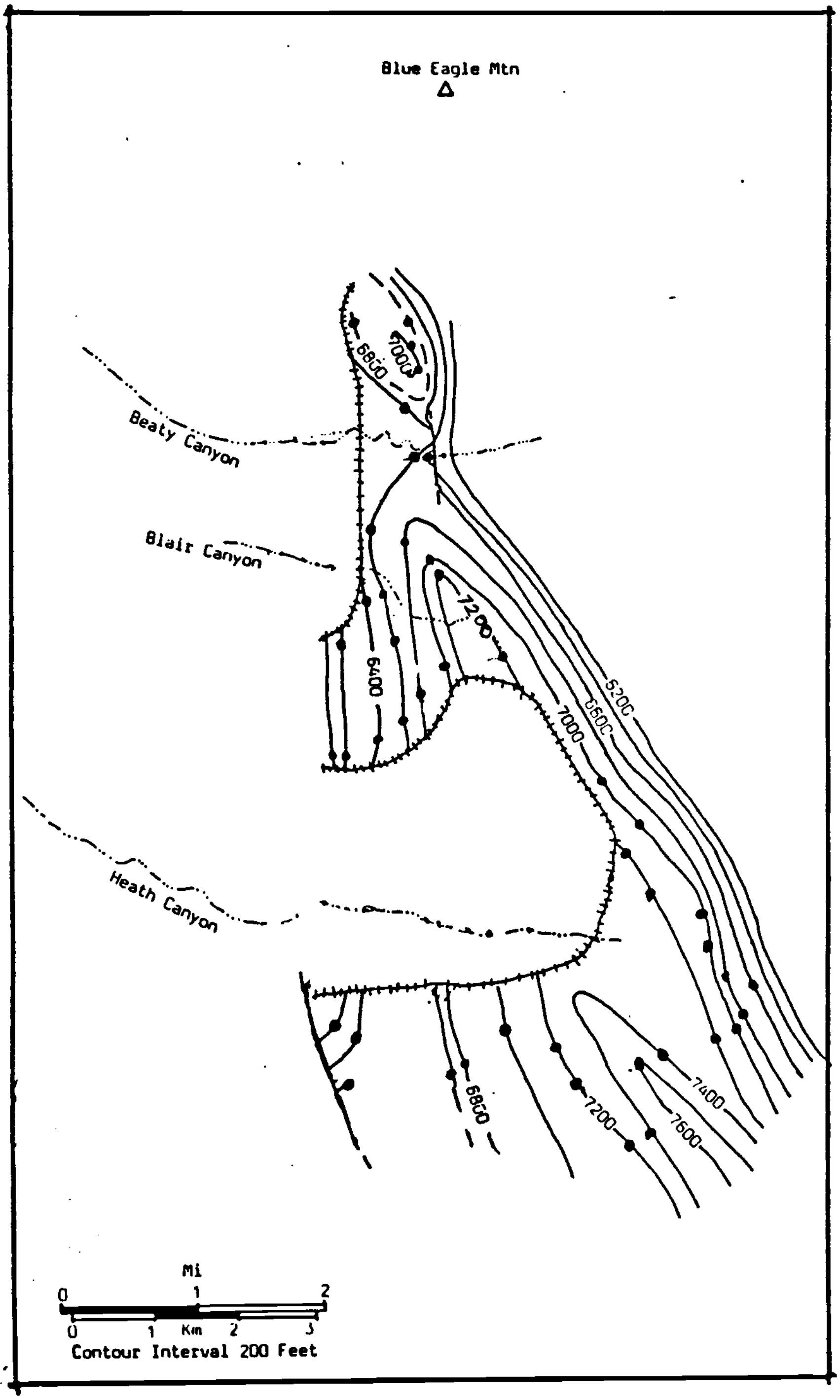


Figure 20. Structure contour map of fault "D". Sea level datum. Solid circles represent data points used in contouring; these data were taken from Plate 1. The hatchured line represents the approximate points at which fault "D" is cut by younger low-angle normal faults.

where fault "D" intersects Beaty and Heath Canyons. Here, bedding above and below the fault dip more steeply east than the fault thus it appears that upper plate rocks moved towards the west relative to lower plate rocks (see Plate 2 cross-sections E-E′ and B-B′ and Fig. 17).

Fault "E"

Fault "E" juxtaposes Devonian Sevy and Simonson Dolomites on top of Silurian Laketown Dolomite to the east, and to the west juxtaposes Sevy and Simonson Dolomite, Laketown Dolomite, and Ordovician Ely Springs Dolomite on top of older Ordovician rocks (Plate 1). Fault "E" cuts down-section to the west in its upper and lower plates (Plate 1). Fault "E" cuts fault "D" and is cut by fault "F" (Plate 1 and Plate 2 [cross-section A-A′]).

Fault "E" is arched, this relationship is best observed due west of Blue Eagle Mountain (Plate 1). Here, the fault dips east at its easternmost exposure and to the west it dips west. In the eastern part of the study area, where considerable thickness of the upper plate of fault "E" is preserved, bedding in the upper plate dips to the east, in the western part of the study area near where fault "E" is in proximity of being cut by fault "F", rocks in the upper plate of fault "E" tend to be thoroughly brecciated and bedding is not discernable. On the western flank of the Grant Range just north of the study area, however, rocks in the upper plate of fault "E" have a prominent west dip.

Fault "F"

Fault "F" cuts fault "E", cuts down-section to the west, and juxtaposes Guilmette Formation and Joana Limestone on top of older Cambrian to Devonian rocks (Plates 1 and 2). Fault "F" dominantly dips to the west, however to the east, the dip of the fault shallows (Plates 1 and 2), and to the east of the study area the fault begins to dip very shallowly (10° or less) to the east (Lund, 1988, personal communication). Bedding above the fault is generally west-dipping and nearly parallel to the fault (Plates 1 and 2).

MESOSCOPIC AND HIGH-ANGLE NORMAL FAULTS

In a few places, rocks in the study area contain small-scale, high- to low-angle mesoscopic normal faults (Fig. 21). These structures overprint compressional fabrics and tend to occur in proximity of the intersections of major low-angle normal faults. Where displacement on these structures is discernable, based on offset layering, it is commonly down- or top-to-the-west. The age and relationship of the mesoscopic normal faults relative to the major low-angle normal faults is not constrained because cross-cutting relationships were not found.

Rocks in the study area are cut by high-angle normal faults that have dominantly north-northwest to north-northeast trends, dip to the east or west, and have minor displacements (Plate 1). Down-to-the-east faults predominate. Many of these faults cut, and therefore are demonstrably younger than, the low-angle normal faults. However, some high-angle normal faults occur entirely within the upper plate of a low-angle normal fault, or appear to sole into a low-angle normal fault. These faults can be older than, synchronous with, or younger than the major low-angle normal fault (s).

Figure 21. Mesoscopic normal faults in the Pole Canyon Limestone from the north side of Blair Canyon.

METAMORPHISM

Metamorphism during extension was predominantly localized along low-angle normal faults.

In various places, poorly exposed phyllite with superposed foliations and marble tectonite occurs at the base of faults "B" and "A". Phyllite at the base of fault "A" is from the Grant Canyon member, and at the base of fault "B" it is from the Pole Canyon Limestone. In thin section, the phyllite exhibits S-C fabrics with synkinematic white mica and sheared chlorite. Pelitic rocks in these same units away from these faults do not exhibit such fabrics thus I suspect that at least part of the deformation-metamorphism may be related to movement on faults "B" and "A".

Limestone adjacent to low-angle faults tends to be altered. Areas of alteration range from 1 dm to several meters thick and tend to occur in zones adjacent to and parallelling the structure. The types of alteration are (1) dolomitization of limestone--commonly associated with tiny (~2 mm thick) quartz veins, (2) silicification, and (3) orange to yellow iron oxide staining. In addition, metamorphic biotite or phlogopite tends to be altered to chlorite in proximity of low-angle faults. Silicified breccias are common along faults that cut thin-bedded limestone. The breccias contain ghosts of randomly oriented limestone clasts that are typically 2 to 5 cm long. The only place along any one particular fault where such a breccia occurs is where it is silicified, thus making a hydrothermal origin more probable than a tectonic origin.

GEOMETRY AND TIMING OF EXTENSION

Low-angle normal faults are broadly arched about north trending axes. All of the faults omit stratigraphic section and structures appear to young structurally and stratigraphically upwards. Although offsets across the low-angle normal faults are not directly observable in the field, the sense of offset on faults "C" through "F" can be inferred geometrically: these faults appear to be west-directed. There is no reliable kinematic information on faults "A" and "B". However, fault "B" cuts down-section to the east in its lower plate and could be an east-directed fault. Nevertheless, the low-angle normal faults in the study area form a unique pattern (Fig. 17) and this suggests that the faults are geometrically and kinematically related.

Compressional structures probably pre-date 70 Ma (see discussion in section entitled "geometry and timing of compression"); thus, the cross-cutting low-angle normal faults probably post-date 70 Ma. Although it might be possible to constrain the age of some of the low-angle normal faults by dating dikes that cross-cut or are cut by the faults, at present there are no direct dates on the timing of extension in the study area. However, some broad time constraints can be placed on extensional structures in the study area through examination of work in other regions.

In the southern Grant Range, Fryxell (1984) has recognized two generations of low-angle normal faults. The earlier generation is east-dipping and is inferred to be east-directed, whereas the latter generation is west-dipping and interpreted to be west-directed. Based on cross-cutting relationships of low-angle normal faults with dated volcanic rock, Fryxell (1987) suggests that the east-directed faults are post-late Oligocene and that the west-directed faults are Miocene or younger (one of the west-directed low-angle normal faults mapped by Fryxell (1984) contains east-dipping Miocene-Pliocene basin-fill strata in its upper plate; Fig. 1). In the northern Grant-southern White Pine Ranges (Fig. 1), the west-directed moderate to low-angle normal faults that contain east dipping basin-fill strata in their upper plates are Miocene or younger in age (Moores et al., 1968). Faults in the study area that appear to be west-directed (faults "C" through "F") may be the same age as west-directed low-angle normal faults to the north and south of the study area; faults "A" and "B" may be temporally correlative with Fryxell's first or second generation of low-angle normal faults.

The arching of the low-angle normal faults could have occurred synchronous with and/or after low-angle normal faulting. The arching pre-dates deposition of Pleistocene(?) pluvial deposits (Kleinhampl and Ziony, 1985) that rest on the western margin of the study area (mapped as "Qu" on Plate 1) because these strata are not tilted.

Thus based on the regional data, low-angle normal faults in the study area can be loosely bracketed as Oligocene to Pleistocene in age, and may be dominantly Miocene or younger in age, broadly coeval with the development of Railroad Valley and the Grant Range. The high-angle normal faults and mesoscopic normal faults can only be bracketed as Oligocene (?) to Recent.

CONCLUSIONS

RELATIVE CHRONOLOGY OF STRUCTURES

The chronology of tectonic events in the study area is shown diagrammatically in Figure 22. The oldest structures in the northwestern Grant Range are mesoscopic east-vergent folds (F_1) with axial-planar cleavage and axes that plunge gently in north to north-northwest and south to south-southeast directions. Regional metamorphism began during the development of F_1 folds. Static metamorphism followed the first deformational event. Static metamorphism was then followed by the development of west-vergent folds (F_{2x}), which are coaxial with F_1 folds, and small-scale thrust faults. West-vergent deformation marks the end of regional metamorphism. Upright folds (F_{2y}) overprint east-vergent folds. Based on regional data, the compressional events can be considered Mesozoic in age and based on correlation of compressional structures in the study area with those in the southern Grant Range, compressional deformation and metamorphism is interpreted to pre-date 70 Ma. Granitic and andesitic dikes discordantly cut compressional fabrics or structures, thus igneous activity post-dates compressional deformation and regional metamorphism.

The development of dominantly brittle low-angle normal faults followed ductile compressional deformation and metamorphism. Some igneous activity took place during the formation of the low-angle normal faults as indicated by dikes that cut or are cut by the faults. The low-angle normal faults juxtapose younger rocks over older rocks and young structurally upwards. Metamorphism during extension is restricted to hydrothermal alteration, although some ductile deformation-

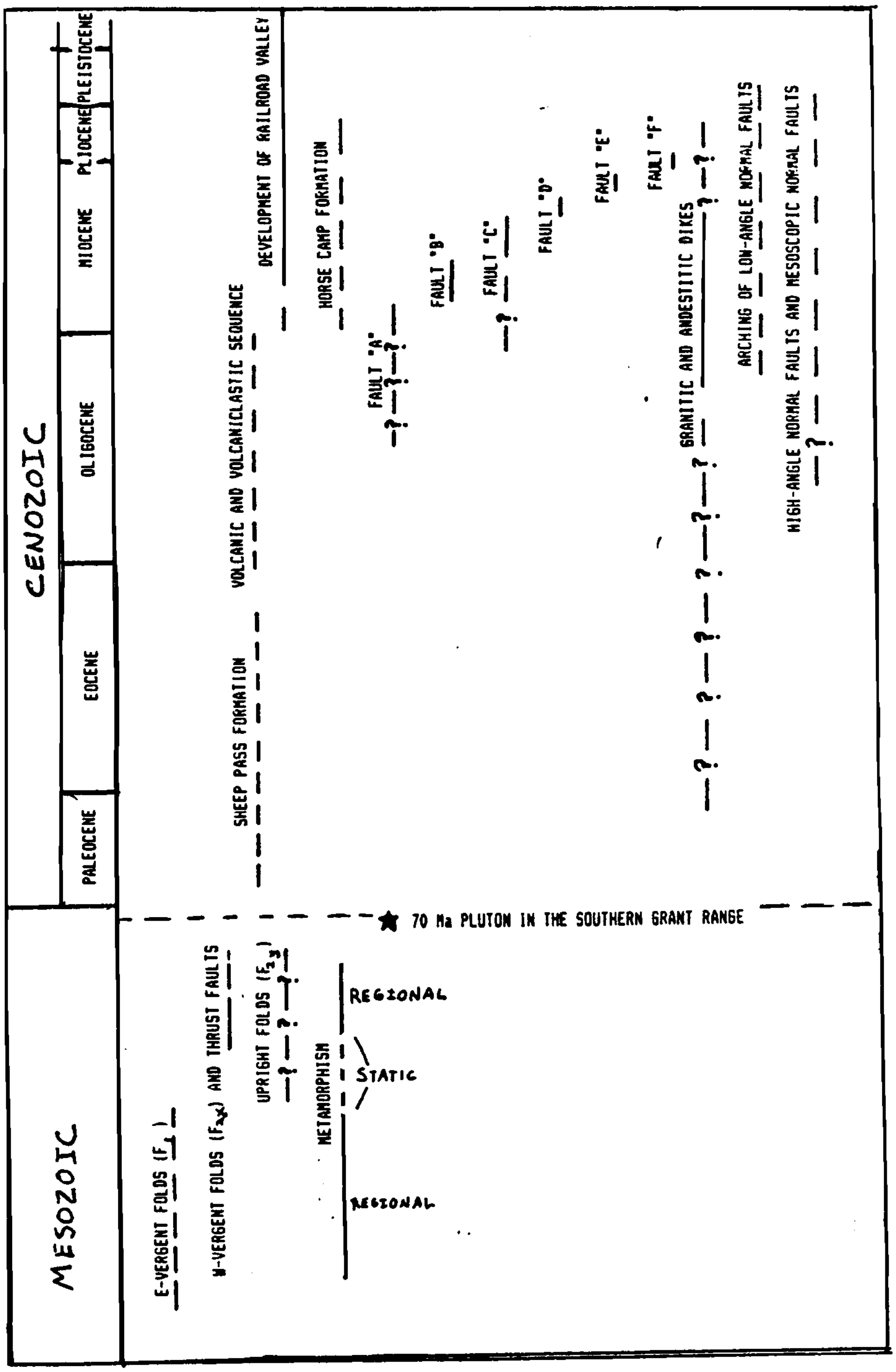


Figure 22. Diagram illustrating the chronology of events in the study area and adjacent areas. Data from this study, Moores et al. (1968), Fryxell (1984) and Fouch (1979).

metamorphism (as indicated by phyllite with "S-C" fabrics) along faults "A" and "B" may have occurred. The arching of low-angle normal faults may have occurred concurrently with or after faulting. Mesoscopic normal faults and high-angle normal faults that appear to sole into low-angle normal faults are either older, the same age as, or younger than the low-angle faults; their timing is not constrained. However, some high-angle normal faults cut low-angle normal faults and are therefore demonstrably younger. Arching of the low-angle normal faults pre-dates the deposition of Pliestocene pluvial deposits on the western flank of the range because they are not deformed. The age of the low-angle normal faults can only be loosely bracketed as Oligocene to Pleistocene; however, the occurrence of Miocene or younger west-directed low- to moderate-angle normal faults along the western flanks of the northern Grant-southern White Pine and southern Grant Ranges would suggest that the family of west-directed low-angle normal faults in the study area could also be Miocene or younger. Thus low-angle normal faulting and arching in the study area may be largely Miocene to Pleistocene in age.

MESOZOIC COMPRESSION

Mesozoic structure in the northwestern Grant Range consists of mesoscopic, open to isoclinal east-vergent folds that are overprinted by small-scale west-vergent thrust faults and folds. East-vergent folds may have composed the upper limb of a map-scale, east-vergent overturned anticline (such as described in the southern Grant Range by Cebull, 1970 and Fryxell, 1984); if this was the geometry, then west-vergent structures would have an unknown relation to this fold. However, it is possible that the rocks in the study area were part of the hanging wall to an east-vergent thrust fault, and west-vergent structures simply represent back-folds and back-thrust faults (Figure 15 shows one possible pre-extension geometry); perhaps a structure similar to that recognized by Bartley and Martin (1986) in the southern Quinn Canyon Range.

The rocks in the study area are cut by low-angle normal faults that omit stratigraphic section; thus, there is no major structural duplication of stratigraphic section. It is possible that metamorphosed rocks in the study area were never structurally buried very deep. The oldest unit exposed in the study area, the Pole Canyon Limestone, probably represents a stratigraphic depth of 6 km (based on thickness of equivalent, undeformed, unmetamorphosed Paleozoic stratigraphic units above and including the Pole Canyon Limestone in the southern Egan Range [Kellogg, 1963]). Mesoscopic folding and small-scale thrust faulting as well as the possible presence of Mesozoic rocks would have also thickened the section somewhat; thus, a 6 km paleodepth is a minimum estimate. If the rocks were never buried to depths significantly greater than stratigraphic depths then the heat needed for metamorphism may have come from friction created during deformation. However, other workers in the Great Basin (Smith and Wright, 1988; Miller et al., 1988) have documented Mesozoic compressional deformation and regional metamorphism as being synchronous with plutonism. These workers suggest that plutons locally provided the heat for metamorphism and associated ductile deformation. It is possible that during compression, rocks in the study area were being warmed by plutons at depth. There is evidence of Mesozoic plutonism in the southern Grant Range, although emplacement and cooling of the pluton post-dates compressional deformation (Fryxell, 1987). However this intrusive, and perhaps others, prior to ascent to their emplacement level in the crust, could have provided the heat necessary for metamorphism and ductile deformation.

CENOZOIC EXTENSION

Extensional Geometry

The low-angle normal faults in the northwestern Grant Range form a unique geometric pattern and several generalizations can be made to characterize this geometry (Plates 1 and 2 and Fig. 17):

(1) Low-angle normal faults omit stratigraphic section.
(2) Bedding-to-fault angles are generally in the range of 5° to 15°.
(3) Where low-angle normal faults dip to the east, bedding above and below a fault dip more steeply east than the fault; where these faults begin to dip to the west, the dip of bedding shallows and in some places becomes west-dipping.
(4) Although offset across the low-angle normal faults can not be observed in the field, faults "C" through "E" appear to be west-directed. Kinematics on faults "A" and "B" are poorly constrained.
(5) There is a general younging of structures structurally and stratigraphically upwards; the youngest (uppermost) fault contains the youngest rocks in its upper plate and cuts lower faults.
(6) The structurally highest and youngest faults have the most shallow dips; the youngest fault is west-dipping, although to the east of the study area it flattens and becomes gently east-dipping.

(7) Where low-angle normal fault slices taper, near where the faults intersect, intervening rocks tend to be remarkably attenuated by small-scale, high- to low-angle faults (mesoscopic extensional faults), and where these intervening rocks are dolomite they tend to be pervasively brecciated and fractured. Away from the tapered end of a fault slice--where the thickness of a fault slice increases--, rocks lack such pervasive deformation.

(8) Low-angle normal faults are broadly arched about north to northwest trending axes.

(9) Low-angle normal faults are cut by north trending, high-angle, down-to-the-east and -west normal faults of minor displacement.

Model for Extension

An extensional geometry similar to that of the northwestern Grant Range could be generated by progressive eastward rotation (caused by arching) of the upper reaches of a moderate- to high-angle normal fault with concomitant creation of new fault segments replacing inactive, rotated segments. Figure 22 is a geometric paradigm that illustrates this process. This model requires that the pre-extension geometry of lower to middle Paleozoic rocks exposed in the northwestern Grant Range was a homoclinal west-dipping section not transected by any major thrust faults. Such a west-dipping section might have been the west limb of a map-scale east-vergent anticline, the hanging wall of an east-vergent thrust fault (see discussion entitled "geometry and timing of compression" and Fig. 15), or might have overlain a west-dipping thrust ramp. Figure 23-A shows this possible pre-faulting geometry; both the first normal fault and bedding, which is upright, dip

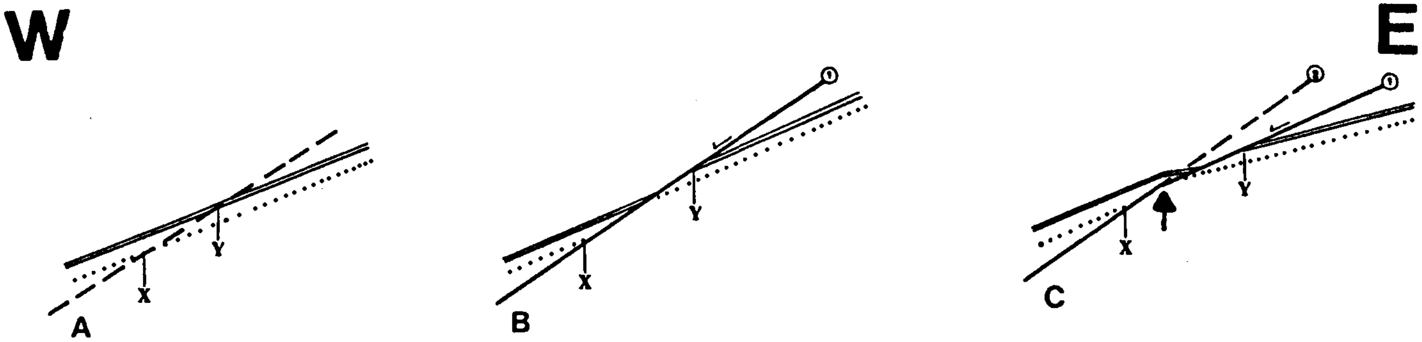


Figure 23. Geometric paradigm illustrating a possible mode of extension responsible for the generation of the extensional geometry in the study area. Bedding is represented by the stippled pattern and two parallel lines. A fault is shown by a solid bold line; a solid bold line is shown dashed where a fault is about to develop. Arrows indicate pivot points about which rocks and structures to the east of rotate in a clockwise direction. "X" and "Y" are fixed reference points. A. Bedding dips to the west and the initial fault, fault # 1, dips more steeply than bedding. B. Translation on fault surface #1. C. Arching of faults and rocks about the pivot point. This renders that part of fault #1 to the right of the pivot point inactive; this consequently results in the development of fault segment #2.

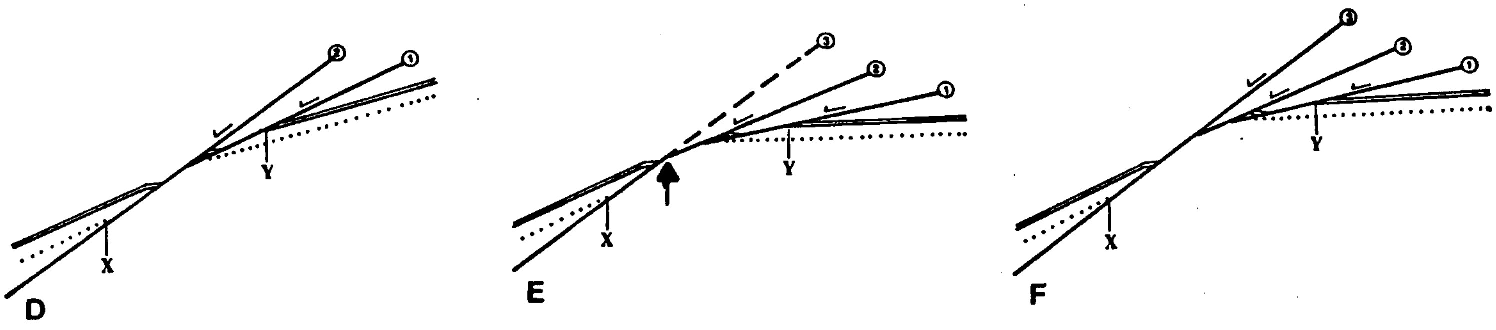


Figure 23 continued. D. Translation of fault surface #2. E. Arching about another pivot point. Arching--rotation--renders part of fault surface #2 inactive and results in the creation of a new fault segment (#3). F. Translation on surface #3.

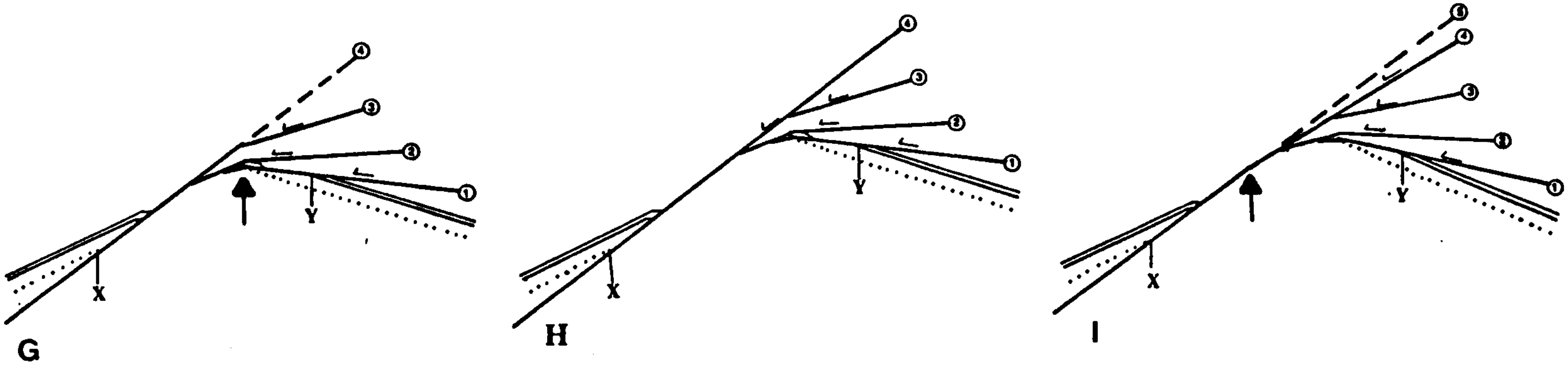
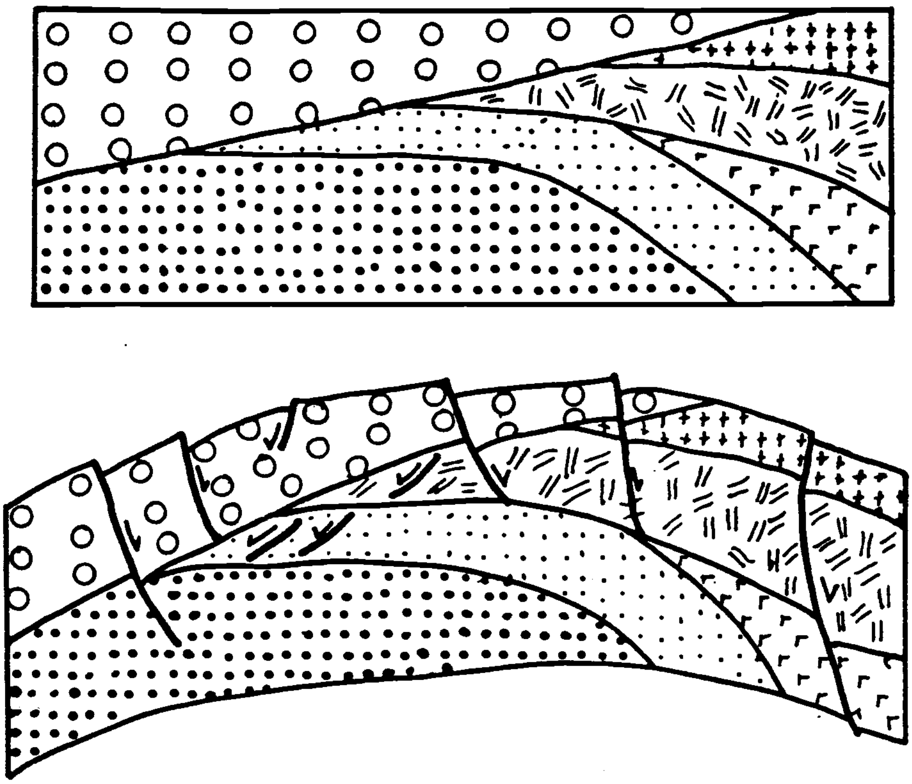


Figure 23 continued. G. Arching about another pivot point; part of fault surface #3 is rendered inactive and another fault segment is created--#4. H. Translation on surface #4. I. Arching about another pivot point, rendering part of fault surface #4 inactive and the consequent creation of surface #5.

west and bedding dips less steeply than the fault. Translation on the first normal fault is accompanied or followed by arching of the upper reaches of the fault rendering that particular part of the structure inactive (Fig. 23 C). This would result in the creation of a new fracture (Fig. 23 C) and subsequent translation on this new surface (Fig. 23 D). This process is then repeated (Fig. 23 E, F, G, H, and I). Note that progresssive arching of the upper reaches of the fault requires a general westward migration of the axis of arching with time. As a consequence of progressive eastward rotation of rocks and structures ("arching"), bedding and faults assume east dips and bedding dips more steeply than the faults (Fig. 23 I). This extensional process requires that structures become younger structurally upwards, that each successively higher fault contains rocks in its upper plate that are younger than the rocks below it, and that the youngest fault is west-dipping and contains the youngest rocks in its upper plate. This general sequence and geometry is observed in the northwestern Grant Range (Plate 2 and Fig. 17). Although kinematic information on faults "A" and "B" is lacking, these faults conform to the general geometry of the all of the low-angle normal faults; these faults could conceivably be west-directed as well. For example fault "B" may cut a small broad fold and locally cut down-section to the east, but regionally cuts up-section to the east. On the otherhand faults "A" and "B" may be east-directed and related to those recognized by Fryxell (1984) in the southern Grant Range. The minor high-angle and mesoscopic normal faults in the study area could be products of tension generated during progressive arching with extension (i.e. as in Fig. 23). Figure 24-A represents a hypothetical stack of wedge-shaped bodies of varying rheologies (which in essence represents the



a.

b.

Figure 24. Hypothetical origin of (1) the high-angle normal faults that cut low-angle normal faults, (2) high-angle normal faults that appear to sole into low-angle normal faults, and (3) mesoscopic normal faults. A. Stack of wedge-shaped bodies with varying rheologies. B. Flexing of this stack produces tension accommodated by the creation of normal faults, some of which transect rheologic boundaries and others that are restricted to certain rheologies or are incapable of transecting certain rheologic boundaries.

stack of wedge-shaped low-angle normal fault slices in the study area). Assuming that all of these wedge-shaped bodies lie above a neutral surface, flexing of this geometrically and rheologically anisotropic stack could produce tensional faults that transect or are restricted to various rheologies (the mesoscopic extensional faults) or even are incapable of transecting certain boundaries, and thus terminate at such boundaries (for example, a "secondary" high-angle fault created during arching may sole into instead of transecting a preexisting, extinct low-angle normal fault because of differences in rheologic properties at or across the fault surface [Fig. 24 B]). The development of these minor normal faults also could form with arching that progresses with "low-angle normal faulting".

The family of inferred west-directed low-angle normal faults in the study area appears to project into the west-directed "range-front fault" (shown diagrammatically in Figure 17). This suggests a possible relationship between the two systems. The age of the low-angle normal faults in the study area can only be crudely bracketed as Oligocene to Pleistocene and the "range-front fault" and its associated basin, Railroad Valley is Miocene to Recent in age. The age constraints on the two fault systems make a relationship between them permissable. It is possible that an extensional process similar to that shown in Figure 23 could be responsible for the uplift, and eastward tilting, of Miocene-Pliocene basin-fill strata ("Horse Camp Formation") on the western flanks of the northern Grant-southern White Pine and southern Grant Ranges (this process is shown diagrammatically in Figure 25). It is interesting to note that the lower part of the Horse Camp Formation in the northern Grant-southern White Pine Ranges records a folding event where basin-fill strata

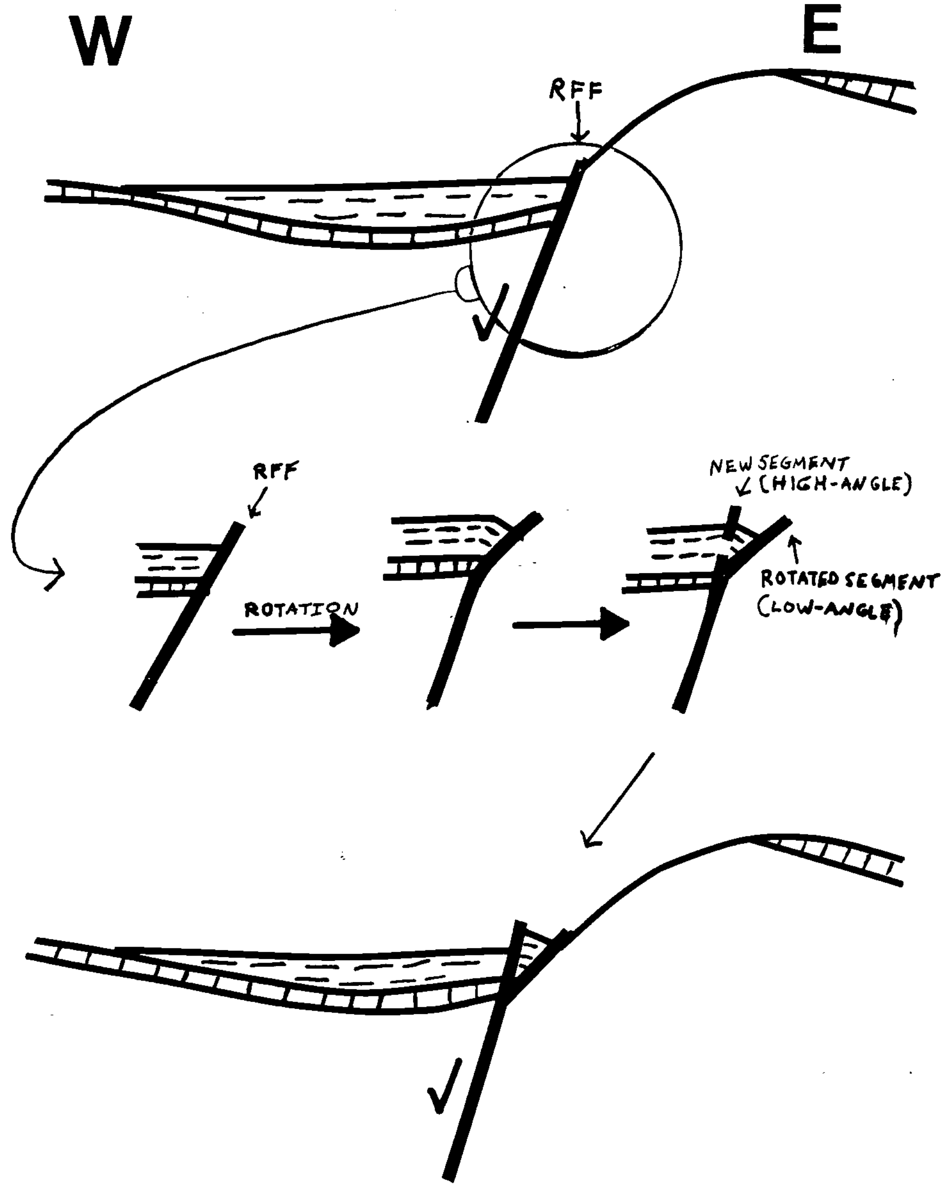


Figure 25. Diagram showing how arching of the upper reaches of the range-front fault (RFF), and consequent creation of a new fault segment replacing the rotated segment, can result in exposing east-tilted basin-fill strata. Dashed lines represent basin-fill strata. Vertically ruled pattern represents the volcanic and volcaniclastic sequence and Sheep Pass Formation. Unpatterned areas are Paleozoic rocks.

were arched about a north-trending axis (Moores et al., 1968). The folded strata were then partly eroded and subsequently deposition of the Horse Camp Formation resumed (Moores et al., 1968). This provides some evidence that folding, or arching, is synchronous with extension. Thus west-directed low- to moderate-angle normal faults on the western flanks of the Grant and southern White Pine Ranges may simply be products of paleo-eastward rotation, or arching, of the upper reaches of the range-bounding normal fault system. One possible mechanism for the arching of major normal faults, suggested by Spencer (1984), is isostatic uplift in response to tectonic denudation. Such a mechanism could be invoked for the low-angle normal faults in the study area, however, it is also possible that part of the arching, or eastward rotation, of the rocks and structures, is related to rotation along a coeval, major west-dipping fault located east of the Grant Range--perhaps the fault system that bounds the southern Egan Range to the east (Fig. 2).

Although the model for extension that I have presented in Figure 23 is one explanation of the extensional geometry in the study area, it needs to be tested by integrating the geometry and timing of extension to the east and west of the study area: this would require an attempt to regionally reconstruct the rocks to their pre-Basin and Range geometry. Moreover, if the extensionally dismembered middle to lower Paleozoic rocks on the western flanks of the Grant and southern White Pine Ranges, prior to extension, were once part of the hanging wall of an east-vergent thrust fault (e.g. a geometry like that shown in Fig. 15) then the possibility that a Mesozoic thrust fault was reactivated during extension needs to be investigated (i.e., is the base of the low-angle normal fault complex on the western flank of the Grant Range a reactivated thrust fault?).

REGIONAL IMPLICATIONS

COMPRESSION

Two compressional events are recognized in the study area: earlier east-vergent and later west-vergent deformation. The ages of compressional deformation can be bracketed only as pre-70 Ma and can be considered as Mesozoic in age. The recognition of a similar sequence of compressional deformation in the Quinn Canyon Range (Bartley and Martin, 1986), the southern Grant Range (Fryxell, 1984, 1987), and in the study area may indicate a belt of kinematically and temporally related structures. The extent and nature of this belt are poorly defined and await detailed mapping between these areas.

The sequence of compressional deformation in the Grant and Quinn Canyon Ranges contrasts with that recognized in the Snake and Schell Creek Ranges approximately 120 km northeast of the study area. Here the earliest compressional event is west-vergent and Jurassic in age and the youngest is east-vergent and Cretaceous in age (Miller et al., 1988). Because dates on compressional deformation in the study area and to the south are lacking, correlation of compressional events in the study area with those to the north is not possible. If west-vergent deformation in the study area is temporally correlative with that in the Snake and Schell Creek Ranges, then east-vergent deformation in the study area would be Jurassic or older in age! However, it is more likely that west-vergent structures in the study area are a local byproduct of major east-vergent deformation. Thus, east-vergent deformation in the study area, if correlative with the event to the north, is probably Cretaceous in age.

EXTENSION

Interpretations of the origin of low-angle normal faults in the eastern Great Basin varies from gravity slides (Boyer, 1987; Moores _et al_., 1968) to low-angle normal faults that formed at low-angles (Bartley and Wernicke 1984; Miller _et al_., 1983) or normal faults that were initially formed at high-angles and with progressive extension rotated to low-angles (Miller _et al_., 1983). This study suggests that the low-angle normal faults in the study area could have formed at higher angles and rotated to lower angles. The low-angle normal faults in the study area are interpreted to have formed as a consequence of progressive eastward rotation of the upper reaches of a major high- to moderate-angle west-dipping normal fault: the low-angle normal faults represent successive replacements of previously rotated parts of the major fault. This extensional process could be responsible for the stranding and eastward tilting of Miocene-Pliocene basin-fill deposits on the western flanks of the southern Grant and northern Grant-southern White Pine Ranges. The age constraints on the low-angle normal faults in the study area are broad, but these faults and the ones that contain the east-tilted basin-fill strata in their upper plates could be products of the normal fault system responsible for the development of the Grant Range and Railroad Valley.

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APPENDICES

APPENDIX A

METHODS

The field area was mapped at a scale of 1:12,000 on a base map made from enlargements of parts of the Troy Peak, Currant, Blue Eagle Springs, and Forest Home USGS 15′ quadrangles.

Nine weeks were spent doing field work during the summer of 1986 and four weeks during the spring and summer of 1987. Most of this time was spent in my thesis area, however, some time was spent looking at Cambrian and Ordovician sections described and mapped by: Humphrey (1960) in the Mt. Hamilton area, Kellogg (1963) in the southern Egan Range, Moores et al. (1968) in the southern White Pine Range, and Cebull (1967) and Fryxell (1984) in the southern Grant Range. I also spent time in various other localities from the southern White Pine Range to the southern Grant Range either to trace structures outside of my area or to observe certain geologic relations on maps of earlier workers.

In addition to the field work, fifty thin sections of metamorphic rock were examined to determine mineral assemblages and the relation of metamorphism to strain.

APPENDIX B

STRATIGRAPHY

PREFACE

Formation thicknesses were determined geometrically from the geologic map and bedding thicknesses given herein are visual estimates.

EUREKA QUARTZITE

The Eureka Quartzite (Hague 1883, 1892) is approximately 60 m thick and is composed of thin- very thick-bedded, white to black to reddish-brown quartzite. Sedimentary structure within beds is generally not discernable, however, in one place tabular crossbeds were observed.

Argillite (up to 3 dm thick) and sericite phyllite (less than 3 cm thick) layers are interbedded with quartzite in the bottom 3 m of the Eureka Quartzite. The top of the Eureka Quartzite is marked by a 3 dm thick bed of brownish red quartzite and the contact with the overlying Ely Springs Dolomite is sharp.

POGONIP GROUP

Introduction

The Pogonip Group exposed in the northern Grant Range was heretofore undivided. I have recognized six mappable units within what was previously mapped as as the Pogonip Group by Hyde and Huttrer (1970). The upper five of these units have been collectively assigned to the Pogonip Group. The lowermost unit is correlative with the Little Meadows formation of Cebull (1967).

I have correlated the upper four units of the Pogonip Group with similar units in the southern Egan Range (Fig. 25), which were mapped and described by Kellogg (1963). These units are the Lehman Formation (Hintze, 1952), Kanosh Shale (Hintze, 1952), Shingle Limestone (Kellogg, 1963), and Parker Spring Formation (Kellogg, 1963). In the southern Grant Range, Cebull (1967) divided the Pogonip Group into three formations and used names for similar strata in the Eureka area; the Antelope Valley Limestone (Nolan _et al_., 1956), Ninemile Formation (Nolan, et al., 1956), and Goodwin Limestone (Nolan _et al_., 1956). The rocks that Cebull mapped as the Antelope Valley Limestone and Ninemile Formation are correlative with my upper four units of the Pogonip Group (Fig. 26). In the northern Grant Range, Ordovician and Cambrian strata below these four units do not correlate with the stratigraphy established in the Egan Range, but do correlate with the stratigraphy established by Cebull (1967). Thus, in the study area the lowest unit in the Pogonip Group is the Goodwin Limestone. This formation is correlative with rocks that Cebull (1967) mapped as two separate units: the Cambrian-Ordovician Goodwin Limestone and the Cambrian Windfall Formation.

The best place to observe the Lehman Formation, Kanosh Shale, Shingle Limestone, and Parker Spring Formation is in Beaty Canyon. The Goodwin Limestone is best exposed in Heath Canyon.

Lehman Formation

The Lehman Formation is approximately 137 m thick

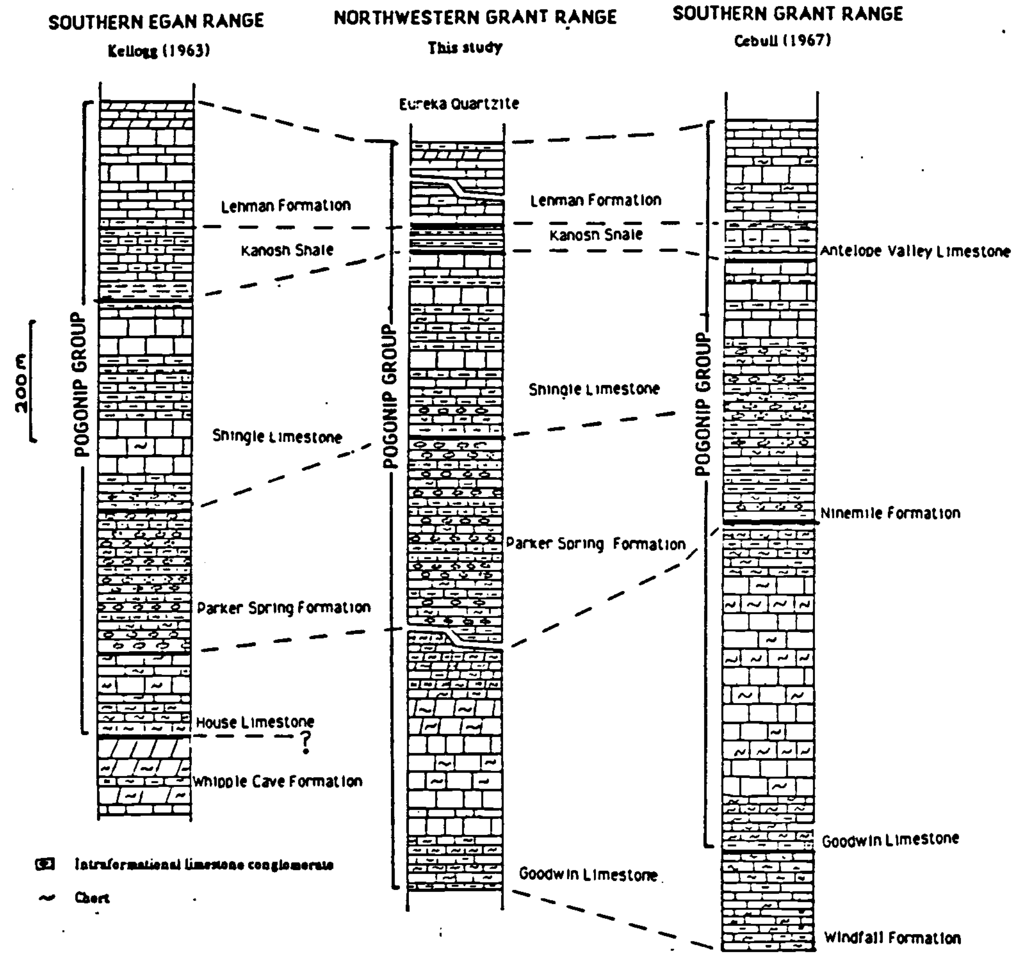

Figure 26. Correlation of the Pogonip Group in the study area with the Pogonip Group in the southern Grant Range and the southern Egan Range. Breaks in the stratigraphic column represent faulted sections.

and is composed of very fine-grained, light bluish-gray limestone wherein bedding is defined by wavy argillaceous/carbonaceous partings. The uppermost 6 m of the Lehman Formation contains 2 dm to 2 m thick beds of tan dolostone. Fossils include ostracodes and pelmatozoan fragments.

Kanosh Shale

The Kanosh Shale is approximately 46 m thick and is composed of 1 cm- to 3 dm-thick beds of laminated argillite that are interbedded with medium- to thick-bedded, fine- to coarse-grained, greenish-gray limestone and argillaceous limestone. Argillite volumetrically composes approximately 20% of the Kanosh Shale. Fossils in this unit include orthocone cephalopods, oncolites, sponges, and abundant <u>Receptaculites</u> sp.

Shingle Limestone

The Shingle Limestone is approximately 305 m thick. This unit can be qualitatively divided into upper, middle, and lower sections. The lower section consists of alternating sets of interbedded fine-grained, thin-bedded, dark grayish-green argillaceous limestone and limestone that alternate with thick-bedded, dark grayish-green, fine-grained limestone that contains minor chert and rare intraformational conglomerate. The middle section is similar to the lower section except that the rocks are dominantly medium bluish-gray, vary from fine- to medium-grained, contain minor argillite in beds up to 1 dm thick and contain rare small-scale, fossil-debris-filled channels. The upper section is thick- to very thick-bedded, coarse- to medium-grained, medium to light grayish-blue limestone. Rocks in the upper section tend to contain abundant fossil hash--in particular abundant tiny (~2 mm in diameter) pelmatozoan fragments.

Parker Spring Formation

The minimum tectonic thickness of the Parker Spring Formation is 355 m. This unit consists of thin- to medium-bedded, light grayish-green, fine-grained limestone and argillaceous limestone intercalated with thin (1 mm to 1 cm) sericite phyllite. Limestone contains chert that tends to transect bedding and that ranges in morphology from lense-shaped to approximating the shape of a rootless fold. Phyllite is a minor constituent, but this as well as intraformational conglomerate are distinctive of this unit.

Goodwin Limestone

The minimum tectonic thickness of the Goodwin Limestone is 396 m. This unit has been variably altered to dolostone, in particular, in proximity of the Heath Canyon detachment.

The Goodwin Limestone can be grossly divided into three sections. The upper section consists of thin-bedded, light gray limestone with bedding-parallel chert lenses and nodules. The middle section consists of very thick-bedded, light gray, massive limestone with minor chert. Gastropods occur near the bottom of the middle section.

The lower section ranges in thickness from 0 to 30 m and consists of thin-bedded, light gray-blue limestone (mostly hydrothermally altered to dolostone) and bedding parallel chert lenses. Near the base of the lower section, the limestone is black to dark gray and laminated, and chert appears to preferentially occur as partial to complete replacement of orange weathering laminated dolomitic layers. The lower section is correlative with rocks mapped as the Windfall Formation in the southern Grant Range (Cebull, 1967) and in the southern White Pine Range (Moores et al., 1970). However, in the Mt. Hamilton area, Humphrey (1960) mapped this same lithofacies as "member one" of the "Goodwin Formation". I have not considered it a mappable unit because of its minor and irregular extent.

The lower section tends to lie above a low-angle normal fault associated with fault "B" and constitutes a "zone of chaotic stratal disruption" (ZCSD). The ZCSD comprises breccia, disharmonic folds, and small faults. Some of the folds evolved into small faults, and in places strata merge into breccia. In the breccias and folds, carbonate is commonly ductiley deformed and quartz and graphitic carbonate are brittlely deformed. Nowhere can the folds be traced for more than 5 m and orientations of strata are both vertically and laterally chaotic: the ZCSD lacks a uniform fabric. Although the rocks record brittle and ductile deformation, the mineral grains are recrystallized. Brittlely or ductilely strained minerals are generally lacking in this zone.

I suspect that much of the deformation in the ZCSD may be syn-sedimentary, although the proximity of fault "B" necessarily raises the question of 1) whether or not the ZCSD, or at least some of the deformation, is a consequence of movement on fault "B" or 2) perhaps even whether of not some of the deformation formed in response to earlier compressional events. However, tectonic fabrics (e. g. foliation and lineation) such as those produced in east- or west-vergent deformation are lacking. Moreover, folds in the ZCSD lack a uniform orientation, a situation that might be expected as a result of syn-sedimentary slumping rather than tectonic processes. My study of these features is reconnassaince scale and it would take a more detailed study to support or refute a synsedimentary origin for the ZCSD. Nevertheless, structures in the ZCSD are morphologically similar to carbonate slump and flow deposits described by Nelson and Lindsley-Griffin (1987), and those described by Cook and Taylor (1977) in coeval (?) rocks 70 km to west in the Hot Creek Range.

Metamorphism

In the Kanosh Shale, Shingle Limestone and Parker Spring Formation, pelite layers greater than 3 cm were metamorphosed to argillite, whereas thinner bedded pelite was converted to sericite phyllite. The degree of development of metamorphic white mica appears to be enhanced in deformed regions. White mica and rare chlorite appear to be the only metamorphic minerals in carbonate.

LITTLE MEADOWS FORMATION

The Little Meadows formation in the study area is everywhere dismembered by extensional faults so that no complete section exists. Cebull (1967) informally named the Little Meadows formation and designated a type section in the southern Grant Range. The type section is 154 m thick (Cebull, 1967), well exposed and very much like the section in the northern Grant Range. Thus those who are interested in this unit should first refer to Cebull's type section. Moreover, because on a range-wide basis this unit tends to be dismembered by extensional faults, the type section may be the only place in the Grant Range where a complete section is exposed.

The tectonic thickness of the Little Meadows formation in the study area generally does not exceed 40 m and in most places is less than 30 m. The dominant lithologies are massive, light blue-gray limestone to marble and white limestone to marble with a pinkish tinge. In the eastern part of the study area, much of the Little Meadows formation has been hydrothermally altered to dolostone. No fossils were found in the northern Grant Range but Cebull (1967) indicates that fossils in the Little Meadows formation in the southern Grant Range are Late Cambrian.

SIDEHILL SPRING FORMATION

Introduction

Cebull (1967) informally named the Sidehill Spring formation for strata between the Little Meadows Formation and Pole Canyon Limestone. I have retained Cebull's (1967) nomenclature for correlative strata but have recognized three mappable units which I informally name, from stratigraphically highest to lowest, the Blue Eagle member, Grant Canyon member, and Willow Springs member. The thicknesses of these units are not known because the rocks are faulted and folded.

Blue Eagle Member

The Blue Eagle member consists of thin-bedded, fine-grained, light blue-gray limestone intercalated with resistant, orange weathering, medium gray limestone. Resistant layers are graphitic(?) and contain quartz. The Blue Eagle member contains rare intrafromational conglomerate. The best place to observe the Blue Eagle member is in Blair Canyon. The minimum thickness of this unit is 20 m.

Metamorphic minerals in the Blue Eagle member are white mica and tourmaline porphyroblasts.

Grant Canyon Member

The upper part of Grant Canyon member consists of dark gray-blue limestone with orange weathering dolomitic mottling (burrows?) interbedded with thin-bedded greenish-gray limestone, phyllite (up to at least 2 m thick) and minor intraformational limestone conglomerate. The lower part of the Grant Canyon Member consists of phyllite and alternating thin-bedded, blue-gray limestone and argillaceous limestone. This unit contains a few disharmonic folds that lack tectono-metamorphic fabric and may be syn-sedimentary in origin. The Grant Canyon member is quite fossiliferous; fossils include trilobites, linguloid brachiopods, and sponge spicules.
The minimum thickness of the Grant Canyon member is 50 m.

Phyllite contains white mica, chlorite and in some places ilmenite (?; altered to leucoxene). The limestone was not observed in thin section and thus its metamorphic mineral assemblage(s) are unknown.

Willow Springs Member

The Willow Springs member consists of light grayish-green to light bluish-green to orangish-brown, thin- to thick-, but dominantly thin-bedded fine-grained calcite marble. Bedding is defined by poikiloblastic amphibole (after phlogopite) and/or phlogopite-rich layers. This unit also contains minor zircon (in phlogopite) and metamorphic chlorite and white mica. Near low-angle normal faults, phlogopite is variably altered to chlorite. The minimum thickness of this unit is 100 m.

Certain parts of the Willow Springs member contain a strong, thick (~3 mm) phlogoptie-bearing cleavage (S_1) and it is in these rocks that two distinct planar elements, both bedding and cleavage, are evident. The differentiation between the two usually can be discerned because large micas are generally concentrated in cleavages whereas micas in bedding tend to be small and sparsely distributed.

POLE CANYON LIMESTONE

The Pole Canyon Limestone (Drewes and Palmer, 1957) consists of thin- to thick-bedded, light blue-gray and dark blue-gray, fine-grained marble with subordinant tabular, resistant, reddish-brown weathering, quartzose-micaceous layers up to 2 dm thick and thin-bedded schist or phyllite. Fossils in this unit include trace fossils, and in Grant Canyon 3 km south of the study area, oncolites.

Marble contains phlogopite, white mica, and chlorite. Phyllite or schist contains white mica, biotite, and chlorite, and the "quartzose-micaceous layers" contain biotite, white mica, and chlorite.