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Structural evolution of the Starhope Canyon-Muldoon Canyon Area, Copper Basin, Pioneer Mountains, south-central Idaho.

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STRUCTURAL EVOLUTION OF THE
STARHOPE CANYON-MULDOON CANYON AREA,
COPPER BASIN, PIONEER MOUNTAINS,
SOUTH-CENTRAL IDAHO

by

Robert C. Marshall

A Thesis
Presented to the Graduate Committee
of Lehigh University
in Candidacy for the Degree of
Master of Science
in Geology

Lehigh University

1983

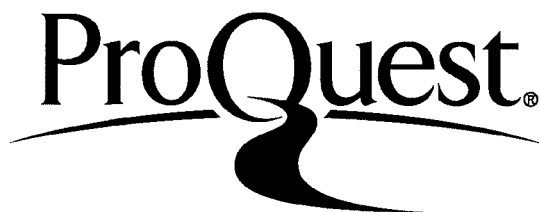
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February 14, 1983
(date)

Professor in Charge

Chairman of Department

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1. Abstract

Geologic mapping, field studies, and detailed analysis of the minor structures within the Starhope Canyon-Muldoon Canyon area in the Pioneer Mountains of south-central Idaho indicate that the Mississippian Copper Basin Formation is contained within two allochthonous plates. Rocks in the upper, Glide Mountain plate are characteristically folded by pervasive small and intermediate scale folds in contrast to the rocks of the lower, Copper Basin plate which typically are only gently folded. Both allochthonous plates and the thrust which separates them are gently folded on a regional scale. The Brockie subplate of Nilsen (1977) is correlated with the Glide Mountain plate of the Copper Basin Formation based on mapping evidence, similarities in structural style, and contrasting structural styles between the Brockie subplate and the Copper Basin plate. Therefore the Brockie subplate and the Glide Mountain plate are considered to be one and the same (the Glide Mountain plate) and the Muldoon Canyon thrust, which Nilsen (1977) maps at the base of his Brockie subplate, is correlated with the Glide Mountain thrust which underlies the Glide Mountain plate. This thrust is a thick breccia zone which can be traced throughout most of the study area, except along the east ridge of Muldoon

Canyon where it is extensively covered by surficial deposits.

Two periods of folding (F_1 and F_2) are identified; but D_1 and D_2 deformational events are not designated due to the fact that F_1 folds and F_2 folds display equivalent axial trends ($N15^\circ W-S15^\circ E$) and may have developed under a stress continuum. Small and intermediate scale folds throughout the Glide Mountain plate, directly beneath the Copper Basin plate, and within the Drummond Mine Limestone of Paull and others (1972) in the Copper Basin plate are all assigned an F_1 notation. F_1 folds developed, to a large degree, during thrusting of the Glide Mountain plate. Gentle to open regional folds which post-date thrusting are assigned an F_2 notation.

A weakly to moderately developed spaced fracture cleavage (S_1) was formed in the less competent stratigraphic units within both thrust plates of the Copper Basin Formation during the F_1 folding and thrusting period. This cleavage is best displayed within the Copper Basin plate in the Drummond Mine Limestone of Paull and others (1972). A well developed joint system, composed of three joint sets, is present in both thrust plates of the Copper Basin Formation and is post-thrusting in age. Eocene quartz monzonite intrusive

stocks, Eocene Challis Volcanics, and the Upper Cretaceous (?) to Lower Eocene (?) Smiley Creek Conglomerate all post-date emplacement of the Glide Mountain plate. Jointing is pre-intrusive in age and minor offsets along joint sets are developed in the a kinematic direction and may have been produced by the same stress system which produced F_1 and F_2 folds. Some minor tectonic adjustments within the rocks of the Starhope Canyon- Muldoon Canyon area are post-intrusive in age and are not related to the dominant regional structural fabric. The Starhope Canyon-Muldoon Canyon area may be intrusively domed, which could be responsible for minor tectonic adjustments.

2. Introduction

The primary objective of this study was to undertake a detailed analysis of the structural geology of the Starhope Canyon-Muldoon Canyon area in the Pioneer Mountains in south-central Idaho in an attempt to resolve the controversy concerning the number of allochthonous plates and stratigraphic relationships within the Mississippian Copper Basin Formation. The stratigraphy of the Mississippian Copper Basin Group (now referred to as the Copper Basin Formation) which was proposed by Paull and others (1972) and modified by Paull and Gruber (1977) has subsequently been rejected by other workers (Nilsen, 1977; Skipp and others, 1979; Dover, 1980, 1981) on the premise that thrust faults complicate the stratigraphic sequence. Paull and others (1972) considered the rocks of their Copper Basin Group to represent a normal stratigraphic sequence. Nilsen (1977) suggested that the Copper Basin Group of Paull and others (1972) actually consists of three allochthonous plates: the Scorpion subplate, Brockie subplate, and Glide Mountain plate. Alternatively, Dover (1980, 1981) proposed that (1) the Brockie subplate of Nilsen (1977) and the Glide Mountain plate of Nilsen (1977) are parts of the same allochthonous plate, and therefore (2) the Copper Basin Group of Paull and others (1972) consists of

only two allochthonous plates.

The Starhope Canyon-Muldoon Canyon area was selected for this study because of its strategic location in relation to the stratigraphic and structural problems concerning the rocks which comprise the Mississippian Copper Basin Formation. The present study area is contained within the larger study areas of Paull and others (1972), Paull and Gruber (1977), Nilsen (1977) and Dover (1980, 1981).

2.1 Geographic Setting

The Starhope Canyon-Muldoon Canyon area consists of an 85 square kilometer region at the headwaters of the East Fork of the Big Lost River in the Mackay 3NW and Mackay 3NE 7.5 minute U. S. Geological Survey topographic quadrangle maps. The area lies at the southern end of the Copper Basin, in the Pioneer Mountains of south-central Idaho; approximately 28 km. southwest of Mackay, 32 km. east of Ketchum, and 40 km. west of Arco (see Figure 2-1 and Plate A). Access to the study area can be gained by two alternative routes: (1) from Mackay via the Burma Road through the White Knob Mountains to the Copper Basin Loop Road, which enters the northern end of the study area; or (2) from either Mackay or Ketchum by exiting the Trail Creek Road at the Copper

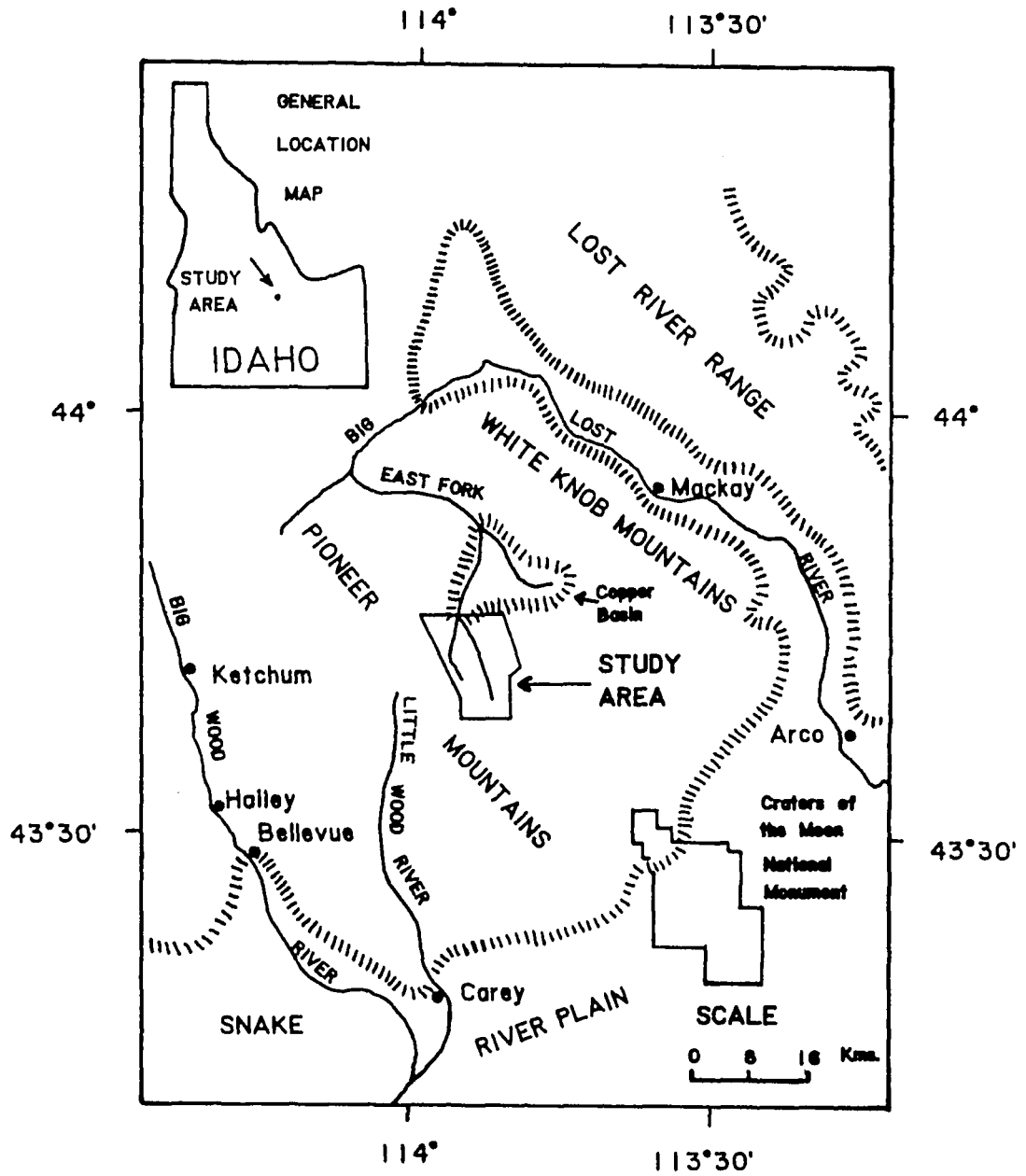


Figure 2-1: Location Map of the Starhope Canyon-Muldoon Canyon Area.

Basin Road which connects with the Copper Basin Loop Road. Unimproved Forest Service and mining roads provide access from the Copper Basin Loop Road into Muldoon and Starhope Canyons, but most of the study area is accessible only by foot.

The study area includes the full extent of both Starhope and Muldoon Canyons, from their mouths, which face Copper Basin to the north, to their headwalls approximately 8 to 10 kilometers to the south. The western border of the study area, which roughly coincides with the Blaine County-Custer County Line, is defined by the points of highest elevation on the west ridge of Starhope Canyon. The eastern border is defined by the points of highest elevation on the east ridge of Muldoon Canyon. The southern boundary is defined by the headwalls of both Starhope Canyon and Muldoon Canyon; and the northern boundary is the 43 degree 45 minute parallel, which constitutes the northern border of the Mackay 3NW and Mackay 3NE 7.5 minute quadrangles.

Elevations in the study area range from a minimum of approximately 7890 feet at the front of the ridge between Starhope Canyon and Muldoon Canyon to a maximum of 11,258 feet on the east ridge of Muldoon Canyon above Green Lake. The lower and middle slopes of the ridges are to a large extent underlain by unconsolidated Quaternary

deposits (talus, colluvium, alluvial fans, and glacial tills) and are vegetated with sagebrush, aspen, and pine. The floors of both canyons are underlain by glacial tills, glacial outwash and recent stream alluvium.

2.2 Geologic Setting

The Pioneer Mountains of Idaho, which include the study area, are located in the Central Rocky Mountain Province of Eardley (1962). They lie south and east of the Idaho batholith, west of the Lost River Range and the White Knob Mountains, and north of the Snake River Plain.

Most of the major mountain ranges in the vicinity of the study area follow a pronounced north-northwest; south-southeast regional structural trend. These ranges include the Beaverhead, Lemhi, Lost River, and Pioneer Mountains. They were formed during Miocene-Pliocene times as "Great Basin" type structures (Anderson, 1934; Baldwin, 1951; Ruppel, 1964). Both Muldoon Canyon and Starhope Canyon also lie parallel to this north-northwest; south-southeast trend.

The north-northwest; south-southeast structural trend is expressed in both (1) macroscopic structural elements, such as major fold axes and compressional and tensional faults; and (2) mesoscopic structural elements, such as minor fold axes, strike and dip of Paleozoic

strata and cleavage. This structural trend is also generally reflected in depositional patterns in Upper Paleozoic sedimentary rocks (Scholten, 1957; Thomasson, 1959; Ross, 1962a, 1962b; Churkin, 1962). Other workers (Skipp and Mamet, 1970; Schramm, 1978), however, suggest that north-south facies changes may complicate previously accepted models of northeast-southwest facies changes and north-northwest; south-southeast depositional patterns.

Rocks exposed in the study area range from Lower Mississippian to Eocene. Upper Paleozoic sedimentary rocks within the study area are restricted to the Mississippian Copper Basin Formation. A few, small, isolated outcrops of the Upper Cretaceous (?) to Lower Eocene (?) Smiley Creek Conglomerate occur within the study area in upper Muldoon Canyon. Extrusive rocks of the Eocene Challis Volcanics are locally present at high elevations. Post-orogenic quartz monzonite intrusive bodies in the form of stocks and associated rhyolite and andesite dikes of Eocene age are also present. Unconsolidated Pleistocene to Holocene surficial deposits line the valley floors and mantle some of the slopes.

The stratigraphic and structural relationships in Upper Paleozoic rocks of the Starhope Canyon-Muldoon Canyon area are obscured by the following:

1. The stratigraphic units which comprise the Copper Basin Formation are relatively thick (generally on the order of thousands of meters), lithologically similar, and contain few distinctive marker beds.
2. Contacts between the stratigraphic units are, for the most part, gradational over considerable vertical intervals (generally on the order of 60 to 160 meters). Also, complex facies changes tend to mask stratigraphic relationships.
3. Extensive cover of unconsolidated Quaternary deposits and Eocene Challis Volcanics has resulted in poor exposures of Upper Paleozoic rocks in much of the area.
4. Eocene intrusive rocks (stocks and dikes) that have been injected into Upper Paleozoic strata and structural zones of weakness (faults, joints, etc.) have obscured structural relationships.

These difficulties have hindered this study and previous studies within the Pioneer Mountains (Paul and others, 1972; Nilsen, 1977; Dover, 1980, 1981). Furthermore, scarcity of fossils has prevented precise age assignments for several of the stratigraphic units which comprise the Copper Basin Formation.

2.3 Tectonic History

Early ideas concerning the depositional and tectonic setting for the Mississippian Copper Basin Formation include:

1. The northern end of a geosynclinal trough adjacent to a source area now obscured by the

Idaho batholith (Ross, 1934a, 1934b, 1934c; Bissell, 1962).

2. A transitional zone between miogeosyncline and eugeosyncline (Kay, 1951; Scholten, 1957).
3. A rapidly subsiding trough (Muldoon trough of Thomasson, 1959) adjacent to the Antler orogenic belt (Churkin, 1962; Roberts and Thomasson, 1964; Roberts and others, 1965; Paull and Gruber, 1977).

Poole (1975) suggested, based on lithologic comparison with similar deposits in Nevada and regional map distribution of these rocks, that the Copper Basin Formation (then the Copper Basin Group) was part of the flysch deposits of the Antler foreland basin. Based on sedimentologic data Nilsen (1977, p. 275) identified the Copper Basin Formation as

"a deep sea fan and related turbidites deposited in a trough or foreland basin that developed between the Antler orogenic belt on the west and the cratonal-miogeosynclinal hinge line to the east."

According to Nilsen's model, terrigenous turbidites preserved in the Copper Basin Formation were derived from the Antler highlands to the west, while the carbonate turbidites of the Copper Basin Formation were derived from the cratonal shelf to the east.

The Late Devonian and Early Mississippian Antler orogeny, which gave rise to the deposition of the Copper

Basin Formation and other flysch deposits of the Antler foreland basin, was first recognized in central Nevada (Roberts, 1964; Smith and Ketner, 1968). The Roberts Mountain thrust has been the principal structure attributed to the development of an Antler highland in Nevada (Roberts and others, 1958). Several workers (Churkin, 1962; Roberts and Thomasson, 1964; Poole, 1974) have presumed that Late Devonian and/or Early Mississippian thrusts equivalent to the Roberts Mountain thrust may project northward into central Idaho, but Late Cenozoic volcanic rocks of the Snake River Plain Group and Early Cenozoic (?) plutonic rocks of the Idaho batholith and associated stocks and plutons prevent tracing of these thrusts (Nilsen, 1977). However, Dover (1980, p. 371) points out that

"evidence which suggests that thrusting accompanied Antler highland development in Idaho has not yet been demonstrated."

More recently, Ketner and Smith, Jr. (1982) have demonstrated that a mid-Paleozoic age for the Roberts Mountain thrust in Nevada can no longer be taken for granted and a post-Paleozoic age cannot be ruled out.

Regional mapping by many workers (Dover, 1969, 1975; Schanz and others, 1974; Hall and others, 1975; Skipp and Hall, 1975a, 1975b; Dover and others, 1976; Skipp and

Halt, 1977) confirms the fact that the central Idaho region is complicated by Mesozoic thrusting commonly attributed to the Sevier orogeny.

Skipp and others (1979) present two possible plate tectonic models for the plate boundary tectonic events along the western margin of a proto-North American continent which gave rise to the Antler orogeny and the deposition of the Copper Basin Formation. Both models involve (1) an arc-continent collision on the western margin of the proto-North American continent culminating in (2) major changes in structural settings for deposition of Carboniferous rocks in latest Devonian to earliest Mississippian time.

Model I, modified from Poole (1974a, 1974b) and Poole and Sandberg is based on back-arc thrusting during a period (Late Devonian to Early Mississippian) of increased plate motion along an eastward dipping subduction zone. As a result of back arc thrusting, obducted inner arc basin sediments formed the Antler orogenic belt which served as a source area for the flysch of the Copper Basin Group of Pauli and others (1972), as well as Mississippian flysch in Nevada south of the Snake River Plain. In this model the inner arc basin continued to close through the Permian, culminating in collision and welding of the island arc (Klamath-North

Sierran arc terraine of Poole and Sandberg, 1977) to the proto-North American continent in the Triassic.

Model II, from Dickinson (1977), is based on eastward thrusting of an Antler highland onto the continental margin along a westward dipping subduction zone during latest Devonian to earliest Mississippian time. According to this model mid-Carboniferous rifting broke up the Antler orogeny, which interrupted maturation of the Antler flysch (including the Copper Basin Formation) to molasse. The rifting produced various fault-bounded basins and highlands in Pennsylvanian and Permian time. In Permian time the volcanic highland collided with the proto-North American continent and was welded to it by mid-Triassic time.

3. Previous Work

3.1 Upper Paleozoic Stratigraphy and Structure

The earliest geological investigations into Upper Paleozoic stratigraphy in south-central Idaho were conducted by Lindgren (1900), Umpleby (1917), and Umpleby and others (1930). Later modifications concerning Upper Paleozoic stratigraphic nomenclature in south-central Idaho were suggested by Bostwick (1955), Sandberg (1975), and Sandberg and others (1975). Ross (1934a, 1934b, 1937, 1947, 1960, 1961, 1962a, 1962b, 1962c, 1963, 1969a, 1969b) and Nelson and Ross (1968, 1969a, 1969b) conducted extensive mapping, stratigraphic, and structural studies in the Pioneer, White Knob, and Lost River Ranges.

Thomasson (1959) described complex stratigraphic relations in the Upper Paleozoic section of the Pioneer Mountains and informally proposed the name Muldoon Formation for these rocks. Ross (1960, 1962a, 1962c) suggested that the Upper Paleozoic rocks of the central Pioneer Mountains merge and interfinger with the White Knob Limestone on the east and with the Milligen and Wood River Formations to the west. Ross (1962a) proposed the name Copper Basin Formation for this complex of interfingering Upper Paleozoic rocks. He assigned an Early Mississippian to Early Permian age to the Copper

Basin Formation and estimated its thickness to lie in the range of 3050 to 4575 meters. Dover (1966, 1969) undertook extensive geologic mapping in the Pioneer Mountains and suggested that the abrupt east-west Upper Paleozoic facies changes were partially the result of thrust faults of great magnitude. He reserved conclusions concerning stratigraphic relations until structural relationships were better understood.

Pauli and others (1972, p. 1372) noted that

"Lithology and thickness comparisons (of the Upper Paleozoic rocks of the Pioneer Mountains) with lithologic sequences in adjacent areas are difficult because faults have telescoped the Paleozoic stratigraphic succession of the Wood River Valley on the west into juxtaposition with the Paleozoic rocks of the central Pioneer Mountains which, in turn, apparently have been faulted into juxtaposition with the Paleozoic rocks of the White Knob Mountains on the east. The age, geometry, and displacement of these faults are not well known, and it may be that none of the Upper Paleozoic rocks in the Pioneer Mountains are autochthonous."

Skipp and Hall (1975a, 1975b) considered Upper Paleozoic rocks of the Pioneer Mountains to be contained within a gently folded and faulted allochthon, which they named the Copper Basin allochthon. The Copper Basin allochthon is considered to be about 19 kilometers wide and to have been thrust over older Paleozoic strata to the west and Mississippian to upper Permian strata to the east (Skipp

and Hall, 1975a, 1975b).

Paul and others (1972) formally proposed raising Ross's (1962a) Copper Basin Formation to group status and designated six formations within the Copper Basin Group. The Copper Basin Group of Paul and others (1972) consists of the following stratigraphic units in ascending stratigraphic order: the Milligen Formation, Drummond Mine Limestone, Scorpion Mountain Formation, Muldoon Canyon Formation including the Green Lake Limestone Member, Brockie Lake Conglomerate, and Iron Bog Creek Formation. Paul and others (1972) stated that the Copper Basin Group consists of a continuous sequence of 5490 meters of Lower (?) Mississippian to Middle (?) Pennsylvanian predominantly terrigenous rocks. Terminology and age assignments within the Copper Basin Group of Paul and others (1972) were later modified by Paul and Gruber (1977). The lowest formation of the Copper Basin Group of Paul and others (1972) was originally named the Milligen Formation and correlated with the Milligen Formation of the type area in the Wood River Valley. When Sandberg and others (1975) later described the Milligen Formation at the type area as a relatively deep-water deposit of Devonian age, Paul and Gruber (1977) renamed the lowest formation of the Copper Basin Group the Little Copper Formation. Skipp and Hall

(1975b) later restricted the age of the Copper Basin Group of Pauli and others (1972) to the Mississippian. However, Nilsen (1977, p. 279) noted that

"no fossils indicating that any part of the type sections of the Copper Basin Group of Pauli and others (1972) is unquestionably younger than Early Mississippian in age have yet been found."

The Copper Basin Group of Pauli and others (1972) was mapped in the Pioneer Mountains by Wolbrink (1970), Bollman (1971), Lukowicz (1971), Grover (1971), Volkmann (1971), Pruitt (1971), Erwin (1972), Rothwell (1973), Larson (1974), Yokley (1974), and Schramm (1978), students of R. A. Pauli at the University of Wisconsin at Milwaukee, and Spoelhof (1972), a student at the Colorado School of Mines. The stratigraphy of the Copper Basin Group as described by Pauli and others (1972) was considered to be essentially correct by all of Pauli's students. However, slight modifications were proposed by Schramm (1978), who suggested that the Brockie Lake Conglomerate and Scorpion Mountain Formation of Pauli and others (1972) merged laterally towards the south in the Pioneer Mountains.

Alternatively, Nilsen (1977, pp. 279-280) noted that

"At least one major fault within the stratigraphic sequence of the Copper Basin Group of Pauli and others (1972) ... has probably

repeated the sequence (and) as a result, the proposed thickness of Pauli and others (1972) for their Copper Basin Group ... is about twice the true thickness."

Nilsen (1977) cited primarily stratigraphic and sedimentologic evidence combined with some structural evidence to support the hypothesis that the Copper Basin Group of Pauli and others (1972) is actually composed of (1) two structural plates (or subplates) which are both age-equivalent and facies-equivalent and repeat the stratigraphic sequence due to the fact that one plate (Brockie subplate) structurally overlies the other plate (Scorpion subplate), and (2) a third structural plate (Glide Mountain plate) containing rocks which lithologically resemble some units within the Scorpion and Brockie subplates and may be a partial facies equivalent of rocks in both of these (see Table 3-1 and Figure 3-1).

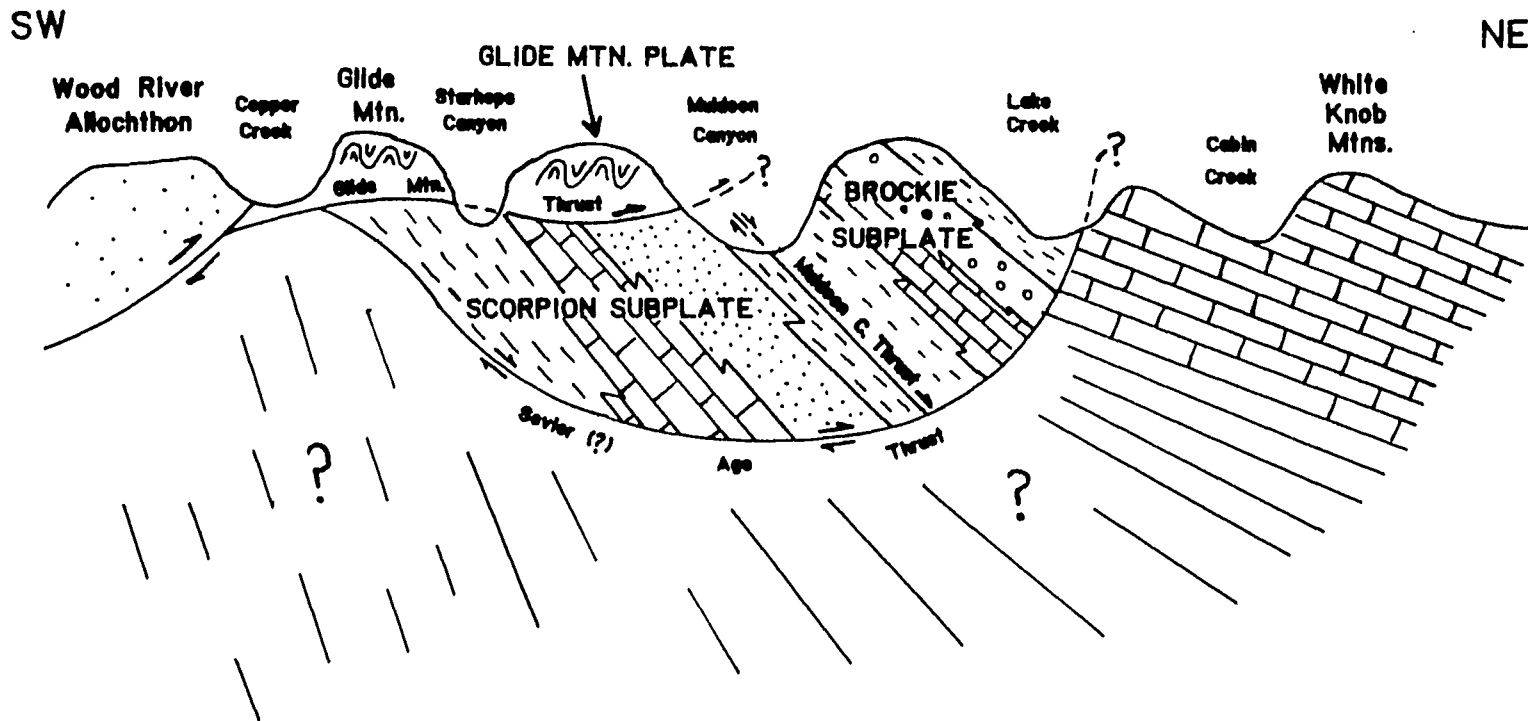
According to Nilsen's model, the Brockie subplate, which contains the more proximal facies of the Antler foreland basin, is thrust atop the Scorpion subplate, which contains the more distal deposits. The (lower) Scorpion subplate of Nilsen (1977) is composed, in ascending stratigraphic order, of the Little Copper Formation, Drummond Mine Limestone, Scorpion Mountain Formation, and lower part of the Muldoon Canyon Formation

	Pauli and others (1972) Pauli and Gruber (1977)	Nilsen (1977)	Dover (1980, 1981)
West Ridge of Muldoon Canyon- East Ridge of Starhope Canyon	Muldoon Canyon Formation in upper thrust plate	Glide Mountain Plate	Glide Mountain Plate
East Ridge of Muldoon Canyon	Iron Bog Creek Fm. Brockle Lake Congl. Green Lake Lmst. Member of Muldoon Canyon Fm. (upper part)	Brockle Subplate	
Starhope and Muldoon Canyons	Muldoon Canyon Fm. (lower part) Scorpion Mountain Fm. Drummond Mine Lmst. Little Copper Fm.	Scorpion Subplate	Copper Basin Plate
	Copper Basin Group	Copper Basin Plate	Copper Basin Formation

Table 3-1: Correlation of stratigraphic and structural nomenclature for the thrust plates of the Copper Basin Formation.

	Paul and others (1972) Paul and Gruber (1977)	Nilsen (1977)	Dover (1980, 1981)
West Ridge of Muldoon Canyon- East Ridge of Starhope Canyon	Muldoon Canyon Formation in upper thrust plate →	Glide Mountain Plate →	Glide →
East Ridge of Muldoon Canyon	Copper Basin Group →	Copper Basin Plate →	Copper Basin Formation →
Starhope and Muldoon Canyons			
	Iron Bog Creek Fm. Brockie Lake Congl. Green Lake Lmt. Member of Muldoon Canyon Fm. (upper part) ----- Muldoon Canyon Fm. (lower part) Scorpion Mountain Fm. Drummond Mine Lmt. Little Copper Fm. →	Brockie Subplate → Scorpion Subplate →	Mountain Plate → Copper Basin Plate →

Table 3-1: Correlation of stratigraphic and structural nomenclature for the thrust plates of the Copper Basin Formation.



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Figure 3-1:
 Schematic cross-section of Nilsen (1977)
 through the Starhope Canyon-Muldoon Canyon Area.
 The Mississippian Copper Basin Formation
 is contained within the Glide Mountain plate,
 Scorpion subplate, and Brockie subplate.

of Paull and others (1972) and Paull and Gruber (1977). The (upper) Brockie subplate of Nilsen (1977) is composed, in ascending stratigraphic order, of the upper part of the Muldoon Canyon Formation containing the Green Lake Limestone Member, Brockie Lake Conglomerate, and Iron Bog Creek Formation of Paull and others (1972). According to Nilsen (1977) the Glide Mountain plate, which is thrust atop both the Scorpion and Brockie subplates, is composed of either (1) a more western, largely terrigenous and plant-rich facies of the Copper Basin Group of Paull and others (1972), or (2) the lower part of the Brockie subplate which became detached (during thrusting of the Brockie subplate) and was then later thrust eastward.

Wolbrink (1970) and Schramm (1978) mapped the strata of the Glide Mountain Plate of Nilsen (1977) as part of the Muldoon Canyon Formation of Paull and others (1972) which they showed was thrust atop the Scorpion Mountain Formation of Paull and others (1972). Wolbrink (1972) and Schramm (1978) recognized this thrust plate on the ridge between Starhope Canyon and Muldoon Canyon. Nilsen (1977), Dover and others (1976), and Dover (1981) also mapped this thrust (the Glide Mountain thrust) on the east ridge of Starhope Canyon, including Glide Mountain.

Nilsen (1977) noted that the strata of the Glide

Mountain plate (referred to as the Glide Mountain sequence) is tightly folded and cleaved in contrast to the strata of the underlying Scorpion and Brockie subplates, which are gently folded. Based on this difference in structural character between the Glide Mountain plate and the Scorpion and Brockie subplates (together), and the lithologic similarities between the Scorpion and Brockie subplates, Nilsen (1977) grouped the Scorpion and Brockie subplates together into a single structural unit, the Copper Basin plate (see Table 3-1 and Figure 3-1). Nilsen (1977) suggested that the Brockie subplate was thrust eastward atop the Scorpion subplate during closing of the Antler foreland basin in Early to Late Mississippian time. This thrust is referred to as the Muldoon Canyon thrust by Nilsen (1977). According to Nilsen's model, the Scorpion and Brockie subplates were later tectonically transported as a unit, the Copper Basin plate, during the post-Paleozoic Sevier orogeny. Nilsen (1977) inferred that the Glide Mountain plate may have been "later" thrust atop the Copper Basin plate along a thrust referred to as the Glide Mountain thrust. Nilsen (1977) does not offer any structural evidence to support his dating of the thrusts.

More recently, Dover (1980) has demonstrated that:

1. Post-Antler, mainly Mesozoic thrusting dominates the tectonic framework of the Antler orogen in Idaho.
2. The severity of the Mesozoic tectonic overprint is such that it seriously complicates the identification of Antler structures, interpretations of Antler orogenesis, and paleogeographic reconstruction.
3. All datable thrusts are post-Permian (age of the Wood River Formation, youngest rock stratigraphic unit involved in thrusting) to pre-Eocene (age of Challis Volcanics and post-orogenic intrusives) in age in Idaho.

Dover (1980, 1981) suggested that the Brockie subplate of Nilsen (1977) is, in fact, part of the (upper) Glide Mountain plate, and therefore the Muldoon Canyon thrust of Nilsen (1977) is a continuation of the Glide Mountain thrust throughout the rest of the Pioneer Mountains. The major evidence which Dover cites in support of this hypothesis is the recent discovery of limestone equivalent to the Green Lake Limestone Member in the Glide Mountain plate at Big Rocky Canyon (located north of the present study area) and "other lithologic similarities and mapping evidence." Therefore, the Glide Mountain plate as defined by Dover (1980, 1981) consists of both the Glide Mountain plate of Nilsen (1977) and the Brockie subplate of Nilsen, 1977 (see Table 3-1). Nilsen (1981, personal communication) still maintains, however, that the Brockie subplate of Nilsen (1977) is part of the

(lower) Copper Basin plate, and the Glide Mountain plate of Nilsen (1977) is a separate, higher structural plate.

3.2 Mesozoic and Cenozoic Stratigraphy

3.2.1 Upper Cretaceous (?) to Lower Eocene (?) Smiley Creek Conglomerate

Although students of R. A. Paull at the University of Wisconsin, Milwaukee (Wolbrink, 1972; Schramm, 1978) have mapped this conglomerate extensively in the Pioneer Mountains, a type section has never been formally proposed and described (Paull, 1981, personal communication). A Late Cretaceous (?) to Early Eocene (?) age is suggested for the Smiley Creek Conglomerate based on limited botanical evidence (Paull, 1974; Dover, 1981). This unit has been mapped beneath the Challis volcanics by several workers including Nelson and Ross (1969) and Dover (1981). This fact and the absence of Challis clasts within the Smiley Creek Conglomerate identify the unit as being older, though perhaps only slightly so, than the Challis volcanics. Paull (1974) considers this unit to be a post-orogenic deposit. Unfortunately the tentative age of this unit prevents precise dating of the latest orogeny in south-central Idaho.

3.2.2 Eocene Challis Volcanics

Ross (1961, p. C178) defined the Challis volcanics as:

"dominantly volcanic strata of early Tertiary age within the part of central Idaho north of the Snake River Plain and south of the westward-flowing segment of the Salmon River (near latitude 45 degrees 30 minutes."

Ross (1961) reported the age of the Challis Volcanics as Eocene to Early Miocene (?). Siems and Jones (1977), however, concluded that the Challis Volcanics were extruded over most of central Idaho in Middle Eocene time.

Ross (1937) distinguished several members within the Challis Volcanics, which are, in ascending stratigraphic order: the latite-andesite Member, basalt and related flows, the Germer Tuffaceous Member, and the Yankee Fork Rhyolite. The most extensive member within the Challis Volcanics is the latite-andesite member, to which all of the rocks in the present study area are assigned (Anderson and Wagner, 1946; Ross, 1962b). Armstrong (1975) reported a K-Ar age of 42.0 m. y. with a possible error of 1.3 m. y. on Challis Volcanics from Smiley Creek, approximately 5 kilometers east of the present study area.

3.2.3 Eocene Intrusive Rocks

Intrusive quartz monzonites cropping out in Muldoon Canyon are lithologically identical and genetically related to the larger stock in Lake Creek Canyon just east of the present study area (Wolbrink, 1970; Dover, 1981). Dover (1981) reported that this stock intrudes both the Glide Mountain plate and Copper Basin plate of the Copper Basin Formation.

The Lake Creek stock lies along an elongate gravity low that trends northwest along the crest of the Pioneer Mountains into the Boulder Mountains where Tschanz and others (1974) have mapped similar stocks and plutons (Dover, 1981). According to Dover (1981) these plutons may connect at shallow depth with a large northwest-trending batholith or zone of batholiths. These quartz monzonite intrusive bodies postdate major thrust faulting and are responsible for local intrusive doming in the Pioneer and Boulder Mountains (Dover, 1981).

Armstrong (1975) reported a K-Ar date of 47.7 m. y. with a possible error of 1.4 m. y. for chloritized biotite from the Lake Creek stock; while Stern and others (in Armstrong, 1975) reported a Pb-alpha age of 50 m. y. with a possible error of 10 m. y. on zircon from the Lake Creek stock. Dover (1981) cites field evidence to suggest that these quartz monzonite intrusive bodies are

younger than (though perhaps only slightly so) or coeval with the Challis Volcanics. Wolbrink (1972) reported xenoliths of Challis Volcanics included within a quartz monzonite intrusion (although he does not mention the location of this occurrence).

4. Purpose of Study

The major goal of this study was to conduct a detailed study of the structural geology (especially the minor structures) of the thrust plates of the Copper Basin Formation in the Starhope Canyon-Muldoon Canyon area of the Pioneer Mountains. A study of this nature, comparing and contrasting the minor structures in the various thrust plates of the Copper Basin Formation, had not been done in this area prior to this study.

The focus of the study was three-fold:

1. To map the Starhope Canyon-Muldoon Canyon area.
2. To determine a chronology of deformation based on relative age relationships between the various structures.
3. To determine whether the Copper Basin plate and Glide Mountain plate of the Copper Basin Formation possess any characteristic structural styles and/or deformation fabrics which would be useful in determining whether the Brockie subplate of Nilsen (1977) is actually part of the lower, Copper Basin plate (Nilsen, 1977) or part of the upper, Glide Mountain plate (Dover, 1980, 1981).

5. Method of Study

Field work for the present study was conducted during July and August of 1981 and July of 1982. Mapping was performed on aerial photographs at the scale of 1:15,600 and 7.5 minute U. S. Geological Survey advance print topographic maps. Unconsolidated Quaternary deposits were mapped in the laboratory on aerial photographs. For production of the final map (Plate A), data was transferred from air photos to 7.5 minute U. S. Geological Survey advance print orthophoto maps and then to 7.5 minute U. S. Geological Survey advance print topographic maps. The purpose of including surficial material (unconsolidated Quaternary deposits) and dikes on Plate A is to document areas where the relationships of thrusts and contacts between stratigraphic units are obscured. Geologic cross-sections (Plate B) have been derived from the geologic-topographic map to illustrate stratigraphic and structural relationships in the Starhope Canyon-Muldoon Canyon area.

Oriented hand specimens were obtained at various localities within the study area for petrofabric analysis of quartz optic axes (0001). This procedure was performed using a Leitz 4-axis universal stage following the techniques of Turner and Weiss (1963).

6. Terminology

6.1 Stratigraphic Nomenclature

The stratigraphy of the Copper Basin Group as proposed by Paul and others (1972) and modified by Paul and Gruber (1977) has recently been rejected by other workers (Nilsen, 1977; Skipp and others, 1979; Dover, 1980, 1981). Nilsen (1977) proposed reducing the rank of the Copper Basin Group to formation status, and other workers have subsequently adopted this terminology.

The Copper Basin Group as modified by Paul and Gruber (1977) consists of the following formations in ascending stratigraphic order: Little Copper Formation, Drummond Mine Limestone, Scorpion Mountain Formation, Muldoon Canyon Formation including the Green Lake Limestone Member, Brockie Lake Conglomerate, and Iron Bog Creek Formation. Most recently, Dover (1981), who divides the Copper Basin Group of Paul and others (1972) into two allochthonous plates, has suggested that:

1. The Little Copper Formation of Paul and Gruber (1977) and the Drummond Mine Limestone of Paul and others (1972) should be reduced to member status (within the Copper Basin plate of the Copper Basin Formation).
2. The Scorpion Mountain Formation and the lower part of the Muldoon Canyon Formation of Paul and others (1972) should be referred to as the upper clastic unit of the Copper Basin plate (of the Copper Basin Formation).

3. The upper part of the Muldoon Canyon Formation, Brockie Lake Conglomerate, and Iron Bog Creek Formation, all of Pauli and others (1972), should be discarded as formation and/or member titles in favor of the term Glide Mountain plate (of the Copper Basin Formation).
4. The Green Lake Limestone Member remain as the only member within the Glide Mountain plate (of the Copper Basin Formation).

Despite these recent developments in stratigraphic nomenclature, the stratigraphy of the Copper Basin Group as modified by Pauli and Gruber (1977) will be used extensively in this manuscript as it provides a useful framework for three important aspects of the present study: (1) mapping, (2) discussion of the structural geology of the Starhope Canyon-Muldoon Canyon area, and (3) evaluation of the structural models of Nilsen (1977) and Dover (1980, 1981). The stratigraphic units within Pauli's Copper Basin Group are much thinner than the allochthonous plates within which they are now contained. Thus, by employing the stratigraphy of the Copper Basin Group, a more precise description of both the macroscopic and mesoscopic structural geology can be provided. However, the thrust plate nomenclature of Nilsen (1977) and Dover (1980, 1981) must also be used when evaluating their structural models. The stratigraphic and thrust plate nomenclature which is employed within the course of

this study is summarized on Table 3-1. When a given worker's stratigraphic or thrust plate nomenclature is used in this report, it will be credited to that worker (e.g. Glide Mountain plate of Dover, 1980, 1981; Glide Mountain plate of Nilsen, 1977) so as to avoid misunderstanding on the part of the reader.

6.2 Other Terminology

The term argillite, as used in this study and other related studies, refers to a siltstone, mudstone, or claystone which has undergone a slightly higher degree of induration than shale without secondary cleavage development (Twenhofel, 1937).

The terminology which is used to describe small and intermediate scale folds is based on the twelve properties of mesoscopic folds of Hansen (1971). Fold hinges are described as either sharp (lacking any appreciable curvature) or broad. Limbs of folds are described as either straight or broadly curved. The ratio of short-limb height to width, which generally ranges from 0.1 to 5.0, expresses quantitatively the relative amount of overfolding within a regular wave train of folds. The greater H/W values correspond to greater degrees of overfolding. An H/W value of 0.1 describes an open fold. An H/W value greater than 1.0

describes a fold which is nearly isoclinal. The ratio of depth to width of a fold, which generally ranges from 1 to 15, expresses the elongation of the fold in profile. A D/W value of 1 describes an intrafolial fold, or an isolated fold that is contained within relatively unfolded strata. Greater D/W values correspond to folds with axial surfaces which extend through greater thicknesses of strata. The reader is referred to Hansen (1971) for a more complete discussion of these fold properties.

7. Stratigraphy

The information included below is largely based on the descriptive work of others (Paul and others, 1972; Nilsen, 1977; Paul and Gruber, 1977; Dover, 1980, 1981) and is used as a basis for approaching the structural and stratigraphic problem with which this study is concerned. The (previously discussed) thrust plate model of Dover (1980, 1981) is largely based on a revision of the stratigraphic, structural, and sedimentological concepts of Nilsen (1977), so the stratigraphy of the various thrust plates of Nilsen (1977) must be reviewed. However, because the structural and stratigraphic model of Nilsen (1977) is based on a reinterpretation of the stratigraphy of the Copper Basin Group of Paul and others (1972) and Paul and Gruber (1977), the stratigraphic nomenclature of the Copper Basin Group as modified by Paul and Gruber (1977) must also be employed.

7.1 Glide Mountain Plate of Nilsen (1977)

The strata of the Glide Mountain plate of Nilsen (1977), which he refers to as the Glide Mountain sequence, consists predominantly of argillite and interbedded quartzitic sandstone turbidites. This strata is generally plant-rich and contains a high proportion of

interbedded mudstones.

The Glide Mountain sequence is characteristically tightly folded. Dover (1980) estimates a tectonic shortening of 50 percent within the Glide Mountain plate. Because of the intense folding of this sequence, its thickness has not been measured or estimated.

Dover (1980) has identified a carbonate-rich zone within the Glide Mountain plate in Big Rocky Canyon north of the Copper Basin in the Pioneer Mountains. This zone is approximately 75 meters thick and consists of interlayered beds of argillite, silty limestone, and bioclastic limestone. The bioclastic limestone yields the same Upper Kinderhookian conodont fauna as the Drummond Mine Limestone of Paul and others, 1972 (Dover, 1980).

The Glide Mountain plate of Nilsen (1977) was mapped within the course of the present study on the east and west ridges of Starhope Canyon and on the west ridge of Muldoon Canyon, where it is in thrust contact with the Scorpion subplate of Nilsen (1977), or Copper Basin plate of Dover (1980, 1981).

7.2 Scorpion Subplate of Nilsen (1977)

The Scorpion subplate of Nilsen (1977) contains the following stratigraphic units of Pauli and others (1972) and Pauli and Gruber (1977) in ascending stratigraphic order: the Little Copper Formation, Drummond Mine Limestone, Scorpion Mountain Formation, and the lower part of the Muldoon Canyon Formation. According to Nilsen's model these stratigraphic units in the Scorpion subplate are more distal (relative to the Antler highland) equivalents of stratigraphic units contained within the Brockie subplate of Nilsen (1977).

The Little Copper Formation of Pauli and Gruber (1977) is the lowermost stratigraphic unit in the Copper Basin Group as modified by Pauli and Gruber (1977) and is also the lowermost unit in the Scorpion subplate of Nilsen (1977) or Copper Basin plate of Dover, 1980, 1981 (see Figure 3-1 and Table 3-1). The true stratigraphic base of the Little Copper Formation is not exposed in the Pioneer Mountains, although at several localities outside the present study area the unit is reported to be in thrust contact with Devonian rocks (Skipp and Sandberg, 1975; Pauli and Gruber, 1977; Schramm, 1978). The top of the unit is gradational into the overlying Drummond Mine Limestone of Pauli and others (1972).

At its type section the Little Copper Formation of

Paul and Gruber (1977) consists of 1,120+ meters of thin- to medium-bedded, blocky-weathering, dark-gray argillite with some interbeds of dark quartzite, granule-size conglomerate and sparse limestone lenses. The unit thins in all directions away from its type section (Paul and Gruber, 1977). Nilsen (1977, p. 286) identified some randomly distributed thinly bedded, carbonaceous, fine-grained facies D or, less commonly, facies C terrigenous sandstone and siltstone turbidites within the unit, but does not distinguish them as marker beds at a particular stratigraphic horizon. According to Nilsen's paleogeographic and sedimentologic model the Little Copper Formation of Paul and Gruber (1977) is the more distal equivalent (relative to the Antler highland) of the upper part of the Muldoon Canyon Formation of Paul and others (1972).

Fossils are scarce within the Little Copper Formation, and the unit has not been precisely dated. Stratigraphic relations indicate that it is older, though perhaps only slightly so, than the Drummond Mine Limestone of Paul and others (1972), which is Kinderhookian in age. Dover (1981) believes an Early Mississippian age is probable.

In the present study the Little Copper Formation of Paul and Gruber (1977) was mapped on the west ridge of

Starhope Canyon, where it stratigraphically underlies the Drummond Mine Limestone of Paull and others (1972). In this general area the Little Copper Formation is also found to be in thrust contact with the overlying Glide Mountain plate of Nilsen (1977). Due to lithologic similarities between the Little Copper Formation of Paull and Gruber (1977) and the rocks of the Glide Mountain plate of Nilsen (1977), these two units are difficult to distinguish in this area, until the thrust fault which separates them has been recognized.

The Drummond Mine Limestone of Paull and others (1972) is contained within the Scorpion subplate of Nilsen (1977), also referred to as the Copper Basin plate by Dover (1980, 1981). The Drummond Mine Limestone of Paull and others (1972) conformably overlies the Little Copper Formation of Paull and Gruber (1977). The contact between these is gradational through an interval of approximately 60 meters. The upper contact of the Drummond Mine Limestone of Paull and others (1972) is gradational with the overlying Scorpion Mountain Formation of Paull and others (1972) through an interval of 92 to 153 meters.

At its type section the Drummond Mine Limestone is 808 meters thick and consists predominantly of limestone with interbedded argillite, siltstone, sandstone,

conglomerate, and chert. The lower 305 meters at the type section is mostly medium- to thick-bedded micritic limestone, with a few thin beds of dark-gray argillite, some chert, and small amounts of granule conglomerate. The upper 488 meters is very thin- to medium-bedded micritic limestone with interbedded very thin- to thin-bedded argillite, calcareous siltstone, and fine-grained sandstone (Paull et al., 1972). According to Nilsen (1977) the Drummond Mine Limestone of Paull and others (1972) generally consists of facies C and D limestone turbidites with interbedded noncalcareous to slightly calcareous argillite. Nilsen (1977) thought that the limestone turbidites were derived primarily from eastern source areas and deposited in a basin plain environment, in contrast to the clastic units within the Copper Basin Formation, which were derived from the Antler highland to the west. According to Nilsen's paleogeographic and sedimentologic model the Drummond Mine Limestone of Paull and others (1972) is stratigraphically equivalent to the Green Lake Limestone Member of the upper part of the Muldoon Canyon Formation. Conodont faunas indicate a Kinderhookian age for the Drummond Mine Limestone (Sandberg, 1975).

In the present study the Drummond Mine Limestone was mapped on the east and west ridges of Starhope Canyon.

Within the study area the base of the Drummond Mine Limestone is poorly exposed on the west ridge of Starhope Canyon where it lies in gradational stratigraphic contact with the underlying Little Copper Formation of Paul and Gruber (1977). On the east ridge of Starhope Canyon and at the head of the canyon, the upper (gradational) contact of the Drummond Mine Limestone of Paul and others (1972) with the overlying Scorpion Mountain Formation of Paul and others (1972) is exposed. At several localities within the present study area, notably Starhope Gulch, the Drummond Mine Limestone of Paul and others (1972) is in thrust contact with the overlying Glide Mountain plate of Nilsen (1977).

The Scorpion Mountain Formation of Paul and others (1972) is also contained within the Scorpion subplate of Nilsen (1977) or Copper Basin plate of Dover (1980, 1981). The Scorpion Mountain Formation conformably overlies the Drummond Mine Limestone of Paul and others (1972) in gradational stratigraphic contact. The top of the unit is gradational with the overlying lower part of the Muldoon Canyon Formation of Paul and others (1972). At its type section, the Scorpion Mountain Formation of Paul and others (1972) consists of 1,106 meters of medium gray to medium dark-gray chert-quartzite-argillite conglomerate, interbedded with similarly colored

quartzite. Granule and pebble conglomerate generally predominate, but cobble conglomerate is also common. Conglomerate clasts are contained within a quartzite matrix (Paull et al., 1972).

The lower 92 to 153 meters of the unit at its type section, which is in close proximity to the present study area, is interbedded very thin- to thin-bedded quartzite and argillite, with some sandy limestone. Bedding in the remainder of the unit is thick to very thick. Graded conglomeratic beds in which conglomerate grades upward into quartzite are common. Graded conglomeratic beds are commonly interbedded with non-graded beds (Paull et al., 1972).

Nilsen (1977) identified the Scorpion Mountain Formation of Paull and others (1972) as outer-fan deposits consisting mostly of quartzitic sandstone turbidites (facies A, B, C and D). According to Nilsen's paleogeographic and sedimentologic model the Scorpion Mountain Formation of Paull and others (1972) is the more distal (relative to the Antler highland) equivalent of the Brockie Lake Conglomerate of Paull and others (1972).

The Scorpion Mountain Formation of Paull and others (1972) has not been dated. Based on stratigraphic relations it is younger than the Kinderhookian Drummond Mine Limestone. Dover (1981) favored a late

Mississippian age for the Scorpion Mountain Formation.

The Scorpion Mountain Formation of Paull and others (1972) was mapped extensively throughout the study area on the east ridge and headwall of Starhope Canyon and on the east and west ridges of Muldoon Canyon. At the head of Starhope Canyon, the lower contact of the Scorpion Mountain Formation with the underlying Drummond Mine Limestone of Paull and others (1972) is well exposed. The upper contact of the Scorpion Mountain Formation with the overlying lower part of the Muldoon Canyon Formation of Paull and others (1972) is exposed on the unnamed mountain in upper Muldoon Canyon (sec. 19, T.4N., R.22E.). On the east ridge of Starhope Canyon and the west ridge of Muldoon Canyon, the Scorpion Mountain Formation of Paull and others (1972) is in thrust contact with the overlying Glide Mountain plate of Nilsen (1977). A laterally continuous section of overturned beds in the Scorpion Mountain Formation of Paull and others (1972) is found at the northern end of Muldoon Canyon on the lowest slopes of the east ridge.

Paull and others (1972) measured and described the type section for the Muldoon Canyon Formation at two locations, one on either side of upper Muldoon Canyon. The basal 337 meters of this unit was measured and described on the unnamed mountain in upper Muldoon Canyon

(sec. 19, T.4N., R.22.E.). The remaining (upper) 793 meters of this unit was measured on the east ridge of Muldoon Canyon. Nilsen (1977) refers to the basal and upper type sections as the lower and upper parts of the Muldoon Canyon Formation of Paull and others (1972), respectively. The lower part of the Muldoon Canyon Formation of Paull and others (1972) is the uppermost unit in the Scorpion subplate of Nilsen (1977) or Copper Basin plate of Dover (1980, 1981). The upper part of the Muldoon Canyon Formation of Paull and others (1972) is the basal unit in the Brockie subplate of Nilsen (1977). Nilsen (1977) considers the lower and upper parts of the Muldoon Canyon Formation of Paull and others (1972) to be separated by the Muldoon Canyon thrust fault.

Paull and others (1972) have shown the lower part of the Muldoon Canyon Formation to be in gradational stratigraphic contact with the underlying Scorpion Mountain Formation. The lower part of their Muldoon Canyon Formation consists of dark-gray to medium light-gray silty, locally shaly-bedded argillite interbedded with subordinate dark gray quartzite and quartzite-chert-argillite granule to pebble conglomerate and micaceous mudstone (Paull et al., 1972). Nilsen (1977) identified this unit as consisting primarily of interbedded quartzitic sandstone turbidites and argillite

with subordinate granule to pebble conglomerate.

Nilsen (1977) made the following observations concerning the sedimentology of the lower part of the Muldoon Canyon Formation of Paull and others (1972):

1. Facies C and D turbidites are most characteristic of the sequence.
2. The unit was deposited primarily in basin-plain environments, although some lobe-fringe deposits are present in the lowest part of the unit.
3. Paleocurrent data indicates sediment transport to the east.

According to Nilsen's model the lower part of the Muldoon Canyon Formation of Paull and others (1972) is the more distal (relative to the Antler highland) equivalent of their Iron Bog Creek Formation.

The lower part of the Muldoon Canyon Formation of Paull and others (1972) has not been dated. Based on stratigraphic relations it is younger than the Kinderhookian Drummond Mine Limestone and the overlying Scorpion Mountain Formation of Paull and others (1972).

In the present study the lower part of the Muldoon Canyon Formation of Paull and others (1972) was mapped on the west ridge of Muldoon Canyon and on the unnamed mountain in upper Muldoon Canyon (sec. 19, T.4N., R.22E.). At the latter site the base of the unit is

exposed in gradational stratigraphic contact with the underlying Scorpion Mountain Formation. On the west ridge of Muldoon Canyon, toward the southern end of the canyon, the lower part of the Muldoon Canyon Formation of Paull and others (1972) is in thrust contact with the overlying Glide Mountain plate of Nilsen (1977).

7.3 Brockie Subplate of Nilsen (1977)

The Brockie subplate of Nilsen (1977) contains the following stratigraphic units of Paull and others (1972) in ascending stratigraphic order: the upper part of the Muldoon Canyon Formation including the Green Lake Limestone Member, the Brockie Lake Conglomerate, and the Iron Bog Creek Formation. According to Nilsen's model the stratigraphic units in the Brockie subplate are more proximal (relative to the Antler highland) equivalents of the units contained in the Scorpion subplate of Nilsen (1977). According to Nilsen (1977) the Brockie subplate is separated from the Scorpion subplate by the Muldoon Canyon thrust fault, which has repeated the stratigraphy of the Copper Basin Formation. Dover (1980, 1981) suggested that the Brockie subplate of Nilsen (1977) is part of the Glide Mountain plate.

The upper part of the Muldoon Canyon Formation of Paull and others (1972) is considered by Nilsen (1977) to

be the lowermost stratigraphic unit in the Brockie subplate of Nilsen (1977), which is included in the Glide Mountain plate of Dover, 1980, 1981 (see Figure 3-1 and Table 3-1). The upper part of the Muldoon Canyon Formation of Pauli and others (1972) corresponds to the upper 793 meters of type section of the Muldoon Canyon Formation of Pauli and others (1972) which was measured on the east ridge of Muldoon Canyon. At the type section 31 to 122 meters of the Green Lake Limestone Member is found near the top of the upper part of the Muldoon Canyon Formation. The Green Lake Limestone Member of the upper part of the Muldoon Canyon Formation of Pauli and others (1972) is separated from the overlying Brockie Lake Conglomerate of Pauli and others (1972) by 30 meters of argillite which they (Pauli and others, 1972) assign to the upper part of the Muldoon Canyon Formation.

The upper part of the Muldoon Canyon Formation consists primarily of gray silty-, locally shaly-bedded argillite with minor amounts of interbedded dark-gray quartzite and quartzite-chert-argillite granule to pebble conglomerate and micaceous mudstone (Pauli et al., 1972). According to Nilsen (1977) the upper part of the Muldoon Canyon Formation of Pauli and others (1972) consists primarily of argillite with some randomly distributed thinly bedded, carbonaceous, fine-grained facies D or,

less commonly, facies C terrigenous sandstone and siltstone turbidites indicative of basin-plain deposition. The paleogeographic and sedimentologic model of Nilsen (1977) considers the upper part of the Muldoon Canyon Formation of Paull and others (1972) to be the more proximal (relative to the Antler highland) equivalent of the Little Copper Formation of Paull and Gruber (1977).

The age of the upper part of the Muldoon Canyon Formation of Paull and others (1972) is probably predominantly Kinderhookian, the age of the Green Lake Limestone Member.

In the present study the upper part of the Muldoon Canyon Formation of Paull and others (1972) was mapped on the east ridge of Muldoon Canyon. However, this unit is poorly exposed due to an extensive cover of unconsolidated slope deposits, including glacial till, colluvium, and especially talus.

The Green Lake Limestone Member of the upper part of the Muldoon Canyon Formation of Paull and others (1972) is contained within the Brockie subplate of Nilsen (1977). It lies below the uppermost 31 meters of argillite in the upper part of the Muldoon Canyon Formation of Paull and others (1972) and it is variable in thickness, ranging from 31 to 122 meters. The basal

and upper contacts of the Green Lake Limestone Member with the upper part of the Muldoon Canyon Formation of Paull and others (1972) are gradational.

The Green Lake Limestone Member consists of gray-black, silty micritic limestone in 2 to 15 centimeter beds that weather grayish-orange to very light-gray. Thin (1 cm.) interbeds of medium-gray to gray-brown argillites are common. Nodules and thin beds of chert are locally present in the middle of the member (Paull et al., 1972). According to Nilsen (1972) the Green Lake Limestone Member consists of facies D limestone turbidites and interbedded noncalcareous to slightly calcareous argillite. The limestone turbidites are thought to have been derived primarily from eastern source areas and deposited in a basin plain environment, in contrast to the clastic units within the Copper Basin Formation, which were derived from the Antler highland to the west. Grain size is finer and turbidite beds are thinner than in the Drummond Mine Limestone of Paull and others (1972). According to Nilsen's paleogeographic and sedimentologic model the Green Lake Limestone Member is equivalent to the Drummond Mine Limestone of Paull and others (1972). Based on dating of conodont faunas, the Green Lake Limestone Member is Kinderhookian in age, the same age as the Drummond Mine Limestone of Paull and

others, 1972 (Paull, 1981, personal communication).

In the present study the Green Lake Limestone Member was mapped on the east ridge of Muldoon Canyon, where the type section of this unit is found near Green Lake. The Green Lake Limestone Member disappears abruptly approximately 2 km north and the same distance south of the type section. Dover (1980, 1981) reported the discovery of a 75 meter thick limestone which he considers to be equivalent to the Green Lake Limestone Member (based on lithologic associations, thickness, conodont assemblage, and age) in the Glide Mountain plate in Big Rocky Canyon, north of the present study area. According to Dover (1980, 1981) this fact along with other lithologic similarities and continuity of map distribution, indicate that the Brockie subplate and Glide Mountain plate of Nilsen (1977) are equivalent. Structural evidence obtained within the course of the present study supports this hypothesis and will be discussed below.

The Brockie Lake Conglomerate of Paull and others (1972) is contained within the Brockie subplate of Nilsen (1977). The contact of the Brockie Lake Conglomerate with the underlying upper part of the Muldoon Canyon Formation is generally an abrupt lithologic change, but Paull and others (1972) note that at several localities

near Green Lake, within the present study area, a thin gradational zone exists. The upper contact of the Brockie Lake Conglomerate of Paull and others (1972) with their Iron Bog Creek Formation is gradational through an interval of 244 meters at the type section, which is located just outside the eastern border of the present study area. The thickness of the Brockie Lake Conglomerate at the type section ranges from 580 to 730 meters (Paull and others, 1972).

The characteristic lithology of the Brockie Lake Conglomerate of Paull and others (1972) is light gray granule, pebble, and cobble conglomerate in a light gray quartzite matrix. Clasts are generally rounded to sub-rounded, spherical, white to light-gray quartzite. Other subordinate clast types include mudstone and chert. Light-gray quartzite beds without clasts are also common. Minor silty argillite beds can also be found. Conglomerate beds are commonly greater than 3 meters thick, while quartzite and argillite beds are generally thinner (Paull et al., 1972).

Nilsen (1977) identified the Brockie Lake Conglomerate of Paull and others (1972) as middle- and inner-fan deposits consisting of channelized, thickly bedded, facies A pebble and cobble conglomerate with interbedded facies B, C, and D quartzitic sandstone

turbidites and argillite. Paleocurrent directions indicate sediment transport from west to east in an outward radiating pattern with the major source of sediment input near the type section of the Brockie Lake Conglomerate of Paull and others (1972). Conglomerate clasts are largest and conglomerate beds are thickest in this area (Nilsen, 1977). These facts led (Nilsen, 1977) to conclude that the Brockie Lake Conglomerate of Paull and others (1972) represents the proximal portion of a deep sea fan deposit (built into the Antler foreland basin). This conclusion is further supported by the observation of Paull and others (1972) that the unit thins both north and south of the type section.

According to Nilsen's model the Brockie Lake Conglomerate of Paull and others (1972) is the more proximal (relative to the Antler highland) equivalent of the Scorpion Mountain Formation of Paull and others (1972). The predominance of larger, lighter clasts generally distinguishes the Brockie Lake Conglomerate of Paull and others (1972) from their Scorpion Mountain Formation, which generally contains a predominance of smaller, darker clasts. However, beds which are characteristic of the conglomerate of the Scorpion Mountain Formation of Paull and others (1972) are found interbedded within the Brockie Lake Conglomerate of Paull

and others (1972), and vice-versa. This was noted by Schramm (1978), who suggested that the two units merge laterally. Nilsen (1977) used this fact coupled with other sedimentologic and stratigraphic data to develop his paleogeographic model.

In the present study the Brockie Lake Conglomerate of Pauli and others (1972) was mapped on the east ridge of Muldoon Canyon (see plates A and B). The lower contact is well exposed along several stretches of this ridge, whereas the upper contact is not exposed within the present study area, but has been mapped by other workers (Volkman, 1972; Nilsen, 1977) in the Iron Bog Creek drainage area just east of the present study area.

The Iron Bog Creek Formation of Pauli and others (1972) is not exposed within the present study area, but is exposed just east of the present study area (Volkman, 1970). The following description of the Iron Bog Creek Formation of Pauli and others (1972) is included below in order to provide a complete description of the Brockie subplate of Nilsen (1977).

The Iron Bog Creek Formation of Pauli and others (1972) is the uppermost stratigraphic unit in the Brockie subplate of Nilsen (1977), which is included in the Glide Mountain plate of Dover (1980, 1981). The Iron Bog Creek Formation overlies the Brockie Lake Conglomerate with a

gradational stratigraphic contact (Pauli and others, 1972). At the type section of the Brockie Lake Conglomerate of Pauli and others (1972), this contact is gradational through an interval of 244 meters. The true stratigraphic top of the Iron Bog Creek Formation of Pauli and others (1972) has not been recognized. The apparent top of the unit is unconformably overlain by Challis Volcanics, talus and/or alluvium. The thickness of the unit is therefore greater than the 458 meters measured at the type section (Pauli et al., 1972).

The Iron Bog Creek Formation of Pauli and others (1972) consists primarily of interbedded silty, thin-bedded, blocky-weathering argillite and shale, with minor interbeds of quartzite and conglomerate. Color of these rocks ranges from black to medium-gray (Pauli et al., 1972). Nilsen (1977) identified the Iron Bog Creek Formation of Pauli and others (1972) as consisting primarily of argillite with interbeds of facies D quartzitic sandstone turbidites and granule and pebble conglomerate. Paleocurrents indicate sediment transport to the east. According to Nilsen's model the Iron Bog Creek Formation of Pauli and others (1972) is the more proximal (relative to the Antler highland) equivalent of the lower part of the Muldoon Canyon Formation of Pauli and others (1972).

Characteristic trace fossils within the Iron Bog Creek Formation of Paull and others (1972) are of deep-marine Nereites-facies (Nilsen, 1977). Skipp (1974b) reported that trilobites found near the base of the unit indicate an Early Mississippian age for the unit.

7.4 Upper Cretaceous (?) to Lower Eocene (?) Smiley Creek Conglomerate

The Smiley Creek Conglomerate consists primarily of (1) varicolored pebble to boulder conglomerate with well-rounded clasts, and (2) varicolored conglomerate or breccia composed of angular fragments. The former is thought to represent stream gravels, while the latter probably represents talus and other colluvial deposits (Dover, 1981). Quartzite clasts set in a poorly sorted, fairly well cemented matrix are most characteristic of the Smiley Creek Conglomerate. Within the present study area clasts are recognizably locally derived from various stratigraphic units of the Copper Basin Formation. Dover (1981) reported that, regionally, volcanic clasts are absent and plutonic clasts are rare to absent in the Smiley Creek Conglomerate.

The Smiley Creek Conglomerate lies in angular discordance over the deformed Copper Basin Formation and

is disconformable with the overlying Eocene Challis Volcanics. Some bedding is visible, but is poorly developed. The thickness at the unproposed type section of the Smiley Creek Conglomerate is 137 meters, but where the unit is exposed within the present study area, at the southern end of Muldoon Canyon, its thickness is much less (approximately 3 meters).

Dover (1981) reported that the Smiley Creek Conglomerate occupied paleovalleys that existed prior to Challis volcanism and are presently being exhumed by modern erosion. Only three isolated remnants of the Smiley Creek Conglomerate were mapped in the study area, in upper Muldoon Canyon (see Plate A). The significance of these outcrops is discussed below.

7.5 Eocene Challis Volcanics

Within the study area the Challis Volcanics consist of a heterogeneous sequence of flows and pyroclastic beds of a wide compositional range including andesite, latite, dacite, rhyolite, and some basalt. These rocks are varicolored, but are usually shades of yellow, green, brown, red, or gray. Flows and pyroclastic beds are variable in thickness. Wolbrink (1970) reported that textural varieties within the present study area include welded tuffs, crystal tuffs, vitric tuffs, lithic tuffs,

volcanic agglomerates and tuffaceous sedimentary rocks. Both the flows and pyroclastic rocks are generally porphyritic, with phenocrysts (sometimes highly altered) of plagioclase, biotite, quartz and mafic minerals.

The total thickness of the Challis Volcanics is unknown due to post-extrusive erosion. Several workers report that the Challis Volcanics are at least 600 meters thick (Ross, 1962b; Axlerod, 1968; Dover, 1969). Wolbrink (1970) estimates that the Challis Volcanics may have been as much as 900 meters thick within the present study area. These rocks presumably once covered the entire study area but have since been partly eroded away leaving discontinuous remnants along the tops of ridges. In some localities resistant beds form ledges, but in the present study area these rocks are generally non-resistant slope formers.

Within the present study area the Challis Volcanics lie in angular discordance above the deformed Mississippian Copper Basin Formation. In other areas, however, some outcrops of the Challis Volcanics unconformably overlie the Smiley Creek Conglomerate (Nelson and Ross, 1969; Dover, 1981).

The Challis Volcanics crop out extensively on the east ridge of Muldoon Canyon and the west ridge of Starhope Canyon. No attempt was made to map individual

flows or beds, however.

7.6 Eocene Intrusive Rocks

Quartz monzonite intrusive bodies are exposed in the northern part of the study area on both the east and west ridges of Muldoon Canyon. These intrusive bodies are composed mainly of porphyritic pyroxene-hornblende-pliotite quartz monzonite (Dover, 1981) with minor granite (Schramm, 1978) that is fine- to medium-grained with a holocrystalline, porphyritic, granitic texture (Wolbrink, 1970). These stocks and plutons are gray in color and are generally easy to differentiate from the finer-grained rhyolite to andesite dikes which occur throughout the present study area. Wolbrink (1970) reported that the small stocks within Muldoon Canyon are lithologically identical to the Lake Creek Stock (a much larger intrusion) just east of the present study area.

Dover (1981) reported non-systematic cross-cutting relationships between Eocene Quartz monzonite intrusive rocks and the Eocene Challis Volcanics, which suggests that the two are coeval in age. However, Dover (1981) also reported that some of the Eocene quartz monzonite bodies may be slightly younger than the Challis Volcanics.

Dikes and other small hypabyssal intrusive bodies

occur throughout the study area. Major dikes (those recognizable on 1:15,600 scale air photos) were mapped within the course of the present study, but no attempt was made to map them all. Dikes within the study area range from less than 1 meter to approximately 20 meters in thickness. Wolbrink (1970) reports that average thickness of dikes is 2 meters.

Dikes range in composition from rhyolite to andesite, but rhyolite dikes are the most common. Rhyolite dikes are generally porphyritic with phenocrysts of quartz and oligoclase. They range in color from yellow-gray to olive-gray and weather to various shades of yellow and brown. The less common andesite dikes contain phenocrysts predominantly of plagioclase. Andesite dikes are generally dark greenish-gray and weather to brown (Wolbrink, 1970). Some or all of the dikes may be associated with the quartz monzonite intrusive bodies occurring within and in close proximity to the present study area (Dover, 1981).

7.7 Unconsolidated Quaternary Deposits

The glacial geology and geomorphology of the Starhope Canyon-Muldoon Canyon area is reported by Pasquini (1976), Wigley (1976), and Evenson, Cotter, and Clinch (1983, in press) and does not directly bear upon

this study. The reason that unconsolidated Quaternary deposits (undifferentiated) are mapped on Plate A is to document where bedrock is covered by surficial deposits which compromise interpretations of local bedrock structure. Glacial till and glacial outwash are the Pleistocene units which occur in the valleys of Muldoon Canyon and Starhope Canyon. Pleistocene to Holocene units include alluvium, talus, and colluvium.

8. Structural Geology

Structural elements include all components such as surfaces (cleavage, bedding, faults), folds, and lineations that collectively form the geologic structure. Structural or tectonic fabric comprises the complete spatial or geometrical configuration of the structural elements. Structural elements which possess a preferred orientation within a given region are said to be penetrative within that fabric domain. Otherwise, they are non-penetrative.

Table 8-1 is a list of the structural element notation which will be employed in the following discussion of the structural geology of the Starhope Canyon-Muldoon Canyon area. The description, orientation, and configuration of structural elements comprises the main body of data which was obtained during the course of this study. Structural elements which will be discussed include bedding, folds, cleavage, faults, tension gashes, joints and microscopic structures.

8.1 Bedding

Bedding (S⁰) is fairly easy to recognize within the rocks of the Copper Basin Formation for the following reasons:

1. Cleavage (S¹) is not developed to the extent

- S₀ - Bedding
- S₁ - Spaced (fracture) cleavage
- L_{0X1} - Bedding - spaced cleavage intersection
- F₁ - Folding developed during the Glide Mountain plate folding and thrusting episode.
- F₂ - Second folding event: post thrust folding
- a - a kinematic direction; direction of tectonic transport (N70°E - N80°E)
- b - b kinematic direction; fold axes (N10°W - N20°W)
- c - c kinematic direction

Table 8-1:
Structural element notation
designated in the present study.

that it obscures bedding.

2. The contacts between the different stratigraphic units gradationally interfinger such that lithologic layering distinctly define bedding.
3. Although the various stratigraphic units are generally dominated by one lithology, they contain interbeds of other lithologies that distinctly define bedding.

8.2 Folds

Styles of folding differ between the Glide Mountain plate of Nilsen (1977), the Scorpion subplate of Nilsen (1977), and the Brockie subplate of Nilsen (1977). The Glide Mountain plate of Nilsen (1977) is more pervasively and tightly folded than the Scorpion subplate of Nilsen (1977), which is generally homoclinal or gently folded. Styles of folding within the Brockie subplate of Nilsen (1977) vary among the different stratigraphic units comprising the plate.

8.2.1 Folds in the Glide Mountain Plate of Nilsen (1977)

The Glide Mountain plate of Nilsen (1977) characteristically possesses two scales of parallel folds: (1) tight intermediate scale folds with amplitudes and wavelengths which range from tens to hundreds of meters; and (2) smaller, mesoscopic folds with amplitudes and wavelengths both ranging from half a meter to several meters. Both of these fold types within

the Glide Mountain plate of Nilsen (1977) are assigned an F_1 notation. The following pieces of evidence suggest that F_1 folds within the Glide Mountain plate of Nilsen (1977) developed concurrently with thrusting of this plate over the Scorpion subplate of Nilsen (1977):

1. Directly beneath the Glide Mountain thrust (the thrust at the base of the Glide Mountain plate of Nilsen, 1977) the strata of the Scorpion subplate of Nilsen (1977) are highly folded in styles equivalent to the F_1 fold styles in the Glide Mountain plate of Nilsen (1977), which were mentioned above and are described more fully below.
2. Some sheared zones within the Glide Mountain plate of Nilsen (1977) are folded by F_1 folds, whereas others cross-cut F_1 folds. These sheared zones resemble the main Glide Mountain thrust zone at the base of the plate (described more fully below under "Faults"), and are therefore assumed to represent break thrusts and/or splays off the main thrust zone.

It is, of course, possible that some F_1 folding of the Glide Mountain plate of Nilsen (1977) pre-dates thrusting of the plate.

Intermediate scale F_1 folds generally have tight interlimb angles. No isoclinal folds were identified within the Glide Mountain plate of Nilsen (1977) within the course of the present study. Short-limb height to width ratios (as defined by Hansen, 1971) range from 2 to

5. Depth to width ratios of intermediate scale folds are extremely variable where determinable, ranging from 1 to 10; but limited exposures at outcrops make these ratios difficult to determine. The nature of hinges and limbs of these intermediate scale folds is also variable. Hinge geometry ranges from broad to sharp. Limbs of intermediate scale folds range from straight to broadly curved. Assymetry of these folds is dominantly clockwise (as viewed down the plunge of the fold axis), but counter-clockwise assymetries were also observed.

These intermediate scale F_1 folds are generally difficult to recognize as one traverses the slopes within the present study area, due to the fact that both Starhope Canyon and Muldoon Canyon and their associated ridges run parallel to the direction of fold axes (N10° W-N20° W) within the Glide Mountain plate of Nilsen (1977). An excellent exposure of these folds can be seen, however, on the southeastern face of Glide Mountain, where a tributary of Starhope Creek cuts the ridge at an angle approximately perpendicular to fold axes. A typical intermediate scale F_1 fold from the Glide Mountain plate of Nilsen (1977) is shown in Figure 8-1.

The smaller, mesoscopic folds within the Glide Mountain plate of Nilsen (1977) are generally less tight



Figure 8-1:
Typical intermediate scale F₁ fold
in the Glide Mountain plate of the Copper
Basin Formation on the west ridge of Starhope
Canyon south of Glide Mountain. Field assistant
is present in the core of fold for scale.

than the intermediate scale folds. Short-limb height to width ratios of the small scale folds range from 0.3 to 1; and depth to width ratios range from 2 to 5. Hinges of the small scale F_1 folds are generally broad (as opposed to tight) and limbs are straight to broadly curved. Assymetry of the small scale folds is variable. Figure 8-2 shows a typical wave train of small scale F_1 folds within the Glide Mountain plate of Nilsen (1977). At several localities within the study area, small scale F_1 folds were recognizably contained within intermediate scale F_1 folds. It is probable that all of the small scale folds are related to the same stress system that produced the intermediate scale folds, and therefore they are all included as F_1 folds.

Equal-area plots of (A) poles to bedding, (B) F_1 fold axes, and (C) poles to axial planes of F_1 folds in the Glide Mountain plate of Nilsen (1977) within the study area are shown in Figure 8-3. Poles to bedding (Figure 8-3-A) form a clear great circle distribution with a pole plunging at a shallow angle to the north-northwest. This pole is indistinguishable from measured F_1 fold axes in Figure 8-3-B. Hence, the folds are coaxial and cylindrical. The average fold axis orientation is $15^{\circ} N 20^{\circ} W$.

Dover (1981, p. 65) reported that, within the Glide

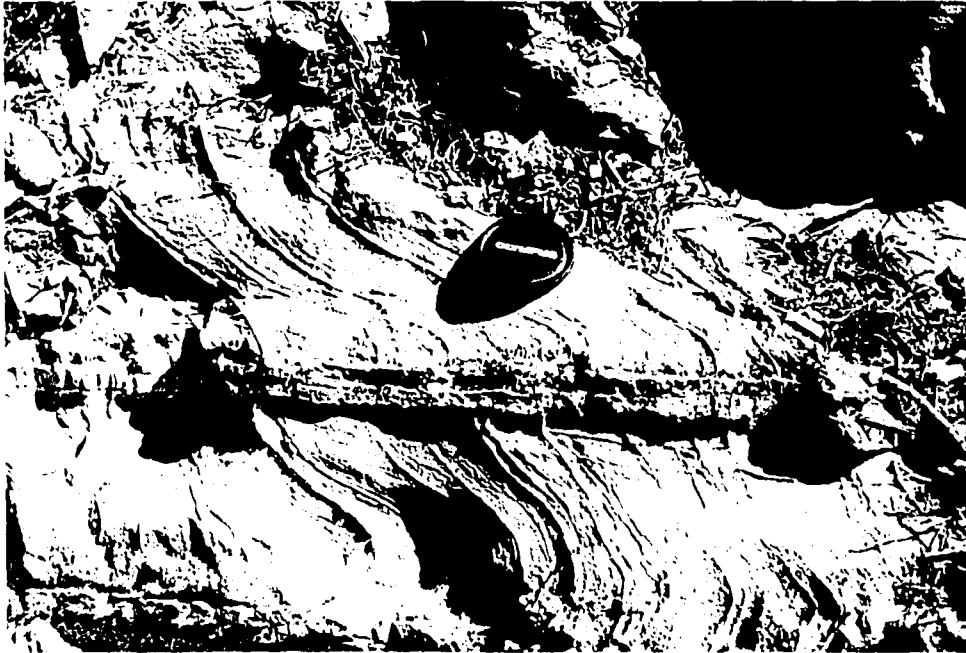


Figure 8-5:
Typical cross-bedding and graded bedding
in the Drummond Mine Limestone of Paul and others
(1972) in Starhope Canyon.

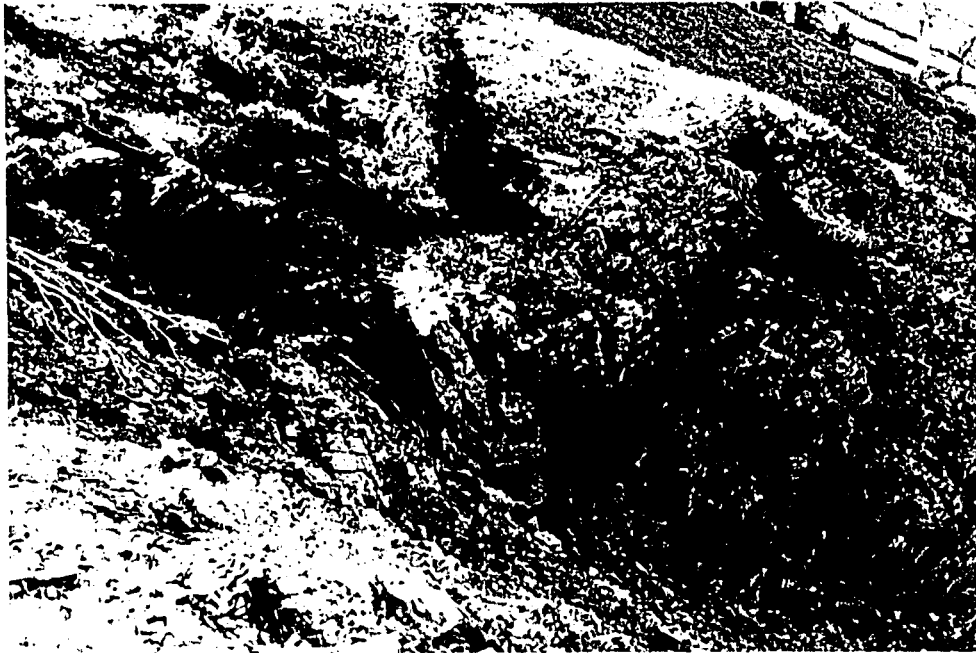


Figure 8-2:
Typical wave train of small scale F¹
folds in the Glide Mountain plate of the
Copper Basin Formation on the west ridge of
Muldoon Canyon. Small pine above folds is
approximately two feet high.

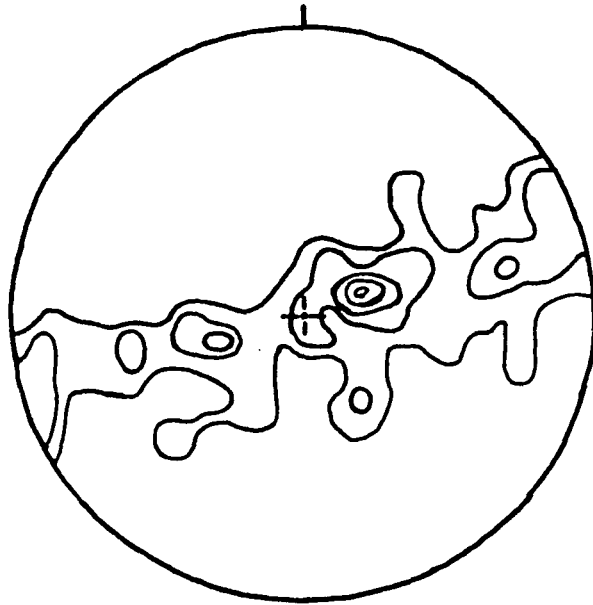
Figure 8-3:
 Equal-area plots of various
 structural elements of the Glide Mountain plate
 of Nilsen (1977) in the Starhope
 Canyon-Muldoon Canyon area:

(A) Poles to bedding showing a major point
 maxima with girdle axis striking N15 W-S15 E;
 max.=72 N69 E; n=77; contours =1,3,5,7,9.

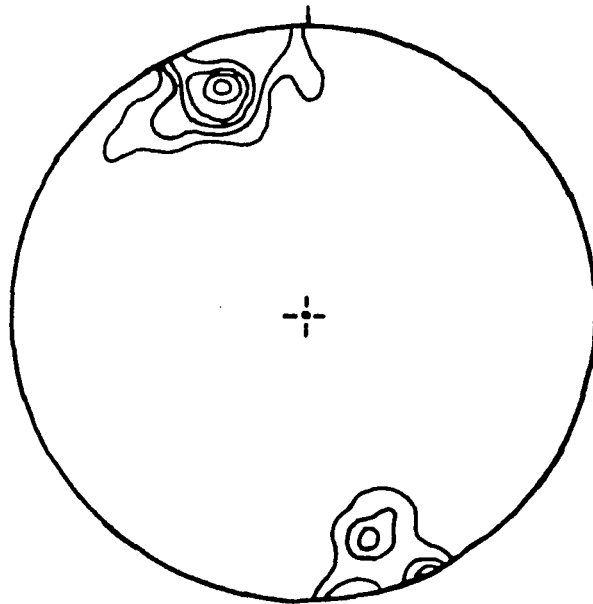
(B) F₁ fold axes showing a strong preferred
 orientation in the N18 W-S18+<0>E direction with
 gentle to moderate plunges in both directions;
 max.=15 N20 W; n=53; contours=2,4,6...12.

(C) Poles to axial planes showing a maximum
 concentration at 28 N60 E; n=45; contours=1,3,5,7.

*
 Contours in this and all succeeding equal-area
 plots are based on the number of points falling
 within a 1% area of the diagram.

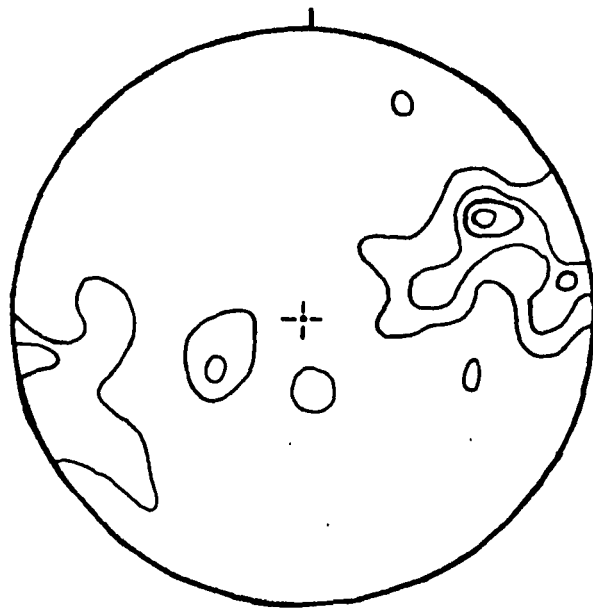


(A)



(B)

Figure 8-3, continued



(C)

Figure 8-3, concluded

Mountain plate, folds are overturned to the east. It can be seen from Figure 8-3-C, however, that although axial planes of F_1 folds dip dominantly westward, they also dip eastward. This dominant dip of axial planes to the west agrees with a tectonic transport of the Glide Mountain plate from west to east, which has been suggested by previous workers (Nilsen, 1977; Dover, 1980, 1981).

The Glide Mountain thrust is recognizably folded on a broad, regional scale in one section of the present study area (sec. 24, T.4N., R.21E.) in upper Muldoon Canyon. Because F_1 folds within the Glide Mountain plate of Nilsen (1977) developed concurrently with (and possibly before) thrusting of the plate, the fold of the thrust zone postdates F_1 folding and is assigned an F_2 notation. F_1 folds within the study area are coaxial with F_2 folds. The previously mentioned variability in orientation of axial planes of F_1 folds probably results from their reorientation during F_1 folding. F_2 folds are discussed below in greater detail.

8.2.2 Folds in the Scorpion Subplate of Nilsen (1977)

In the study area the strata within the Scorpion subplate are dominantly homoclinal or gently folded on a regional scale. This regional folding is generally subtle and can only be recognized by minor changes in the

attitude of bedding as one traverses the study area. Three exceptions to this generalization were observed and these are discussed below.

(1) Where the Glide Mountain plate of Nilsen (1977) has tectonically overridden the Scorpion subplate of Nilsen (1977), the strata of the latter immediately below the Glide Mountain thrust are generally folded (with accompanying shearing and tension gashes) in both tight intermediate scale folds and small scale mesoscopic folds resembling those in the Glide Mountain plate of Nilsen (1977). These folds immediately below the Glide Mountain thrust are especially well developed within the less competent stratigraphic units of the Scorpion subplate of Nilsen (1977), as can be seen in: (a) the Little Copper Formation of Paull and Gruber (1977) on the west ridge of Starhope Canyon, (b) the Drummond Mine Limestone of Paull and others (1972) in the vicinity of Starhope Gulch, and (c) the lower part of the Muldoon Canyon Formation of Paull and others (1972) on the west ridge of Muldoon Canyon. These folds resemble those within the Glide Mountain plate of Nilsen (1977) in every characteristic. Short-limb height to width ratios range from 0.3 to 1; and depth to width ratios range from 2 to 5. Hinges are generally broad (as opposed to tight); and limbs are straight or broadly curved. Assymetry of these folds is

variable. Because these folds lie only directly below the Glide Mountain thrust, it is suggested herein that they developed in response to thrusting of the Glide Mountain plate. The style of these folds is identical to F_1 folds within the Glide Mountain plate of Nilsen (1977), and therefore it is reasonable to assume that they formed in response to the same system of stresses. Therefore the folds occurring directly beneath the Glide Mountain thrust within the Scorpion subplate of Nilsen (1977) are also assigned an F_1 notation.

(2) Several mesoscopic folds were encountered in the Drummond Mine Limestone of Paull and others (1972) in upper Starhope Canyon at vertical distances in excess of 200 meters below the Glide Mountain thrust. These folds occurred as isolated intrafolial folds within the relatively unfolded strata of the Drummond Mine Limestone of Paull and others (1972). A typical example of one of these folds is shown in Figure 8-4. The clockwise vergence of these folds (as viewed down the plunge of the fold axis) was consistent throughout the study area. If these folds were to be considered parasitic (related to folding on a larger scale) they would indicate that the Drummond Mine Limestone of Paull and others (1972) in upper Starhope Canyon is overturned. Cross-bedding and graded bedding in the Drummond Mine Limestone (Figure



Figure 8-4:
Typical mesoscopic F₁ intrafolial
fold within the Drummond Mine Limestone of Pauli
and others (1972) in upper Starhope Canyon.
View is to the south. Fold axis plunges 12°
to the north, therefore fold vergence is clockwise
(as viewed down the plunge of the fold axis).
Wavelength is approximately 8 meters.

8-5) are generally unequivocal in terms of documenting upright sedimentary facings. However, bedding-cleavage relationships (discussed more fully below in "Cleavage") throughout most of the Drummond Mine Limestone in Starhope Canyon indicate that the unit is right-side up. Therefore it is probable that these folds are related either to thrusting of the Glide Mountain plate of Nilsen (1977) over the Drummond Mine Limestone of Paull and others (1972), or thrusting of the Scorpion subplate of Nilsen (1977). These intrafolial folds are therefore assigned an F notation. The axial planes of these intrafolial folds all dip to the east (in contrast to the dominantly west-dipping axial planes of F₁ folds within the Scorpion subplate of Nilsen, 1977, immediately below the Glide Mountain thrust). Consequently, it appears that these folds were tectonically rotated during thrusting of the Glide Mountain plate of Nilsen (1977) over the Drummond Mine Limestone of Paull and others (1972).

(3) At one locality in upper Muldoon Canyon (sec. 24, T.4N., R.21E.) the Glide Mountain thrust fault and the strata of the Scorpion subplate of Nilsen (1977) immediately below the thrust are recognizeably folded in an open (in terms of interlimb angle) regional fold. Because the Glide Mountain thrust is folded, it can be

assumed that at least some of the regional folding of the Scorpion subplate of Nilsen (1977), and the other thrust plates of the Copper Basin Formation, postdates thrusting. An F_2 notation is assigned to this period of folding. It is possible, however, that some of the gentle to open regional folding of the Scorpion subplate may predate thrusting or have developed concurrently with thrusting. These possibilities are impossible to establish, however, with existing field relationships. F_2 folds recognized within the study area display the same axial trend as F_1 folds within the study area (N15 W-S15 E).

Equal-area plots of (A) poles to bedding, (B) F_1 fold axes; and (C) poles to axial planes of F_1 folds within the Scorpion subplate of Nilsen (1977) in the study area are shown in Figure 8-6. Figure 8-6-A displays a maximum concentration of poles to beds at 30 S80 W, illustrating the generally homoclinal nature of the Scorpion subplate of Nilsen (1977) within the study area. Poles to bedding are somewhat spread along a great circle girdle striking N80 E, due to local F_1 folding occurring beneath the Glide Mountain thrust and/or to regional F_2 folding. F_1 folds within the Scorpion subplate of Nilsen (1977), including those folds directly beneath the Glide Mountain thrust and intrafolial folds

Figure 8-6:
Equal-area plots of various
structural elements of the Scorpion subplate
of Nilsen (1977) in the Starhope Canyon-
Muldoon Canyon area:

(A) Poles to bedding showing a major point maximum

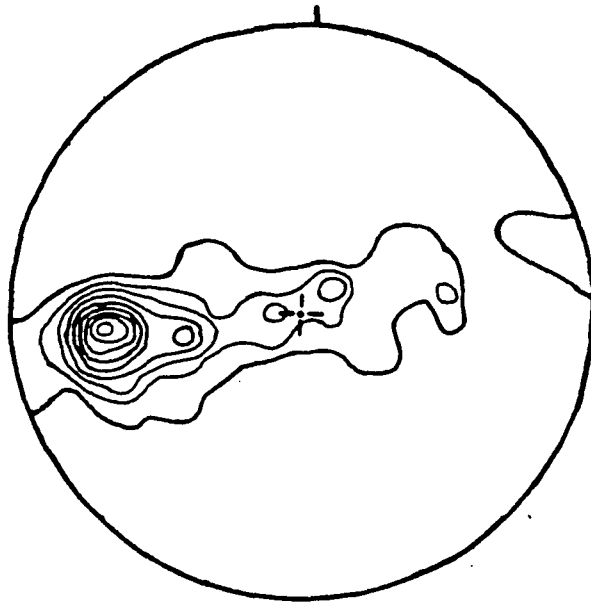
within girdle with axis N10 W-S10 E;
max.=30 S80 W; n=74; contours=1,3,5...17.

(B) F₁ fold axes showing a preferred orientation

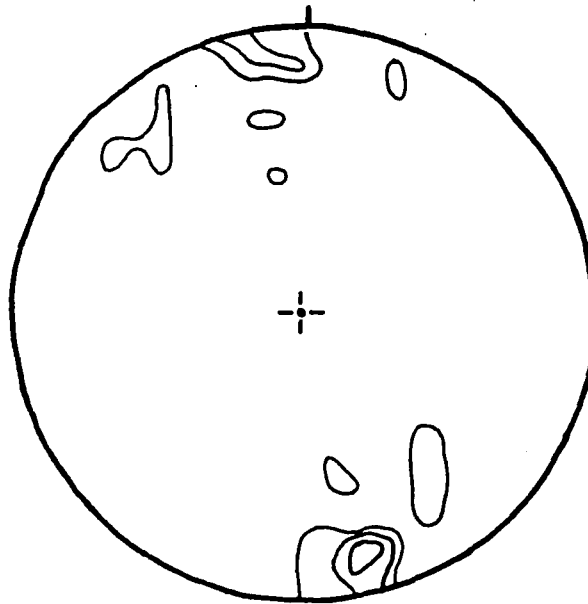
in the N15 W-S15 E direction with gentle to
moderate plunges in both directions; n=18;
contours=1,3,5.

(C) Poles to axial planes of F₁ folds with
several maxima in a poorly defined girdle with

axis N15 W-S15 E; n=14; contours=1,3.

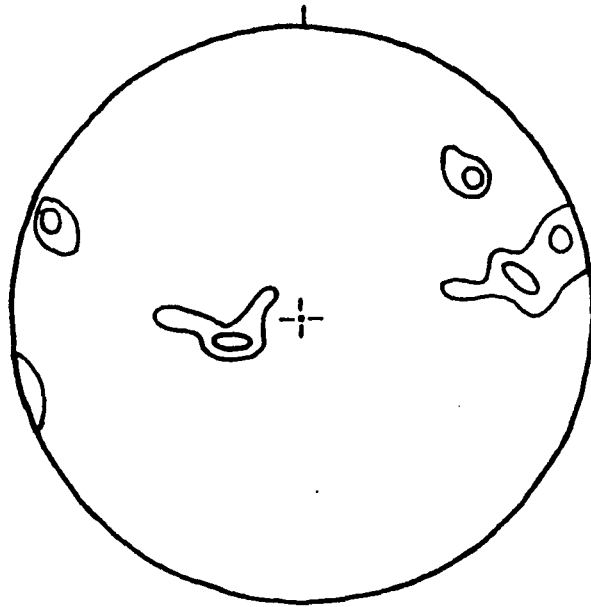


(A)



(B)

Figure 8-6, continued



(C)

Figure 8-6, concluded

within the Drummond Mine Limestone of Paull and others (1972) maintain a statistical fold axis orientation at N15^o W-S15^o E (illustrated in Figure 8-6-B). This orientation is equivalent to that maintained by F¹ fold axes within the Glide Mountain plate of Nilsen (1977).

Poles to axial planes of F¹ folds within the Scorpion subplate of Nilsen (1977), shown in Figure 8-6-C, do not show a well developed point maximum due to the fact that folding within the Scorpion subplate was not extensive enough to yield the number of data points necessary for statistical analysis. It is important to note, however, that axial planes of F¹ folds in the Scorpion subplate of Nilsen (1977) directly¹ beneath the Glide Mountain thrust dipped dominantly to the west (in response to overturning to the east). Axial planes of gentle to open F² regional folds within the Scorpion subplate of Nilsen (1977) were generally vertical or dipping to the west (overturned to the east) where recognizable. Axial planes of F¹ intrafolial folds within the Drummond Mine Limestone of Paull and others (1972) at considerable vertical distances below the Glide Mountain thrust all dip to the east.

8.2.3 Folds in the Brockie Subplate of Nilsen (1977)

Folding is generally difficult to recognize within the Brockie subplate of Nilsen (1977), due to the fact that, in the study area, this "plate" is only exposed along the east ridge of Muldoon Canyon, an area extensively covered by unconsolidated surficial deposits. However, detailed mapping within the course of the present study establishes that the upper part of the Muldoon Canyon Formation of Paull and others (1972) including the Green Lake Limestone Member (both of which comprise the lower part of the Brockie subplate of Nilsen, 1977) are highly folded in styles which are identical to those of F_1 fold styles in the Glide Mountain plate of Nilsen (1977). A typical small scale F_1 fold in the Green Lake Limestone Member is shown in Figure 8-7.

Well-developed folding in the upper part of the Muldoon Canyon Formation of Paull and others (1972) and in the Green Lake Limestone Member occurs as (1) tight intermediate scale folds with amplitudes and wavelengths which range from tens of meters to hundreds of meters, and (2) small scale, mesoscopic folds with amplitudes and wavelengths ranging from half a meter to several meters. Both of these fold types possess similar short-limb height to width ratios, depth to width ratios,

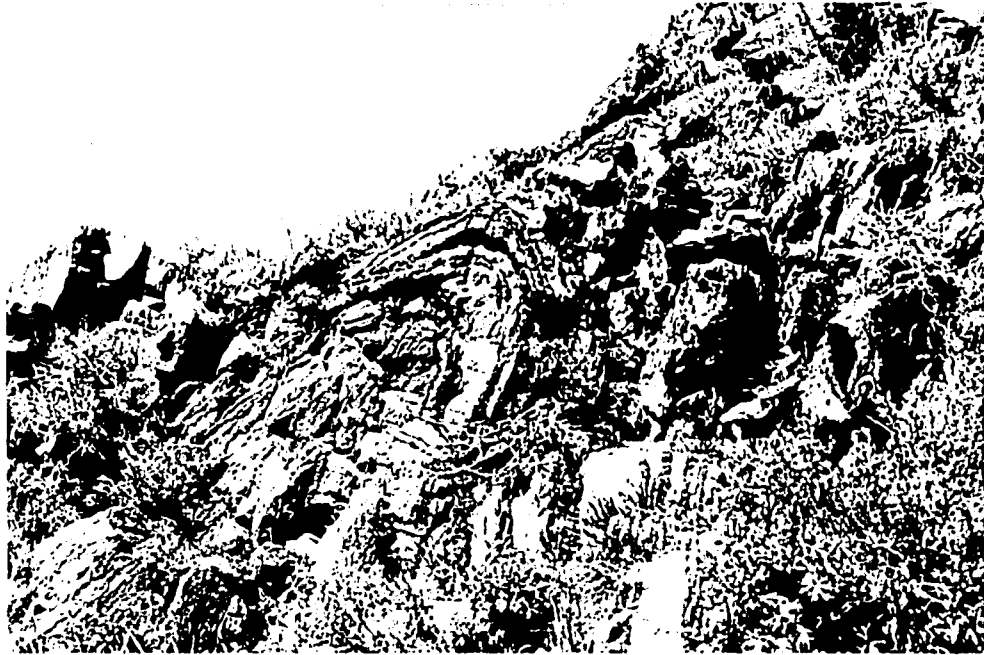


Figure 8-7:
Typical small scale F₁ fold in
the Green Lake Limestone Member. View is to
the north along the east ridge of Muldoon
Canyon.

assymetries, hinges, and limbs to those F_1 folds in the
Glide Mountain plate of Nilsen (1977). It is suggested
here that these are also F_1 folds which developed in
response to thrusting. Further evidence for this is
discussed in the section entitled "Thrust Faults."

Alternatively, the Brockie Lake Conglomerate of
Paul and others (1972), which stratigraphically overlies
their upper part of the Muldoon Canyon Formation, is
homoclinal within the study area. It seems reasonable to
assume that the Brockie Lake Conglomerate in this area
resisted the tendency to fold during thrusting due to its
greater competence than the underlying units. However,
approximately 11 km south of the present study area, in
the valley of Friedman Creek, Lukowicz (1971) mapped
intense folding in the Brockie Lake Conglomerate of Paul
and others (1972) in a plate overlying a thrust fault.

Volkman (1972) reported that the Iron Bog Creek
Formation, which is the uppermost stratigraphic unit in
the Brockie subplate of Nilsen (1977) contains several
recognizable zones of complex folding and faulting along
the Left Fork of Iron Bog Creek in sec. 23, T.4N., R.22E,
to the east of the present study area.

Equal-area plots of (A) poles to bedding, (B) F_1
fold axes, and (C) poles to axial planes of F_1 folds
within the Brockie subplate of Nilsen (1977) in the study

Figure 8-8:

Equal-area plots of various structural elements of the Brockie subplate of Milson (1977) in the Starhope Canyon-Muldoon Canyon area:

(A) Poles to bedding with a maximum concentration

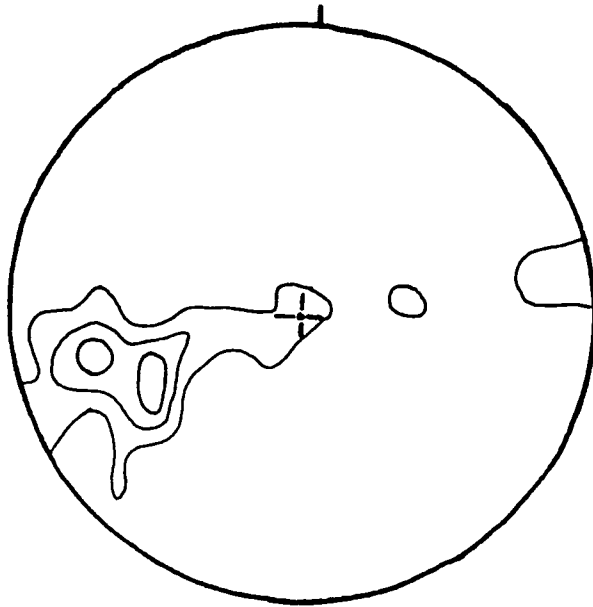
within girdle with axis N20 W-S20 E;
max.=35 S70 W; n=39; contours=1,3,5.

(B) F₁ fold axes showing a strong preferred

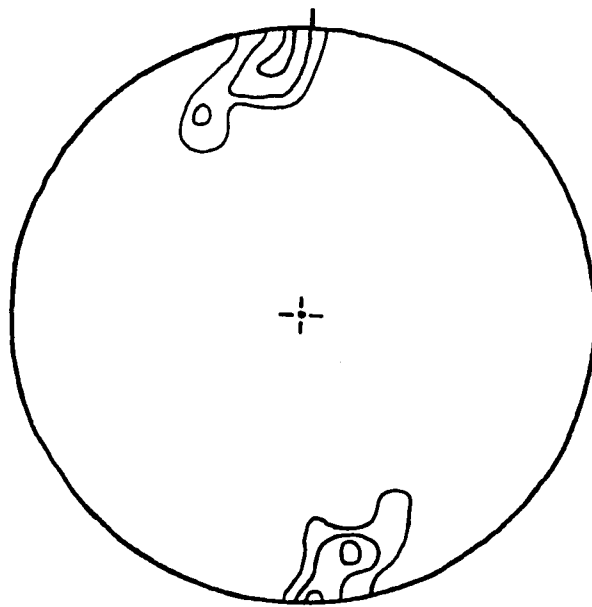
orientation in the N10 W-S10 E direction with
gentle to moderate plunges in both directions;
n=15; contours=1,2,3.

(C) Poles to axial planes of F₁ folds in a

poorly defined girdle with axis N20 W-S20 E;
n=16; contours=1,2,3.

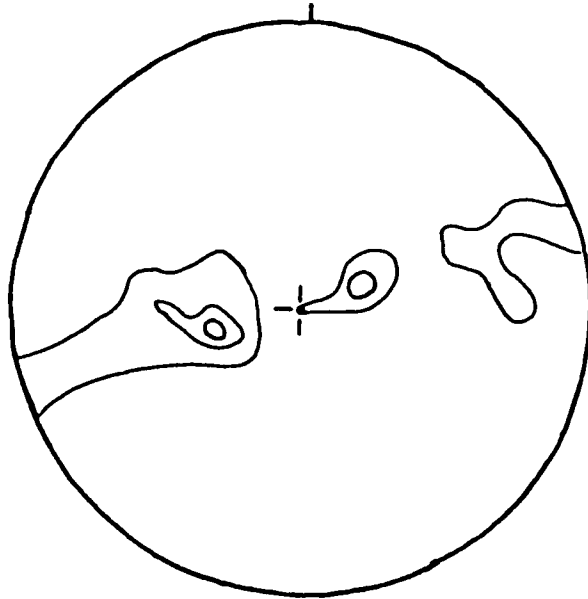


(A)



(B)

Figure 8-8, continued



(C)

Figure 8-8, concluded

area are shown in Figure 8-8. Figure 8-8-A displays a maximum of poles to bedding at 35 S70 W, but poles are somewhat spread along a great circle girdle striking N70 E. Given better exposure of the highly folded upper part of the Muldoon Canyon Formation of Paull and others (1972) and the Green Lake Limestone Member, poles to beds would probably show a greater distribution along the girdle N70 E-S70 W, reflecting a more pronounced effect of F₁ folding. Poles to beds in the Brockie Lake Conglomerate of Paull and others (1972) maintain a tight statistical maximum at 35 S70 W, which accounts for the strong maximum shown in Figure 8-8-A.

Figure 8-8-B illustrates that F₁ fold axes within the Brockie subplate of Nilsen (1977) have been developed with a preferred orientation of N10 W-S10 E with gentle to moderate plunges to both the northeast and southwest. This trend agrees with the orientation of F₁ folds within the Glide Mountain plate and Scorpion subplate of Nilsen (1977). No dominant plunge direction is evident for F₁ fold axes within the Brockie subplate of Nilsen (1977). This may be due to a shortage of data points resulting from poor outcrop exposure.

Poles to axial-planes of F₁ folds within the Brockie subplate of Nilsen (1977) presented in Figure 8-8-C show that these poles maintain an orientation (N70 E-S70 W)

which is coincident to that shown for the same structures in the Scorpion subplate and Glide Mountain plate of Nilsen (1977). However, paucity of data within the Brockie subplate of Nilsen (1977) precludes the determination of any concentration of poles to axial planes in this structural unit.

8.3 Cleavage

Within the study area no slaty cleavage is developed in the rocks of the Copper Basin Formation. A spaced fracture cleavage has formed in some of the less competent rocks (argillites and limestones) and in some of the quartzites of the Glide Mountain plate of Nilsen (1977). The only unit within the Copper Basin Formation that possesses a fairly penetrative fracture cleavage is the Drummond Mine Limestone of Paull and others (1972).

8.3.1 Cleavage in the Glide Mountain Plate of Nilsen (1977)

Dover (1981, p. 64) reported that

"the Glide Mountain plate of the Copper Basin Formation is characterized by pervasive folding on small and intermediate scales and by well-developed axial-plane cleavage."

However, in the Starhope Canyon-Muldoon Canyon area, this is not the case.

Within the present study area, a weak to moderate, spaced (fracture) cleavage has been developed throughout most of the Glide Mountain plate of Nilsen (1977). This spaced fracture cleavage is assigned the notation of S_1 and is most likely a pressure solution-volume loss phenomenon induced by shortening during F_1 folding and thrusting. This is the mechanism that Alvarez and others (1978) propose for the generation of spaced cleavage. Where a fold within the thrust plate has developed in a sequence of rocks which is dominated by argillites, the spaced cleavage is generally moderately developed and "fans" about the axial planes of intermediate scale F_1 folds. Alternatively, where a fold has developed in a sequence of rocks which is dominated by quartzitic beds, the argillite interbeds exhibit a strong bedding plane fissility while the quartzite beds exhibit a radial spaced fracturing. This radial spaced fracturing probably formed at the same time as the spaced cleavage in the other rock types, and, hence, is also considered to be a cleavage. The radial spaced fracture cleavage in the quartzites exhibits a much more pronounced "fanning" around the fold axes, while "axial-planar" cleavage in the argillites "fans" to a much lesser degree. Bedding-cleavage lineations in all cases parallel the fold axes and are assigned a structural notation of L_{OX1} .

The cleavage within the Glide Mountain plate of Nilsen (1977) is more strongly developed in close proximity to (1) the thrust zone at the base of the plate, and (2) sheared zones above the main thrust zone. This fact supports the hypothesis that cleavage (S₁) in the Glide Mountain plate of Nilsen (1977) is related to the folding and thrusting episode (F₁) during which the Glide Mountain plate of Nilsen (1977) was tectonically transported.

8.3.2 Cleavage in the Scorpion Subplate of Nilsen (1977)

The Scorpion subplate of Nilsen, 1977, (Copper Basin plate of Dover, 1980, 1981) is gently folded on a regional scale (F₁) although some earlier F₂ folding can still be found in the present study area. The only stratigraphic unit within the plate which contains a fairly penetrative spaced fracture cleavage is the Drummond Mine Limestone of Pauli and others, 1972 (see Figure 8-9). This spaced fracture cleavage is generally closely spaced and moderately well-developed. The Scorpion Mountain Formation of Pauli and others (1972) is highly jointed and generally lacks a cleavage. Both the Little Copper Formation and the lower part of the Muldoon Canyon Formation of Pauli and others (1972) contain only weakly developed, spaced fracture cleavages. The

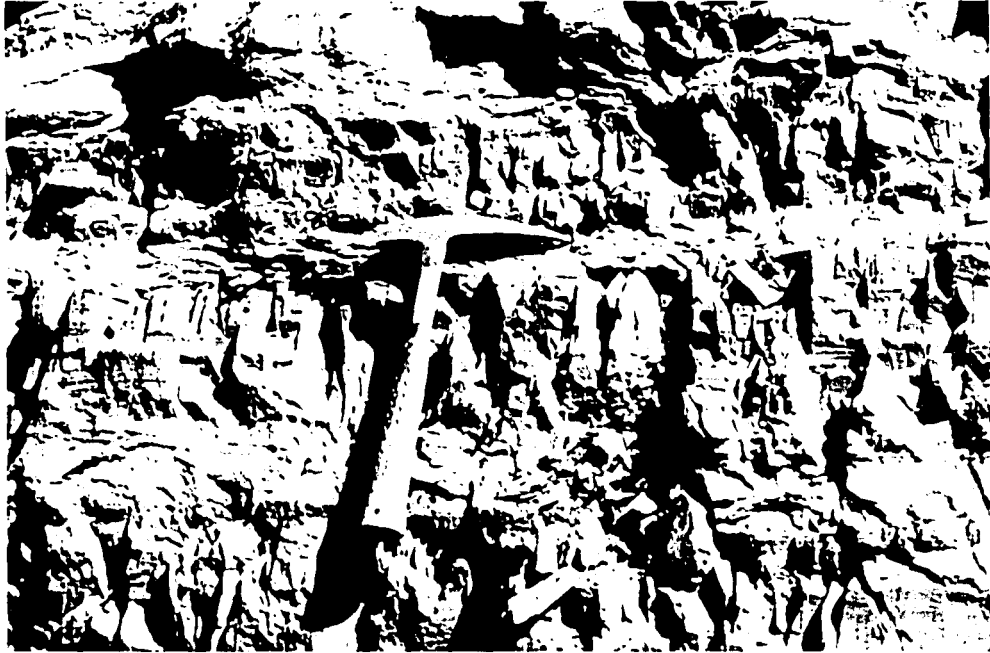


Figure 8-9:
Typical moderately developed
spaced fracture cleavage in the Drummond
Limestone of Paul and others (1972).


cleavage in these units is found only in the most fine-grained, argillaceous beds and directly beneath the Glide Mountain thrust where they have been tectonically overridden. Therefore it is assumed that these cleavages developed during F_1 folding and thrusting of the Glide Mountain plate of Nilsen (1977) over the Scorpion subplate of Nilsen (1977). Bedding-cleavage intersection lineations associated with this cleavage are assigned a notation of L_{OX1} .

Equal-area plots of (A) poles to cleavage, (B) poles to bedding, and (C) bedding-cleavage intersection lineations in the study area within the Drummond Mine Limestone of Paull and others (1972) are shown in Figure 8-10. It is apparent from these plots and the plot of F_1 fold axes within the Scorpion subplate of Nilsen, 1977, (Figure 8-6-B) that cleavage within the Drummond Mine Limestone of Paull and others (1972) is related to F_1 folding. Poles to cleavage (Figure 8-10-A) are distributed along a great circle girdle coincident with bedding (Figure 8-10-B). Bedding-cleavage intersection lineations (Figure 8-10-C) are parallel to F_1 fold axes (Figure 8-6-B) in the Scorpion subplate of Nilsen (1977).


The cleavage in the Drummond Mine Limestone of Paull and others (1972) appears to have developed in response to F_1 folding and thrusting, as did other cleavages below

Figure 8-10:
 Equal-area plots of various
 structural elements of the Drummond Mine
 Limestone of Paull and others (1972) in the
 Starhope Canyon-Muldoon Canyon area:


(A) Poles to S₁ cleavage within a girdle

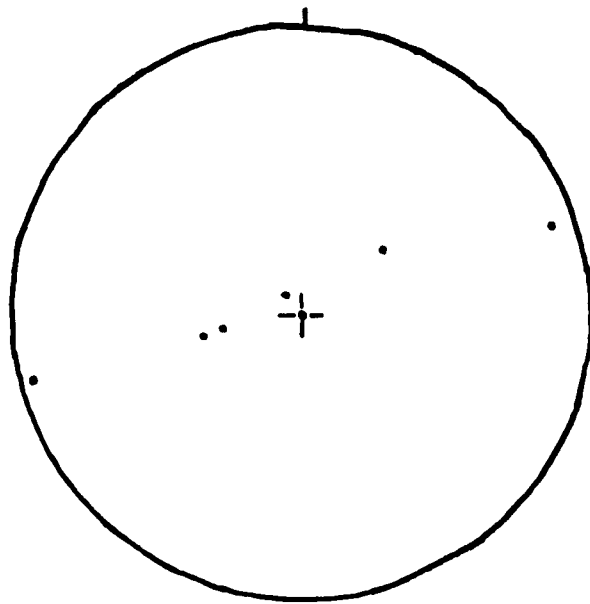

 with axis N20 W-S20 E; n=6.

(B) Poles to bedding showing a major point

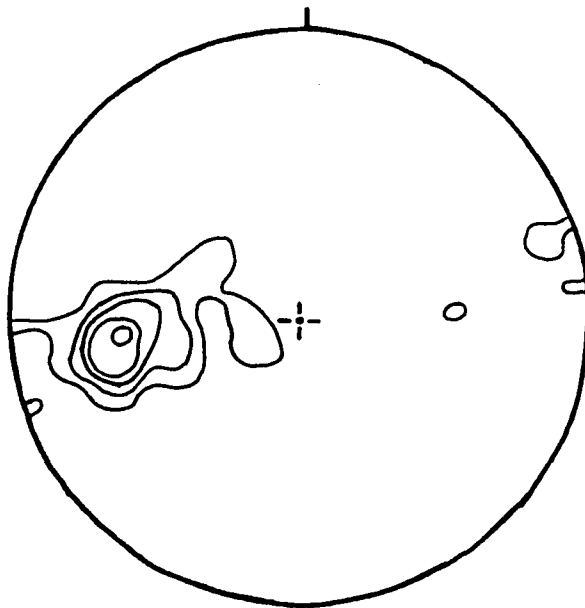

 maximum at 39 S87 W; n=32;
 contours=1,3,5,7,9.

(C) Bedding-cleavage intersection lineations
 (L₁) showing a strong preferred orientation
 OX₁


 in the N16 W-S16 E direction with gentle
 plunges in both directions; max.=10 S8 E;
 n=7; contours=1,2,3,4.

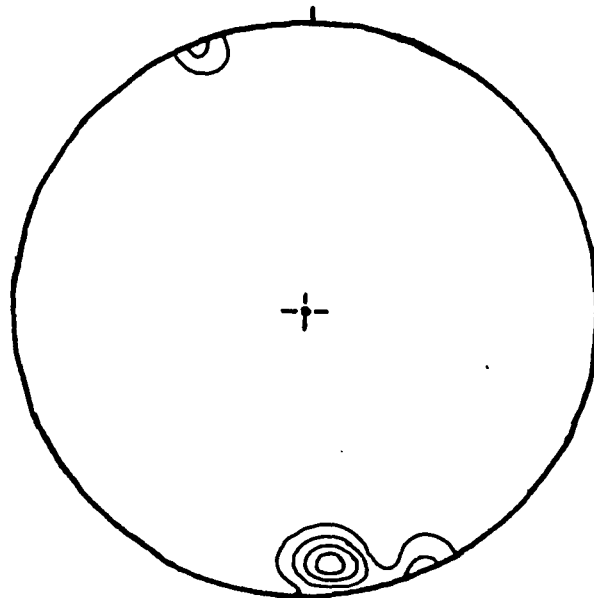


(A)



(B)

Figure 8-10, continued



(C)

Figure 8-10, concluded

the Glide Mountain thrust within the Scorpion subplate of Nilsen (1977). It is highly probable that this cleavage developed concurrently with the development of F₁ intrafolial folds within the Drummond Mine Limestone of Paull and others (1972) since this cleavage "fans" about the axial planes of these folds (see Figure 8-4). F₂ regional folding in the Scorpion subplate of Nilsen (1977) does not seem to have produced any associated cleavage, probably due to the gentleness of the interlimb angle of these folds.

In upper Starhope Canyon most outcrops of the Drummond Mine Limestone of Paull and others (1972) show cleavage dipping more steeply than bedding, indicating that beds are right-side up stratigraphically. At a few outcrops, however, adjacent to Starhope Creek, cleavage is less steeply dipping than bedding, suggesting that bedding is overturned. At these "overturned" outcrops, as well as within most of the outcrops of the Drummond Mine Limestone of Paull and others (1972) within the study area, cross-bedding and graded bedding are unequivocal in terms of documenting upright sedimentary facing (see Figure 8-5). Furthermore, bedding-cleavage relationships cannot be considered unequivocal in terms of documenting upright sedimentary facing where cleavage "fans" about the axial planes of folds. Hence, the

Drummond Mine Limestone of Paull and others (1972) may be locally overturned by F₁ intrafolial folds where beds are more steeply dipping than cleavage adjacent to Starhope Creek, but the unit is dominantly right-side up.

8.3.3 Cleavage in the Brockie Subplate of Nilsen (1977)

A generally weakly-developed, spaced fracture cleavage is present in both the upper part of the Muldoon Canyon Formation and the Green Lake Limestone Member of Paull and others (1972). However, the cleavage in the Brockie subplate of Nilsen (1977) is not as consistently or strongly developed as that in the Drummond Mine Limestone of Paull and others (1972) within the Scorpion subplate of Nilsen (1977). The Brockie Lake Conglomerate of Paull and others (1972), being a massive, competent, quartzitic unit, does not possess any significant cleavage. The Iron Bog Creek Formation of Paull and others (1972) is not exposed within the present study area; consequently its cleavage characteristics were not determined.

The development of a weak, spaced fracture cleavage in the upper part of the Muldoon Canyon Formation and the Green Lake Limestone Member of Paull and others (1972) appears to be related to F₁ folding within these units, and to the thrust at the base of the Brockie subplate of

Nilsen (1977), which will be discussed in more detail below.

8.4 Faults

High-angle faults (dip-slip faults, oblique-slip faults, and strike-slip faults) and thrust fault(s) are all present within the study area. However, most of the major deformation of the Copper Basin Formation within the study area is related to folding and thrust faulting. The Glide Mountain thrust was studied and mapped extensively within the study area. Although Dover (1980, 1981) and Nilsen (1977) report that the lower plate of the Copper Basin Formation (Copper Basin plate of Dover, 1980, 1981; or Scorpion subplate of Nilsen, 1977) is allochthonous, the thrust beneath this plate or penetrative deformational effects associated with the thrusting of this lower plate were not recognized within the study area.

8.4.1 Thrust Faults

8.4.1.1 Glide Mountain Thrust of Nilsen (1977)

The thrust beneath the Glide Mountain plate of Nilsen (1977) was mapped within the course of the present study, and its distribution is outlined on Plate A. This thrust is difficult to recognize in the field for the following reasons:

1. The thrust is largely covered by slope debris (talus, colluvium, and alluvial fans) and glacial deposits, especially in the northern portion of the study area where the thrust is located close to the valley floors.
2. Cataclastic rocks associated with the thrust are often highly weathered.
3. Facies of the Copper Basin Formation above the Glide Mountain thrust in the Glide Mountain plate of Nilsen (1977) are structurally and stratigraphically similar to facies of the Copper Basin Formation within the underlying Scorpion subplate of Nilsen (1977), especially the Little Copper Formation of Paul and Gruber (1977) and the lower part of the Muldoon Canyon Formation of Paul and others (1972). The structural resemblance manifested in highly folded zones below the Glide Mountain thrust, however, is confined to only the less competent units beneath the thrust, and is not characteristic of the entire Scorpion subplate of Nilsen, 1977).
4. The fault zone, being a zone of weakness, is commonly intruded by Eocene intrusive rocks (stocks and dikes).

The Glide Mountain thrust fault was recognized in enough discontinuous exposures, however, to permit its mapping

with a high degree of certainty.

An excellent exposure of the Glide Mountain thrust can be seen in upper Muldoon Canyon (sec. 24, T.4N., R.21E.). At this locality a zone of brecciation marks the thrust (see Figure 8-11). This cataclastic zone is extremely thick (approximately 30 meters at one locality in upper Muldoon Canyon) and its contacts with both the overlying plate (Glide Mountain plate of Nilsen, 1977) and the underlying plate (Scorpion subplate of Nilsen, 1977, or Copper Basin plate of Dover, 1980, 1981) are visible. The fault breccia locally displays iron-staining, bleaching, silicic mineralization, and chaotic fracturing. At most localities the breccia contains angular clasts of the Copper Basin Formation which are variable in size. At a few localities the breccia is a gouge, containing very few angular fragments in a finely granulated matrix. Slickensided zones are generally present within the fault zone, but they do not show a systematic sense of displacement.

As one approaches the thrust zone from either the lower plate (Scorpion subplate of Nilsen, 1977; or Copper Basin plate of Dover, 1980, 1981) or the upper plate (Glide Mountain plate of Nilsen, 1977), the following structural changes can be observed:



Figure 8-11:
Glide Mountain thrust zone in upper
Muldoon Canyon (sec. 24, T.4N., R.21E.).

1. Cleavage becomes more closely spaced and strongly developed and merges into a highly fractured zone immediately above and below the thrust.
2. Brecciated and sheared zones resembling the thrust breccia become more numerous and more intense.
3. Folds become tighter and more numerous.

Sheared zones and other evidence of structural discordance (disrupted bedding and tension gashes) occur in the Glide Mountain plate of Nilsen (1977) at vertical distances in excess of two hundred meters above the Glide Mountain thrust fault. Some sheared zones are folded whereas others cross-cut folds. These sheared zones, which resemble the fault breccia at the base of the Glide Mountain plate, may be splays off of the major thrust zone or local break thrusts developed during F₁ folding. A typical sheared zone in the Glide Mountain plate of Nilsen (1977) found on the west ridge of Muldoon Canyon approximately 200 meters above the main Glide Mountain thrust zone is shown in Figure 8-12.

Structures associated with thrust faulting, such as disrupted bedding, slickensides, micro-faults, and tension gashes, were also locally imparted to the rocks of the Scorpion subplate of Nilsen (1977) where it was tectonically overridden by the Glide Mountain plate of Nilsen (1977). Also, where the Glide Mountain plate of

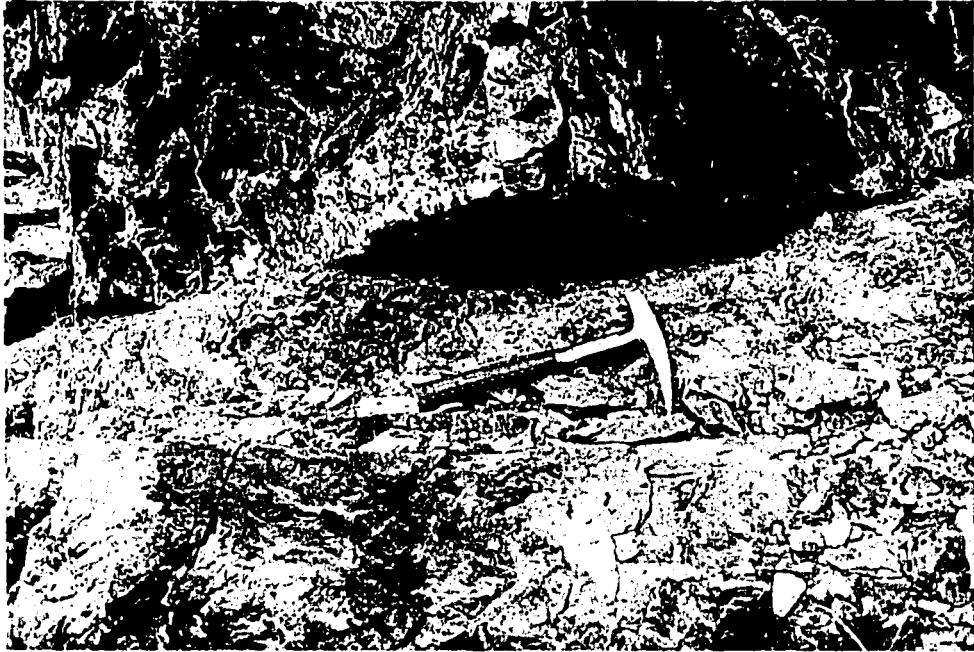


Figure 8-12:
Glide Mountain thrust splay or
break thrust in Glide Mountain plate of
Nilsen (1977) on west ridge of Muldoon Canyon.

Nilsen (1977) has tectonically overridden those less competent stratigraphic units within the Scorpion subplate of Nilsen (1977), the Little Copper Formation, Drummond Mine Limestone, and lower part of the Muldoon Canyon Formation of Paul and others (1972), these rocks are highly folded in styles resembling the folding within the Glide Mountain plate of Nilsen (1977). Locally, the Glide Mountain thrust has incorporated slices of the Scorpion subplate of Nilsen (1977) within the fault zone. One such example of this phenomenon occurs on the west ridge of Starhope Canyon where a block of Drummond Mine Limestone has been transported an indeterminable, but probably short distance along the fault zone. A thin breccia was recognized below this highly folded block of Drummond Mine Limestone. This relationship is shown on the geologic map (Plate A).

An equal-area plot of poles to slickensided surfaces within the Glide Mountain thrust zone is shown in Figure 8-13. It is apparent from this plot that the Glide Mountain thrust generally dips to the northwest within the present study area. This observation is supported by the mapping of the main thrust surface within the study area. At one locality, however, at the southernmost extent of the thrust on the west ridge of Muldoon Canyon, the thrust was observed to be broadly folded (F) into a

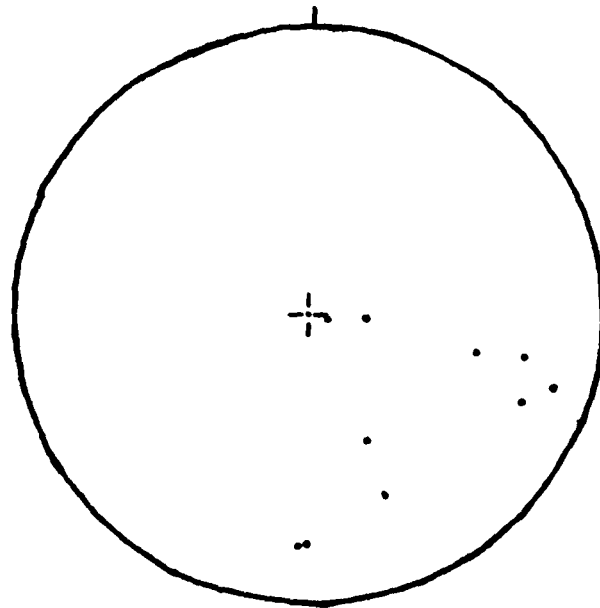


Figure 8-13:
Equal-area plot of poles
to slickensided surfaces within the Glide
Mountain thrust zone in the Starhope
Canyon-Muldoon Canyon area; n=10.

syncline along a fold axis that trends approximately
N15^o W.

Dover and others (1976), Nilsen (1977), and Dover (1981) mapped a portion of the Glide Mountain thrust through the unnamed mountain at the southern end of Muldoon Canyon (sec. 19, T.4N., R.22E.). However, in the present study no thrust fault was mapped at this locality for the following reasons:

1. Excellent exposures of relatively undeformed strata can be seen at this locality which is, in fact, the type section of the lower part of the Muldoon Canyon Formation of Paul and others (1972). The strata on this mountain generally do not display deformational features characteristic of the Glide Mountain plate and are gently folded on a regional scale.
2. A well-developed fault breccia characteristic of the Glide Mountain thrust was not recognized anywhere on this mountain.

Although the map distribution of the Glide Mountain thrust suggests that it should cut through this mountain, it is herein suggested that the Glide Mountain thrust does not exist in this portion of the map area. Two alternative hypotheses can be formulated to explain the absence of the Glide Mountain thrust in this area. Firstly, the thrust surface may have been anticlinally folded (F) over this mountain and subsequently removed

by erosion. Alternatively, the thrust surface may have been displaced upward along normal faults at this locality (or downfaulted in other localities) and then removed by erosion. Either of these hypotheses is equally possible, because both F₂ folding of the thrust and post-thrust normal faults have been recognized elsewhere within the study area.

8.4.1.2 Muldoon Canyon Thrust of Nilsen (1977)

The Muldoon Canyon thrust of Nilsen (1977) separates the Scorpion and Brockie subplates of Nilsen (1977). Nilsen (1977, p. 283) stated that

"the thrust is well exposed at the south end of Muldoon Canyon where a thick zone of gouge and cataclastic material separates the Brockie Lake Conglomerate of Paull and others (1972) from the Scorpion Mountain Formation of Paull and others (1972)."

Nilsen (1977) inferred that the Muldoon Canyon thrust extends from this outcrop down the center of Muldoon Canyon where it is covered by unconsolidated Quaternary deposits. Nilsen (1977) also mapped a small outcrop of the Glide Mountain plate at the headwall of Muldoon Canyon, immediately overlying, and in contact with the Muldoon Canyon thrust. Dover (1980, 1981), however, suggested that the Muldoon Canyon thrust is an extension of the Glide Mountain thrust, thus correlating the

Brockie subplate of Nilsen (1977) with the Glide Mountain plate of Nilsen (1977).

Field mapping undertaken in the present study reinforces the suggestions of Dover (1980, 1981) concerning the Muldoon Canyon thrust. This conclusion is based on the following:

(1) The Scorpion Mountain Formation of Paul and others (1972) crops out extensively on the east ridge of Muldoon Canyon (see Plate A). If the Muldoon Canyon thrust separates the Scorpion subplate from the Brockie subplate, then the "Muldoon Canyon thrust" must actually lie on the east ridge of Muldoon Canyon somewhere above the highest outcrops of the Scorpion Mountain Formation of Paul and others (1972). Placement of the "Muldoon Canyon thrust" on the east ridge of Muldoon Canyon, which is supported by structural evidence discussed below, strongly supports the hypothesis of Dover (1980, 1981) that the "Muldoon Canyon thrust" is an extension of the Glide Mountain thrust. Map distribution of these faults (see Plates A and B) also supports Dover's interpretation.

(2) As previously described, the structures in the Green Lake Limestone Member and the upper part of the Muldoon Canyon Formation of Paul and others (1972) resemble those in the Glide Mountain plate of Nilsen

(1977). The similarity of structural style between the Brockie subplate of Nilsen (1977) and the Glide Mountain plate of Nilsen (1977) strongly supports the hypothesis that these are both parts of the same allochthonous plate. (The Brockie Lake Conglomerate of Paul and others, 1972, is relatively homoclinal where it was mapped within the study area. This unit is much more competent than the underlying Green Lake Limestone Member and upper part of the Muldoon Canyon Formation of Paul and others, 1972, and it apparently resisted F₁ folding in this area.)

(3) Schramm (1878) demonstrated that the Scorpion Mountain Formation and Brockie Lake Conglomerate of Paul and others (1972) are interbedded at the southern end of Muldoon Canyon. Nilsen (1977) also implies this by suggesting that these two units are time-equivalent and facies-equivalent. In the locality where Nilsen (1977) mapped the "Muldoon Canyon thrust," beds of Brockie Lake Conglomerate and beds of Scorpion Mountain Formation are interbedded such that it is mere speculation to suggest that one unit is thrust atop the other.

(4) A fault breccia strongly resembling the fault breccia of the Glide Mountain thrust at other localities within the study area crops out approximately 100 meters east of the locality where the "Muldoon Canyon thrust" is

mapped by Nilsen (1977). This breccia is overlain by highly folded argillites and quartzites and displays silicic mineralization, iron staining, sheared zones and angular clasts of various Copper Basin Formation lithologies. This thrust was mapped as an extension of the Glide Mountain thrust by Dover and others (1976), Nilsen (1977), and Dover (1980, 1981) and this study confirms that interpretation. Mapping on the east ridge of Muldoon Canyon within the course of the present study failed to trace this breccia laterally within the sparse outcrops that occur among the extensive slope deposits (talus, colluvium, alluvial fans, and glacial tills) on this ridge (see Plate A). However, several outcrops did display structures similar those that occur in close proximity to the Glide Mountain thrust (silicic mineralization, chaotic fracturing, and sheared zones). A well-developed breccia was not recognized at the locality where Nilsen (1977) maps the "Muldoon Canyon thrust," although a fractured zone was seen at this locality. This zone is characteristic of deformations which occur where a slice of the lower, Copper Basin plate of Dover (1980, 1981) has been transported with the overlying Glide Mountain plate.

(5) An extensive area of beds of the Scorpion Mountain Formation of Paull and others (1972) which are

overturned to the east were mapped along the lower slopes of the east ridge of Muldoon Canyon in the northern part of the study area. The overturning of these beds is clearly shown by overturned groove casts and graded bedding. This overturning to the east agrees with the eastward direction of thrusting, suggesting that the Glide Mountain thrust should be continued across Muldoon Canyon above these overturned beds and through the east ridge.

Therefore it appears that the structural model of Dover (1980, 1981) best fits the geologic mapping and structural data collected during the course of the present study. Consequently, the Glide Mountain thrust has been projected beneath the unconsolidated Quaternary deposits that mantle the slopes of the east ridge of Muldoon Canyon (see Plates A and B). The mapping of the Glide Mountain thrust on the east ridge of Muldoon Canyon is based on the identification of deformational features recognized to be in close proximity to the Glide Mountain thrust at other localities within the study area. These deformational features were recognized at enough localities to establish reasonable, although tentative mapping of the Glide Mountain thrust on the east ridge of Muldoon Canyon.

8.4.1.3 Stratigraphic Correlation of the Glide Mountain Plate and the Brockie Subplate of Nilsen (1977)

Dover (1980, 1981) suggested that the Brockie subplate of Nilsen (1977) and the Glide Mountain plate of Nilsen (1977) are stratigraphically and structurally equivalent. This interpretation is based on (1) the discovery of a 75 meter thick zone of silty and bioclastic limestone and argillite equivalent to the Green Lake Limestone Member (based on conodont faunas) in the Glide Mountain plate in Big Rocky Canyon (north of the present study area), and (2) other lithologic similarities and the continuity of map distribution (Dover, 1980, 1981). As previously discussed, this interpretation is supported by structural data obtained within the course of the present study.

This interpretation, however valid in the northern Pioneer Mountains, introduces an apparent stratigraphic problem in the Starhope Canyon-Muldoon Canyon area. The Glide Mountain sequence, which consists of highly folded argillite and interbedded quartzitic sandstone turbidite sequences, is well exposed on the west ridge of Muldoon Canyon. The Brockie subplate of Nilsen (1977) occurs on the east ridge of Muldoon Canyon and also contains highly folded argillites and quartzitic sandstone turbidites. The Brockie subplate, however, also contains the Green

Lake Limestone Member and Brockie Lake Conglomerate of Paul and others (1972). Nilsen (1977) proposed that his Brockie and Scorpion subplates represented lateral facies equivalents separated by a thrust. The apparent stratigraphic discontinuity between Nilsen's (1977) Brockie subplate and Glide Mountain plate led him to suggest that the thrust beneath the Brockie subplate (Muldoon Canyon thrust) and the thrust beneath the Glide Mountain plate (Glide Mountain thrust) are two different faults.

However, the observations of Dover (1980, 1981) and structural evidence discussed above suggest that the Brockie subplate and the Glide Mountain plate of Nilsen (1977) are different parts of the same allochthon. In order to accommodate this hypothesis, the stratigraphic discontinuity between the Brockie subplate and the Glide Mountain plate of Nilsen (1977) must somehow be explained. In other words, why aren't the Green Lake Limestone Member and the Brockie Lake Conglomerate of Paul and others (1972) found within the Glide Mountain plate of Nilsen (1977) on the west ridge of Muldoon Canyon and elsewhere within the study area?

The Brockie Lake Conglomerate of Paul and others (1972) is homoclinal and dipping to the east on the east ridge of Muldoon Canyon. Apparently this structurally

competent unit was tectonically transported within the Glide Mountain plate of Nilsen (1977) without becoming highly folded. If this unit were to be geometrically projected across Muldoon Canyon to the west ridge, it would project into space above the present elevation of the top of the ridge. Therefore, it is probable that this unit has been removed by erosion where it may have once occurred above the present west ridge of Muldoon Canyon.

Similarly, the Green Lake Limestone Member may have been removed by erosion in localities other than the east ridge of Muldoon Canyon. Another possible explanation for the absence of the Green Lake Limestone Member on the west ridge of Muldoon Canyon and in Starhope Canyon, however, arises from the fact that the unit thins to the west. The Green Lake Limestone Member may thin to the west so dramatically that it does not exist within the partly terrigenous, most proximal fan facies found in the western part of the Glide Mountain plate.

Based on the assumptions outlined above and the previously discussed structural and stratigraphic similarities, it appears reasonable to assume that the highly folded argillite and quartzite sequences of the Glide Mountain plate of Nilsen (1977) in Starhope Canyon and on the west ridge of Muldoon Canyon are

stratigraphically correlative with the upper part of the Muldoon Canyon Formation of Paull and others (1972) found on the east ridge of Muldoon Canyon. Wolbrink (1970) and Schramm (1978) have also suggested that the highly folded upper thrust plate on the west ridge of Muldoon Canyon contained the Muldoon Canyon Formation of Paull and others (1972).

The nomenclature of Dover (1980, 1981) concerning the allochthonous plates of the Copper Basin Formation appears to incorporate the most reasonable structural interpretation and therefore will be adopted in the remainder of this report. The Copper Basin plate will refer to the lower thrust plate of the Copper Basin Formation (equivalent to the Scorpion subplate of Nilsen, 1977) and the Glide Mountain plate will refer to the upper thrust plate of the Copper Basin Formation (equivalent to the Brockie subplate and Glide Mountain plate of Nilsen, 1977). Dover's (1980, 1981) interpretations have been incorporated into the construction of Plate A, the Geologic Map of the study area, and Plate B, the Geologic Cross-sections through the study area. However, the stratigraphic units of Paull and others (1972) and Paull and Gruber (1977) within the Glide Mountain plate (including the Brockie subplate of Nilsen, 1977) and the Copper Basin plate of

the Copper Basin Formation have been retained as a means of documenting macroscopic structure within the study area.

8.A.2 High-Angle Faults

High-angle faults with small to moderate displacements occur throughout the study area. The amount of displacement is often difficult to determine due to the lack of significant marker beds within the Copper Basin Formation. At one locality on the east ridge of Muldoon Canyon south of Green Lake, several east-west trending vertical to normal faults with displacements ranging from 2 to 20 meters can be recognized by offsets at the abrupt stratigraphic contact between the upper part of the Muldoon Canyon Formation and the overlying Brockie Lake Conglomerate of Paul and others (1972). The Glide Mountain thrust is cut by a younger east-west trending vertical fault just north of Starhope Gulch. On its north side this vertical fault displaces the thrust surface such that it is covered by unconsolidated slope deposits. Consequently the amount of displacement is indeterminable at this locality.

Slickensides are developed along many of the high-angle faults and indicate a variety of displacements, including dip-slip, strike-slip, and

oblique-slip. Dip-slip faults (normal to vertical) are the most common and were recognized in both the Glide Mountain and Copper Basin plates of Dover (1980, 1981). Reverse dip-slip faults, strike-slip faults, and oblique-slip faults were observed in the Glide Mountain plate of Dover (1980, 1981), and also in the Copper Basin plate of Dover (1980, 1981) in close vertical proximity to the Glide Mountain thrust. Thin zones of breccia, gouge, and silicic mineralization commonly occur locally within these fault zones and often obscure slickensides and determination of displacement directions. Drag features are only locally developed and, where present, generally die out within a few meters of these high-angle faults. Because these faults are difficult to recognize and their displacement is difficult to estimate not enough data was gathered for a comprehensive structural analysis. However, the following general relationships were observed:

1. Most of the normal and vertical dip-slip faults recognized within the study area range in strike orientation from N65° E to N90° E. Another, minor set was recognized, however, which ranged in orientation from N30° W to N30° E.
2. Reverse dip-slip faults generally range in orientation from N10° E to N30° W.

3. Strike-slip faults and oblique faults generally range in orientation from N50° E to N90° E.

Exact age relationships for these various high-angle faults are difficult to determine due to (1) the lack of a complete Mesozoic stratigraphic record, and (2) difficulties in tracing these faults in the field. Several of the normal to vertical dip-slip faults were observed to cut the Glide Mountain thrust, and are therefore post-thrusting in age. Many of the normal to vertical faults also cut Challis Volcanics, and are therefore post-Challis in age. However, it was impossible to determine whether any of the pre-Challis normal faults are post-thrusting in age. No faults were observed to displace Quaternary deposits in the study area. Because the reverse dip slip faults generally parallel the major structural fabric (N10° W-N20° W), it is herein suggested that they may represent splays off of the Glide Mountain thrust or associated break thrusts which developed during F₁ folding.

The orientation of the strike-slip faults and the oblique-slip faults (approximately perpendicular to fold axes and parallel to the direction of thrusting) suggests that these may be tear faults related to thrusting of the Glide Mountain plate of Dover (1980, 1981). However, one

strike-slip fault on the northern end of the ridge between Starhope and Muldoon Canyons cuts an Eocene quartz monzonite intrusive body and therefore has at least some motion that post-dates thrusting. This fault strikes N6^o W, which is approximately perpendicular to the direction of tectonic transport during thrusting and is therefore not related to the major thrusting episode. This suggests that some, probably minor, tectonic adjustment within the study area is post-intrusive in age.

The only high-angle faults which are mapped are those that were recognized as having considerable amounts of displacement. In all cases these are normal to vertical dip-slip faults. The Glide Mountain plate of Dover (1980, 1981) locally contains numerous high angle faults of apparently small displacement. These faults are so small and closely spaced that it is impractical to map them at the scale of 1:24,000.

8.5 Tension Gashes

Tension gashes such as those illustrated in Figure 8-14 are ubiquitous throughout the Starhope Canyon-Muldoon Canyon area in both the Glide Mountain plate (including the Brockie subplate of Nilsen, 1977) and the Copper Basin plate of the Copper Basin Formation. In the



Figure 8-14:
Typical set of tension gashes
in the Copper Basin Formation (Glide
Mountain plate) on the east ridge of
Muldoon Canyon.

Green Lake Limestone Member and Drummond Mine Limestone of Paul and others (1972) tension gashes are composed of calcite; whereas in the argillites, quartzites and conglomerates, they consist of quartz. Tension gashes vary in both length (.5 to 2 meters) and width (5 mm to 5 cm).

Some tension gashes are post-deformational joint-fillings (veins) of quartz and calcite, but the following evidence suggests that at least some of the tension gashes were formed during the F₁ folding and thrusting episode:

1. Tension gashes increase in abundance markedly toward the Glide Mountain thrust zone, and occur most frequently in the thrust breccia.
2. Many tension gashes in the Glide Mountain plate (including the Brockie subplate of Nilsen, 1977) are folded by F₁ folds and therefore must be at least slightly older than the folding.

Stereonet analysis of the orientation of all tension gashes showed a random orientation within the Starhope Canyon-Muldoon Canyon area. This is undoubtedly due to the fact that there were several episodes of mineralization of open fractures, such as tension gashes, post-deformational joints and the like, and, unless these mineralized fractures can somehow be separated

temporally, their distribution in space is relatively meaningless.

8.6 Jointing

Jointing is ubiquitous throughout the study area in both the Glide Mountain plate (including the Brockie subplate of Nilsen, 1977) and Copper Basin plate of the Copper Basin Formation. Jointing is especially well developed in the most competent units within the Copper Basin Formation (The Scorpion Mountain Formation and Brockie Lake Conglomerate of Paull and others, 1972). Figure 8-15 shows a typical outcrop of the Scorpion Mountain Formation of Paull and others (1972), illustrating its highly jointed character.

An equal-area plot of poles to joint surfaces within both thrust plates of the Copper Basin Formation is shown in Figure 8-16. This plot was originally subdivided into two separate plots, one for each thrust plate (the Glide Mountain plate, including the Brockie subplate of Nilsen, 1977, and the Copper Basin plate, both of Dover, 1980, 1981). However, because no important differences in orientations of joints in the two plates were found in the two plots, they are presented in Figure 8-16 as a composite plot of poles to 169 joint planes measured throughout the study area. As can be seen in Figure

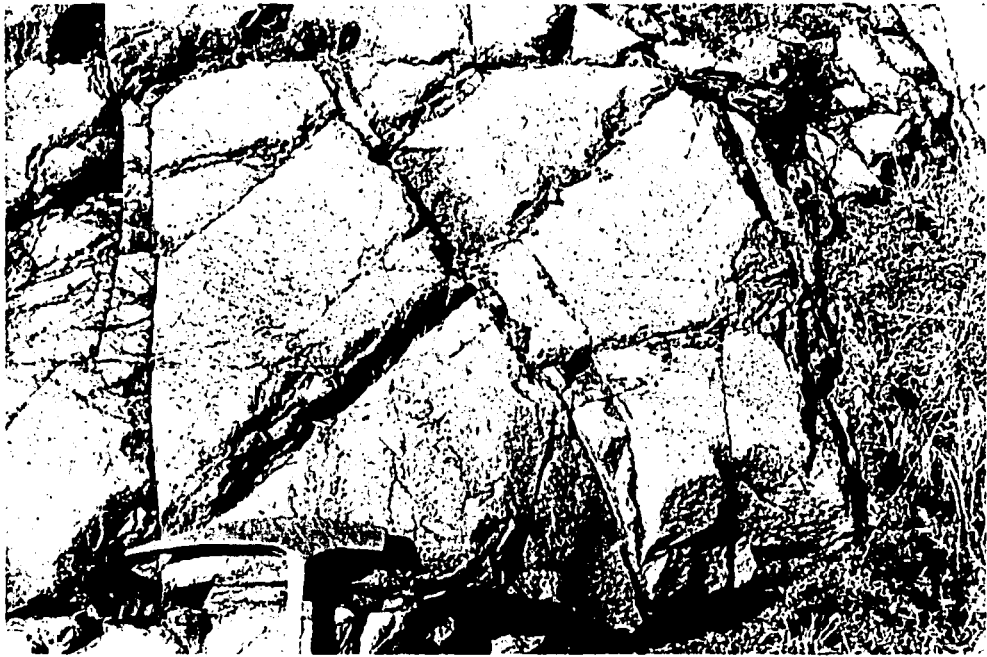


Figure 8-15:
Typical outcrop of the Scorpion
Mountain Formation of Paul and others (1972)
in upper Muldoon Canyon illustrating its
highly jointed character. Note offsets
along joint sets.

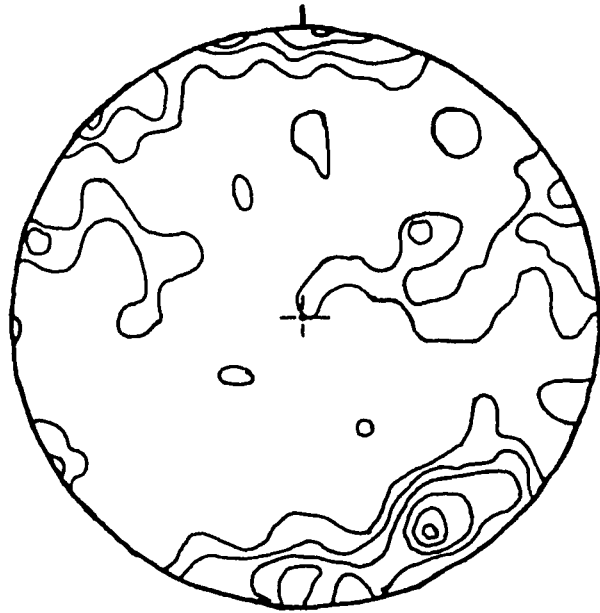


Figure 8-16:
Equal area plot of poles
to joint surfaces within both the Glide
Mountain and Copper Basin thrust plates
of the Copper Basin Formation; n=169;
o o
max.=12 S28 E; contours=2,4,6...12.

8-16, the joint system in the Starhope Canyon-Muldoon Canyon area contains three major sets of joints, all preferentially oriented at high angles to the horizontal. The dominant high-angle joint set, represented by a maximum concentration of poles at $12^{\circ} S 28^{\circ} E$ is oriented approximately perpendicular to fold axes. These are a-c joints. A second high-angle joint set (with poles located at $N 5^{\circ} E - S 5^{\circ} W$) lies close to this orientation and may, in fact, also be part of the a-c joint set. A subordinate joint set is oriented approximately parallel to fold axes (radial joints). This subordinate joint set is also dominantly oriented at high angles to the horizontal, but, as would be expected in a radial joint set, the dip of the joint planes vary with their position in the fold.

It is suggested that the joint system within the Copper Basin Formation formed after the thrusting episode, but before or during Eocene intrusion of quartz monzonite for the following reasons:

1. No joint planes within the present study area were recognized to be folded. If the joints had formed before the F₁ folding and thrusting episode, one would expect to encounter folded joints.
2. Metasomatic alterations and small intrusions occur along joint surfaces in the vicinity of Eocene quartz monzonite intrusive bodies.

Minor offsets along joint planes (up to 5 cm) are common in the massive quartzite and conglomerate beds of the Scorpion Mountain Formation of Paul and others (1972). These are recognizable in Figure 8-15. Northwest-southeast trending (radial) joints are commonly offset along northeast-southwest trending (a-c) joints. This direction of displacement is in agreement with the direction of thrusting and suggests that some minor tectonic adjustments within the study area occurred after the main thrusting episode, as a result of a similar, or possibly the same stress field.

8.7 Relationship of Eocene Intrusive and Extrusive Events to Deformation

Rhyolite dikes, which are probably associated with Eocene quartz monzonite intrusives (Dover, 1980, 1981) are present within both thrust plates of the Copper Basin Formation within the study area. They cross-cut F₁ folds within the Glide Mountain plate (including the Brockie subplate of Nilsen, 1977) and locally they have been intruded along the thrust zone. Two general orientations of rhyolite dikes are found within the study area. The dominant orientation is N30° E-N45° E. A subordinate orientation is N10° W-N20° W, which parallels the general structural trend that is expressed by many of the

structural elements of the Copper Basin Formation.

Eocene quartz monzonite intrusive rocks likewise are present in both the Glide Mountain plate and Copper Basin plate of the Copper Basin Formation. However, only a few outcrops on the northern part of the east ridge of Muldoon Canyon show that the Copper Basin plate has been intruded (see Plate A). These facts suggest that Eocene intrusion was a post-thrusting (and post-F₁ folding) event. Dover (1980, 1981) also states that intrusion post-dates thrust-faulting throughout this region. However, Dover (1981) also reported shearing along contacts between quartz monzonite intrusives and the Copper Basin Formation at the Glide Mountain thrust zone northwest of the present study area. This suggests that some, probably minor, tectonic movement along the Glide Mountain thrust was post-intrusive. This hypothesis is supported by the strike-slip fault which cuts a quartz monzonite intrusive at the north end of the ridge between Starhope and Muldoon Canyons (discussed above).

Dover (1981) proposed that plutonism in the Pioneer Mountains is dominantly post-orogenic, and is responsible for intrusive doming of both thrust plates of the Copper Basin Formation. It may be that minor, post-orogenic movement along the Glide Mountain thrust and within the Glide Mountain plate itself are related to this intrusive

doming and/or other post-orogenic tectonic adjustments. Intrusive doming may also be partly or wholly responsible for the absence of the Green Lake Limestone Member and Brockie Lake Conglomerate of Paul and others (1972) within the Glide Mountain plate west of the east ridge of Muldoon Canyon.

Within the Starhope Canyon-Muldoon Canyon area, Eocene Challis Volcanics unconformably overlie only the Glide Mountain plate, and not the Copper Basin plate (see Plate A). However, to the northwest of this study area, near the junction of Summit Creek and the Big Lost River, Dover (1981) mapped the Challis Volcanics atop the Copper Basin plate and the Glide Mountain plate. The mapping of Dover (1980, 1981) and the fact that the Challis Volcanics are thought to be slightly older than or coeval with Eocene intrusives clearly shows that extrusion of the Challis Volcanics postdates thrusting. Furthermore, within the course of this study, the Smiley Creek Conglomerate was mapped atop both the Glide Mountain plate and Copper Basin plate of the Copper Basin Formation. Therefore the Smiley Creek conglomerate is post-thrusting in age; and since the Smiley Creek Conglomerate is older than the Challis Volcanics, the Challis Volcanics must also be post-thrusting in age.

The mapping of Dover (1981) establishes the fact

that, regionally throughout the Pioneer Mountains, the Challis Volcanics have been deposited primarily atop the upper (Glide Mountain) thrust plate of the Copper Basin Formation. This occurrence suggests that pre-Challis erosion did not expose large portions of the Copper Basin plate by removal of the overlying Glide Mountain plate. Therefore, it is assumed that at least some of the deformation, including F₂ folding and intrusive doming, which has locally upwarped the Copper Basin plate within the study area, is post-Challis in age. Also, Dover (1981) shows that the outcrops of Challis Volcanics on the west ridge of Starhope Canyon and the east ridge of Muldoon Canyon are continuous with extensive areas of Challis cover to the west and east of the study area. The ridge between Starhope and Muldoon Canyons is devoid of Challis Volcanics. Furthermore, the southern end of the west ridge of Starhope Canyon, where the Copper Basin plate has been upwarped and the overlying Glide Mountain plate has been removed by erosion, is also devoid of Challis Volcanics (see Plate A). However, several workers (Wolbrink, 1970; Dover, 1981) suggest that the Challis Volcanics once covered most, if not all of the present study area. These facts suggest that the entire central portion of the study area may be uplifted by intrusive doming and/or some other post-Challis tectonic

adjustment. This hypothesized uplift would have resulted in extensive removal of both the Challis Volcanics and the upper stratigraphic units of the Glide Mountain plate (Green Lake Limestone Member, Brockie Lake Conglomerate, and Iron Bog Creek Formation of Paull and others, 1972) within the central part of the study area. This hypothesis is compatible with all mapping and structural data obtained during the course of the present study.

8.8 Microscopic Structure

The orientation of optic axes (0001) in quartz was determined for several oriented hand samples of the Copper Basin Formation obtained from within the study area in order to determine whether or not deformations within the area had been penetrative on a microscopic scale. Samples of quartzite from various units within the Copper Basin Formation, a sample of argillite from the Glide Mountain plate, and a sample of the Glide Mountain thrust breccia, were all analyzed for quartz fabrics using the techniques described in Turner and Weiss (1963). In all cases, plots of quartz optic axes from 300 grains showed a random distribution, indicating that deformations of the Copper Basin Formation in the Starhope Canyon-Muldoon Canyon area were not penetrative to the microscopic scale. The lack of a strongly

developed cleavage in the Copper Basin Formation in the study area is compatible with the results of this analysis. However, it should be noted that microfibrils in the carbonate units within the Copper Basin Formation (Green Lake Limestone Member and Drummond Mine Limestone of Paul and others, 1972) were not measured and the presence or absence of penetrative deformation in these units has not been determined.

9. Summary of Deformation

The D_1 and D_2 notations, which correspond to two different deformational periods, have not been established within the course of this study due to the fact that F_1 and F_2 folding within the Copper Basin Formation display the same axial trend ($N10^{\circ}W-N20^{\circ}W$) and may be the result of a stress continuum, and not necessarily two different tectonic pulses.

Field relationships support the interpretation that F_1 folding (on small and intermediate scales), which is characteristic of the Glide Mountain plate, developed concurrently with the thrusting of the Glide Mountain plate. This is based on the following observations:

1. The same style of folding and structural discordance which is characteristic of the Glide Mountain plate is locally developed beneath the Glide Mountain thrust in the Copper Basin plate.
2. Some sheared zones within the Glide Mountain plate, which represent break thrusts and/or splay off of the main thrust zone at the base of the plate, are folded, whereas others cross-cut folds. These structural relations suggest that, although some folding of the Glide Mountain plate may have occurred before thrusting of the plate, at least some (and possibly most) of the folding within the Glide Mountain plate occurred concurrently with thrusting.

Although tectonic transport of the Glide Mountain

plate was principally accommodated along one major thrust zone (Glide Mountain thrust), displacements in the a kinematic direction ($N70^{\circ}E-N80^{\circ}E$; perpendicular to fold axes) also occurred along splays off of the main thrust and/or along break thrusts. Strike-slip faults, which are interpreted to be tear faults due to their preferred orientation in the a kinematic direction, also accommodated movement of the Glide Mountain plate.

Dover (1980) reported a tectonic shortening of approximately fifty percent within the Glide Mountain plate. This shortening was accomplished by the development of pervasive F_1 folding and a generally weakly developed spaced fracture cleavage. F_1 fold axes define the b kinematic direction ($N10^{\circ}W-N20^{\circ}W$). Petrofabric analysis performed during the course of this study has established that tectonic shortening of the Glide Mountain plate (and the Copper Basin plate) was not penetrative on a microscopic scale (at least not within the quartzites and argillites of the Copper Basin Formation).

Effects of the thrusting of the Glide Mountain plate were imparted to the Copper Basin plate immediately beneath the Glide Mountain thrust zone. These effects include the development of: (1) F_1 folds displaying the

same fold styles as F_1 folds in the overlying Glide Mountain plate, (2) sheared zones, (3) chaotic cleavage and fracturing, (4) micro-faults, (5) slickensides, (6) tension gashes, and (7) disrupted bedding. It is also possible that the moderate, spaced fracture cleavage and intrafolial folds within the Drummond Mine Limestone of Pauli and others (1972) were formed and rotated during overthrusting of the Glide Mountain plate.

The Glide Mountain plate and Copper Basin plate of the Copper Basin Formation display radically different structural styles. The Glide Mountain plate (with the exception of the Brockie Lake Conglomerate of Pauli and others, 1972) is highly folded by pervasive small and intermediate scale F_1 folds with associated weakly developed spaced fracture cleavage and sheared zones. Both the Glide Mountain plate and Copper Basin plate are regionally folded by gentle to open F_2 folds, but the Copper Basin plate generally lacks the pervasive F_1 folds. This structural contrast is one of the most useful means by which the Glide Mountain plate and Copper Basin plate of the Copper Basin Formation can be distinguished; although recognition of the Glide Mountain thrust zone separating the two plates is still the best method by which to differentiate them.

Gentle to open folds on a regional scale within the

Copper Basin Formation are assigned an F notation. Because the Glide Mountain thrust zone is folded in an F²-style fold, it is apparent that at least some of this folding occurred as a post-thrusting deformation. However, this does not necessarily mean that F² folding is entirely post-thrusting in age. Some F² folding may have occurred as pre-thrusting or concurrent with thrusting, but these possibilities are indeterminable given the existing field relationships within the study area.

High-angle normal to vertical dip-slip faults cut through the Glide Mountain thrust zone in some localities and the Challis Volcanics in other localities. It is apparent that at least some of these faults are not only post-thrust, but also post-Challis in age; but whether or not some of these faults were post-thrust, pre-Challis displacements can not be determined within the study area. Therefore, it is assumed that all high-angle normal to vertical dip-slip faults within the study area are post-thrust, and most probably post-Challis in age.

Well developed joint sets were observed in both the Glide Mountain plate and the Copper Basin plate of the Copper Basin Formation. This joint system is post-thrust but pre-intrusive in its development. Minor offsets along joints are developed along the a kinematic

direction, suggesting that these post-jointing tectonic adjustments within the study area may have developed under the same stress field that produced F_1 and F_2 folding, or a later stress field with very similar kinematic axes. Offsets along joints may also have been produced by later movements related to intrusive doming of the study area.

One strike-slip fault within the study area was observed to cut an Eocene quartz monzonite intrusive body, suggesting that some, minor tectonic adjustment within the study area was post-intrusive. This fault is oriented approximately perpendicular to the direction of tectonic transport related to folding and thrusting, and is therefore unlikely to be related to the dominant regional structural fabric.

Based on regional considerations throughout the Pioneer Mountains, Dover (1980) has demonstrated that all datable thrusts (including the Glide Mountain thrust) are post-Middle Permian to pre-Eocene in age. No data obtained, or structural relations observed, within the course of the present study can modify or improve this dating. The lack of a definitive Mesozoic stratigraphic record prevents more precise dating. If thrusting occurred during the late Paleozoic to Early Mesozoic, a single, long-lived stress field exemplified by F_1 and F_2

folding and minor offsets along joint sets, may have dominated this region for a considerable span of geologic time. It is equally possible that the different deformational events outlined above developed under different (albeit similar in orientation) stress fields which resulted from different tectonic pulses. If thrusting occurred at a period of geologic time closer to the Cenozoic, the various deformational events outlined above may have occurred in steady succession as a result of a single or several closely related tectonic pulse(s).

10. Conclusions

(1) the structural model of Dover (1980, 1981), which suggests that the Copper Basin Formation is contained within two allochthonous plates, best fits the mapping and structural data obtained during the course of this study in the Starhope Canyon-Muldoon Canyon area. The upper Glide Mountain plate (including the Brockie subplate of Nilsen, 1977) has tectonically overridden the Copper Basin plate (equivalent to the Scorpion subplate of Nilsen, 1977).

(2) The Glide Mountain plate of the Copper Basin Formation is much more highly deformed by folding and thrusting than the Copper Basin plate of the Copper Basin Formation. Throughout the Glide Mountain plate, tight folding is developed on intermediate and small scales, and shearing occurs along break thrusts and/or splays off of the main thrust zone at the base of the plate (Glide Mountain thrust). The Copper Basin plate is only tightly folded and sheared directly beneath the Glide Mountain thrust zone.

(3) The Glide Mountain thrust is marked by a zone of intense shearing and cataclasis which is extremely thick (thirty meters in one locality). Deformation within this zone has produced a chaotic fault breccia which locally displays iron-staining, bleaching, silicic

mineralization, and chaotic fracturing.

(4) The thrust at the base of the Copper Basin plate is not exposed within the Starhope Canyon-Muldoon Canyon area. Thrusting of the Copper Basin plate apparently produced very few deformational effects within the plate.

(5) F_1 folding resulted in (a) intricate folding of the Glide Mountain plate of the Copper Basin Formation on small and intermediate scales, (b) intricate folding of the less competent strata of the Copper Basin plate immediately below the Glide Mountain thrust, and probably (c) the development of intrafolial folds within the Drummond Mine Limestone of Paull and others (1972) within the Copper Basin plate.

(6) Some F_1 folding may have occurred before thrusting of the Glide Mountain plate, but much of this F_1 folding occurred concurrently with the thrusting.

(7) Weakly to moderately developed spaced fracture cleavage (S_1) within the Glide Mountain plate and Copper Basin plate of the Copper Basin Formation in the study area developed in response to F_1 folding and thrusting.

(8) Deformations associated with F_1 or F_2 folding or thrusting within both thrust plates of the Copper Basin Formation did not produce any preferred orientations of quartz c-axes (i.e. these deformations were not penetrative to the microscopic scale).

(9) F_2 folding resulted in gentle to open regional folds in both thrust plates of the Copper Basin Formation, as well as the Glide Mountain thrust zone separating these two plates.

(10) Some F_2 folding may have occurred before or during thrusting of the Glide Mountain plate of the Copper Basin Formation, but much of this F_2 folding occurred after thrusting of the Glide Mountain plate.

(11) F_2 fold axes are developed along the same axial trend as F_1 fold axes ($N10^{\circ}W-N20^{\circ}W$).

(12) Some or all F_1 folds may have been the result of a stress continuum associated with F_2 folding, and not necessarily a different tectonic pulse.

(13) The central portion of the study area may be intrusively domed. This hypothesis is supported by the absence of the Challis Volcanics and upper stratigraphic units within the Glide Mountain plate (Green Lake Limestone Member, Brockie Lake Conglomerate, and Iron Bog Creek Formation of Paull and others, 1972) from the central portion of the study area.

(14) All high-angle normal to vertical dip-slip faults with small to moderate displacements within the Starhope Canyon-Muldoon Canyon area are post-thrust, and most probably post-Challis in age.

(15) A well developed joint system is present in both thrust plates of the Copper Basin Formation in the Starhope Canyon-Muldoon Canyon area. This joint system is composed of three joint sets: a dominant set of a-c joints, another high-angle joint set which may be part of the a-c joint set, and a subordinate set of radial joints.

(16) The joint system in the Copper Basin Formation was post-thrusting and pre-intrusive in its development.

(17) Eocene Challis Volcanics and Eocene quartz monzonite intrusive bodies were both emplaced after thrusting of the Glide Mountain plate of the Copper Basin Formation.

(18) Some minor tectonic adjustments within the Copper Basin Formation are post-thrusting, post-jointing in age, as indicated by minor displacements along joints in the Copper Basin Formation. These displacements are developed in the a kinematic direction (N75° E) of the regional structural fabric and may have developed under the same stress system that produced F₁ and F₂ folds.

(19) Some minor tectonic adjustments within the rocks of the Starhope Canyon-Muldoon Canyon area are post-intrusive in age, as indicated by one strike-slip fault through a quartz monzonite intrusive body. This displacement is oriented approximately perpendicular to

the a kinematic direction and is not related to the dominant regional structural fabric. This displacement may have been produced by intrusive doming or some other post-thrusting tectonic adjustment.

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Vita

Robert C. Marshall was born in Plainfield, New Jersey on March 24, 1957. He is the son of Mr. Matthew G. Marshall of Ballymena, Northern Ireland, and Ms. Carol L. Marshall of East Orange, New Jersey. He attended Hackettstown Middle School and Hackettstown High School, from which he was graduated in June, 1975.

Mr. Marshall graduated from Moravian College in September, 1979 with a Bachelor of Science degree in Geology (in cooperation with Lehigh University). After a semester in the real world, he began graduate studies at Lehigh University in January, 1980, working toward the Master of Science degree in the Department of Geological Sciences. During his tenure at Lehigh University, Mr. Marshall was a Research Assistant (May, 1980-May, 1981) and Teaching Assistant (June, 1981-December, 1982) in the Department of Geological Sciences. He left Lehigh University in February, 1983, to pursue a career in the petroleum industry with the Shell Oil Company in Houston, Texas.

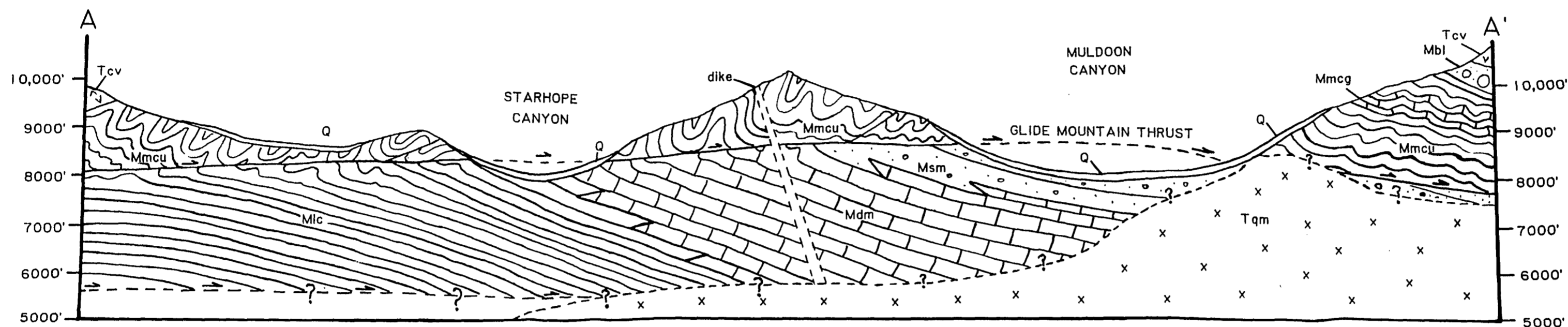
Mr. Marshall is a member of the American Association of Petroleum Geologists, the American Geophysical Union, and the Society of Economic Paleontologists and Mineralogists. His main geological interests lie in the fields of structural geology, tectonics, stratigraphy,

and sedimentology.

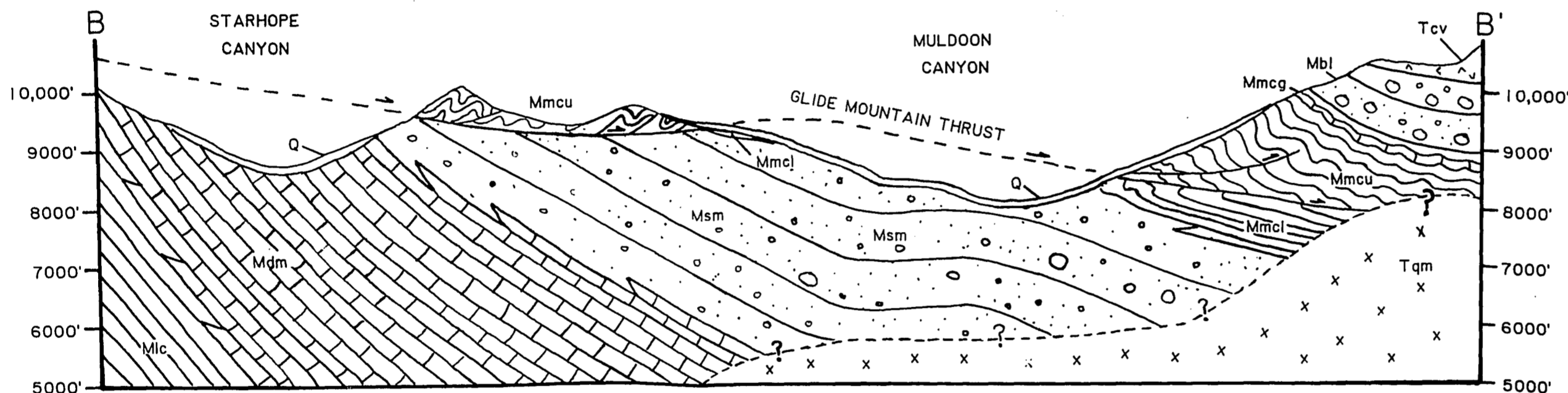
Plates A and B

GEOLOGIC CROSS SECTIONS

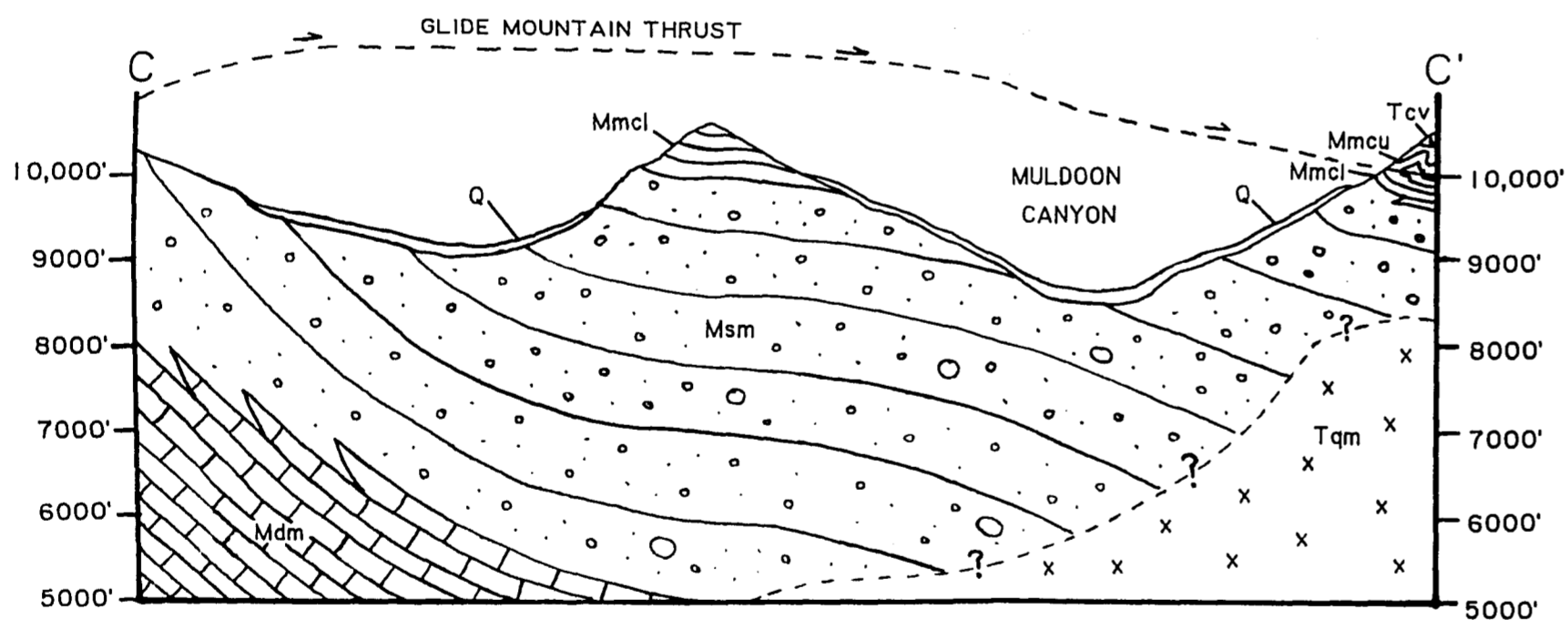
TO ACCOMPANY PLATE A



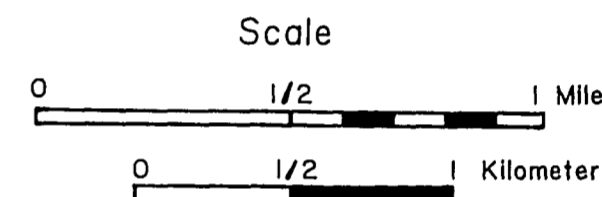
Cross Section A-A'



Cross Section B-B'



Cross Section C-C'



No Vertical Exaggeration

Explanation

- Q Unconsolidated Quaternary Deposits, Undifferentiated—includes alluvium, talus, colluvium, & glacial drift
- Tqm Eocene Quartz Monzonite Intrusive Stocks
- Tcv Eocene Challis Volcanics
- KTsc Upper Cretaceous (?) to Lower Eocene (?) Smiley Creek Conglomerate

Mississippian Copper Basin Formation

Glide Mountain Plate of Dover (1980, 1981) Copper Basin Plate of Dover (1980, 1981)

- Mlb Iron Bag Creek Formation of Paull and others (1972) Not present in study area.
- Mbl Brackie Lake Conglomerate of Paull and others (1972)
- Mmcg Green Lake Limestone Member of Muldoon Canyon Formation of Paull and others (1972)
- Mmcu Upper part of Muldoon Canyon Formation of Paull and others (1972)
- Mmcl Lower part of Muldoon Canyon Formation of Paull and others (1972)
- Msm Scorpion Mountain Formation of Paull and others (1972)
- Mdm Drummond Mine Limestone of Paull and others (1972)
- Mic Little Copper Formation of Paull and Gruber (1977)

Lithology

- Argillite
- Conglomerate
- Limestone
- Quartzite
- Intrusives
- Volcanics

Symbols

- Contact, dashed where inferred
- Facies change
- Thrust fault, arrows indicate direction of tectonic transport

Note:

1. Orientation of cross sections A-A', B-B', and C-C' is N75°E-S75°W (no vertical exaggeration).
2. Folded bedding and splays off Glide Mountain Thrust in Glide Mountain Plate and dike are shown schematically.

Geology by
Robert C. Marshall,
1983

GEOLOGIC MAP OF THE STARHOPE CANYON -MULDOON CANYON AREA, COPPER BASIN, PIONEER MOUNTAINS, SOUTH-CENTRAL IDAHO

