The completion of this thesis
was supported by the
Randy Seeling Award
given, in his memory, to another
outstanding graduate student
of the Geology Department,
University of Minnesota, Duluth.

THE POLYMETAMORPHISM OF THE
LITTLE WILLOW FORMATION
WASATCH MOUNTAINS, UTAH

A THESIS
SUBMITTED TO THE FACULTY OF THE GRADUATE SCHOOL
OF THE UNIVERSITY OF MINNESOTA

BY
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IN PARTIAL FULFILLMENT OF THE REQUIREMENTS
FOR THE DEGREE OF
MASTER OF SCIENCE

May, 1980

"Perhaps part of the mission of a natural science, like ours is, should be to define the role of randomness in historical processes. For it is true that randomness, or anarchy, is in some measure the law of nature, and that some of the order we think we see is a dream in our own minds. As geologists we are in a better position than most to set the limits to the anarchy, and also limits to the dream."

Krauskopf, 1968

ABSTRACT

The Little Willow Formation, of apparent Middle Precambrian age, crops out along the Wasatch Fault Zone, in the foothills of the Wasatch Mountains, approximately 13 miles south-southeast of Salt Lake City, Utah. This metamorphic complex, exposed at the intersection of the Uinta and Wasatch tectonic trends, is composed of quartzofeldspathic gneisses, pelitic mica schists, amphibolites and a heterogeneous migmatitic unit. The formation is unconformably overlain to the east by quartzites and phyllites of the late Precambrian Big Cottonwood Formation; intruded on the southeast by the Tertiary Little Cottonwood quartz monzonite stock; and is covered and bounded on the remaining sides by Quaternary glacial and alluvial sediments.

The migmatite is believed to be part of the Little Willow Formation, not the Big Cottonwood. Both the migmatite and the typical Little Willow are characterized by spaced schistosity, crenulated foliation, similar orientation of the S_1 foliation, recrystallized granoblastic textures and moderate-grade metamorphism. The Big Cottonwood displays a minor phyllitic foliation which is subparallel to bedding, minor recrystallization and except for thermal effects close to the stock, a low-grade metamorphic assemblage.

Metamorphic phenomena of several regional events are recognizable within Little Willow rocks. Spaced schistosity, crenulated foliation and rotated porphyroblasts of andalusite and garnet are among the evidence for the Middle Precambrian regional metamorphism. During the Laramide Orogeny (late Cretaceous-early Tertiary) the unconformable contact between the Little Willow and Big Cottonwood was apparently the locus of

thrust movement. This has resulted in a cataclastic texture in many of the rocks. Intrusion of the Little Cottonwood quartz monzonite stock, 24-31 million years ago, thermally metamorphosed the surrounding rocks. Estimated metamorphic pressures of approximately 4 kb and temperatures in excess of 600^{o}C appear to have allowed preservation of some regional features while initiating some new ones. Characteristics of this thermal event include recrystallization of the earlier cataclastic textures, growth of second-generation andalusite, growth of sillimanite (possibly second generation) and replacement of foliated sillimanite by muscovite. Wasatch normal faulting, which occurred about 20 million years ago, and subsequent erosion are responsible for the uplift and exposure of the Little Willow Formation and apparently formed localized zones of re-brecciated mylonite within the recrystallized rock.

Migmatitic structures, foliated sillimanite and evidence for possible anatexis found in the southeastern portion of the Little Willow area are easiest explained by the regional metamorphic event; however some question remains, since their development during the later thermal event is also a possibility.

ACKNOWLEDGMENTS

I wish to express my sincere gratitude to Dr. James A. Grant who suggested this study and served as my mentor and advisor. He provided much enthusiasm, guidance and expertise throughout all aspects of this project.

The Geology Department of the University of Minnesota, Duluth provided money for the preparation of some thin sections, which is gratefully acknowledged.

The suggestions and helpful criticism of Drs. John Green and Ronald Morton, who also served on my graduate committee, and of Dr. Tim Holst, are greatly appreciated.

Special thanks go to Dr. T. R. Neff of Weber State College, Ogden, Utah, who provided useful suggestions and some of his personal hand samples and thin sections for the study.

I also extend my thanks to Dave and Lola Britton and family, who helped make my stay in Utah enjoyable, to Rick Foss and Terry Langer, for their companionship and assistance in the field, to Jack Swenson, for the use of his truck, to Jim Domenico of the USGS, for his encouragement, and to Jackie Gollinger, who helped with the typing of the manuscript.

The needed criticism, support, encouragement and friendship of my fellow grad students, especially Elizabeth Ann Palmer, Bruce Yeomans, Tim Flood, Peter Giangrande, Diane Carlson and Neil Rismeyer, is cherished and appreciated, as is the comraderie of the entire geology department.

Finally, I wish to thank my folks, who first introduced me to the joy of understanding the world.

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INTRODUCTION

THE PROBLEM

This thesis concerns the polymetamorphism of the Precambrian Little Willow Formation in the Wasatch Mountains southeast of Salt Lake City, Utah (Fig. 1). Metamorphic phenomena seen in the Little Willow rocks are apparently the product of at least three metamorphic events, including regional metamorphism, dynamic (cataclastic) metamorphism, and contact (thermal) metamorphism. This study of the Little Willow Formation and associated rocks was conducted to provide a detailed description of the metamorphic rocks and to integrate the textural evidence and mineral assemblage data into a comprehensive history of the polymetamorphism.

LOCATION AND ACCESS

The Little Willow complex crops out at the mouth of the Little Cottonwood Canyon, with outcrops extending north along the Wasatch Front, in the foothills, to Deaf Smith Canyon (formerly Little Willow Canyon) (Fig. 2). The specific area studied includes parts of Sec. 1, 2, 11, 12 of T.2S, R 1 E., as well as some unsurveyed national forest land to the east. Utah Highway 210, south from Salt Lake City, provides good access to the western perimeter of the study area and Little Cottonwood Canyon. Two dirt trails, former mining roads, one in Deaf Smith Canyon and one originating in the former Gold City, afford access, by foot, into the interior of the outcrop area. Recent housing developments at the mouths of Deaf Smith and Paulson Canyons have altered the natural landscape, removed some potential outcrop and complicated access into the foothills.

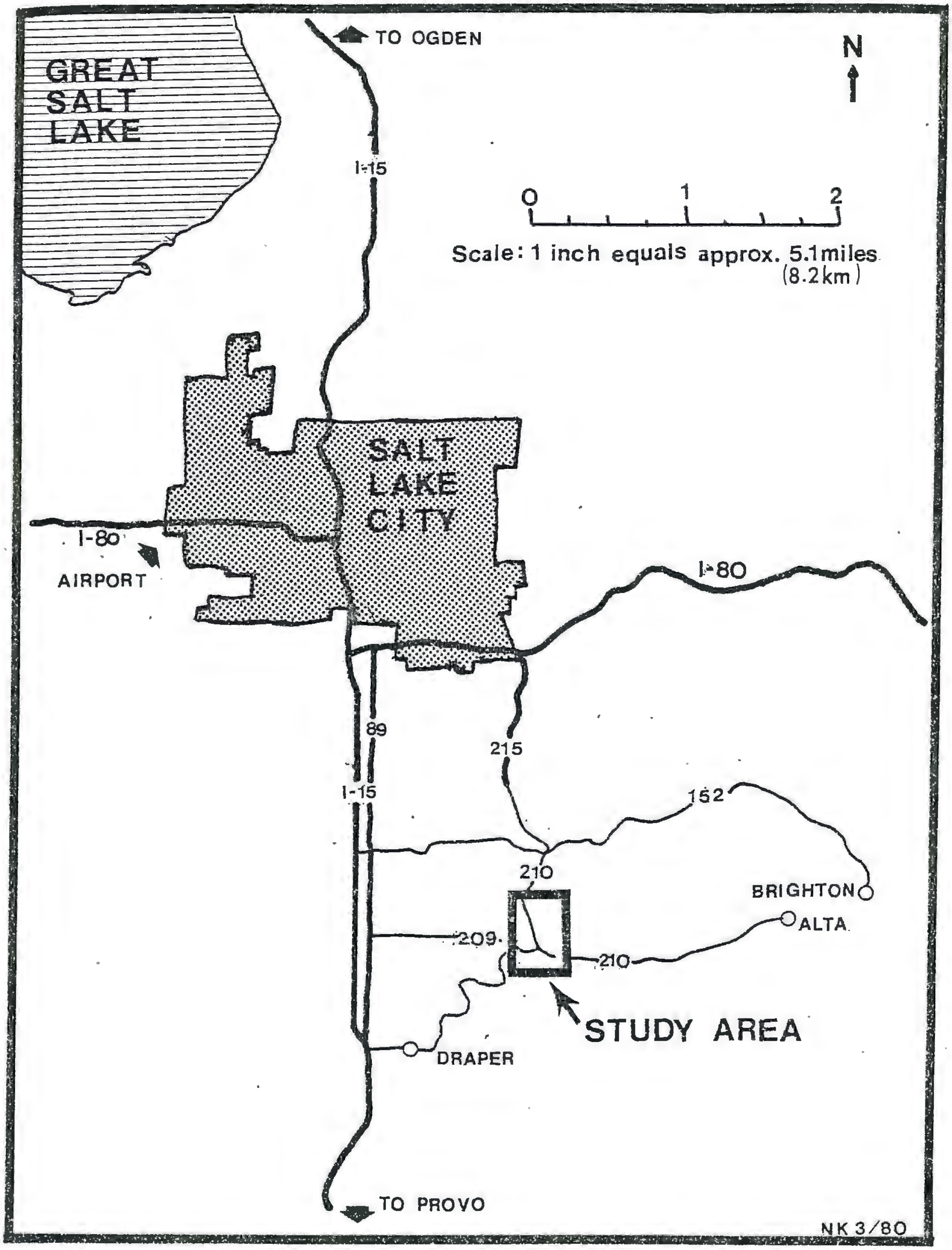


Figure 1. Location of study area.

PHYSIOGRAPHY

Although the Little Willow Formation is located in the foothills of the Central Wasatch Mountains, topography within the study area is characterized by an impressive three-dimensional display of the Wasatch Front. An airphoto of the area shows high, sharp ridges separated by steep-walled canyons; the topographic map illustrates elevation increases of over 1000 feet per 2000 feet of horizontal distance (Fig. 2). The mountains end abruptly to the west, where they meet the Jordan Valley, a downfaulted block, covered with many hundred feet of sediments (Fig. 3). The only significant topography variation on this sediment plain are Pleistocene lake terraces which once rimmed Lake Bonneville and fault scarps in unconsolidated material, due to recent movement along the Wasatch Fault.

As much as 90 per cent of the foothills are covered with *Quercus gambelii* (scrub oak), the heaviest cover being found on the north slopes (Fig. 4). A few other types of deciduous trees, such as cottonwoods, can be found growing on the banks of rivers and creeks. Conifers occupy the mountains at higher elevations.

Wildlife is rare because of the proximity of residential areas; however, numerous small rodents, lizards and rattlesnakes still make their home in these foothills.

These canyonlands were placed in a forest reserve in 1904 and are protected today because of their importance as a watershed for Salt Lake City.

PREVIOUS INVESTIGATIONS

Hague and Emmons, working with King's 40th Parallel Survey (1871),

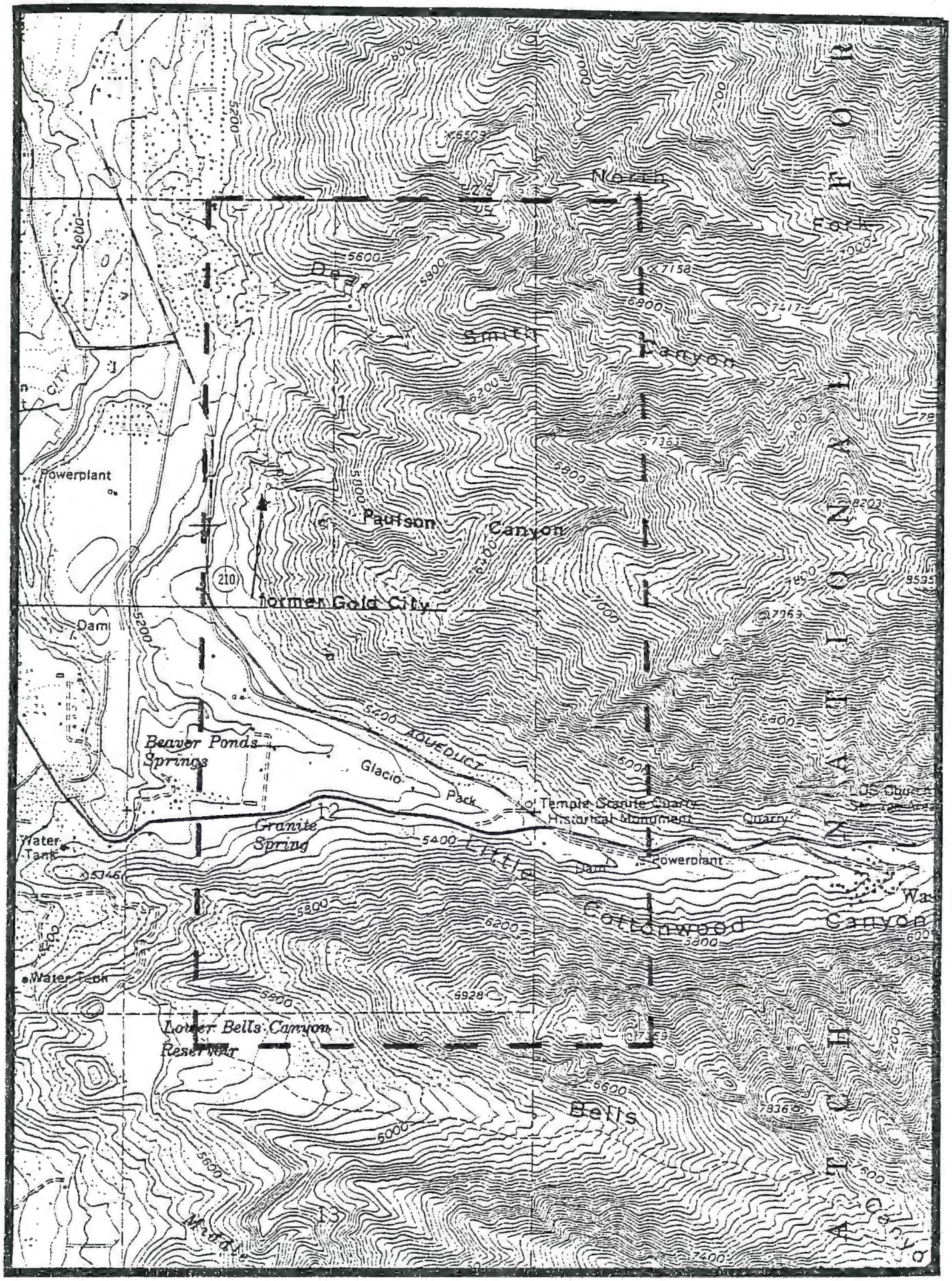


Figure 2. Topographic map of study area. (from Draper 7½ degree quadrangle map, USGS, photorevised 1975, scale: 1:24000) Dotted line represents approximate boundaries of the study area.

Figure 3. A view north from the south ridge of Little Cottonwood Canyon, along the Wasatch Fault Zone. Utah Highway 210 in foreground.

Figure 4. A view east into the study area. The Little Willow Formation is in the foreground; the Big Cottonwood forms the steeper cliffs. Thick scrub oak cover is common, especially on the north slopes, which are protected more from the sun.

first mapped the rocks along the Wasatch Front at a published scale of one inch equals four miles. During the next decades, several geologists (Walcott, 1886 and 1891; Blackwelder, 1910 and 1931; Hintze, 1913; and Schneider, 1925), discussed the rocks of the Little Willow area, but no detailed studies were attempted.

Paris (1935) completed a petrologic study of the rocks between Big and Little Cottonwood Canyons. He postulated that the highly metamorphosed schists and gneisses were the contact metamorphic equivalents of the overlying quartzites and argillites. Birch (1940) disagreed and stated that these complex rocks of the "Little Willow District" were older than the associated quartzites and argillites. He also credited the foliated texture of these rocks to a regional metamorphism, rather than to the thermal event associated with the Little Cottonwood Stock.

Crittenden, et al. (1952) mapped the rocks of this area as part of a Central Wasatch regional mapping project. It was at this time that the name "Little Willow Series" was proposed for the highly metamorphosed rocks of the area.

The glacial history of this area was described by Marsell and Jones in 1955. Determination of three ice epochs and correlation with Lake Bonneville sediments were among the results of their study.

Neff (1962) mapped the formation and carried out a petrological and structural study of the rocks. He considered that the migmatitic rocks flanking the Little Cottonwood Stock, previously mapped as Big Cottonwood Formation, more closely resembled the Little Willow rocks and hence mapped it as such.

A USGS geologic map of the Draper quadrangle, of which the Little Willow area is a part, was published in 1965 as a result of the work of Crittenden and others. This and subsequent papers by Crittenden have resulted in the following, pertinent to the Little Willow rocks:

1) change the name "Little Willow Series" to "Little Willow Formation", to include all regionally metamorphosed rocks of that area (Crittenden, 1964);

2) continued association of the questionable migmatic unit with the Big Cottonwood Formation, in opposition to Neff's (1962) work (Crittenden and others, 1965);

3) change the name of Little Willow Canyon, for which the polymetamorphic complex was named, to Deaf Smith Canyon (1975 edition of Draper 7½ degree quadrangle map);

4) a radiometric age (K-Ar method) for the Little Cottonwood Stock of 24-31 million years old (Crittenden and others, 1973).

The most recent work on the formation is that of James (1979), which delves into the mineralization and mining history of the Big Cottonwood Mining District, which includes the Little Willow Formation. Also included was a remapping of the Little Willow Formation which closely parallels the work of Crittenden (1965).

MINING HISTORY

The mining history of the area is discussed at length by James (1979) from which this section is taken.

The Little Willow Formation had been known to contain ore minerals since the early 1870's with the development of the Big Cottonwood Mining District. Mormon authorities discouraged prospecting, so it was not until the mid 1890's, when gold was discovered in Paulson Canyon, that interest was renewed. The town of Gold City grew up at the mouth of Paulson Canyon in response to the rush that ensued. Several prospects were started but no major ore bodies were discovered. Any ore that was mined was hand-picked vein material and high-graded. For instance, in 1900 one sample reportedly assayed at $4,516 per ton in gold, 46 ounces of silver and four per cent copper, as reported in the Salt Lake Mining Review.

Three types of metaliferous mineralization have been recognized in the Little Willow Formation: 1) low-grade non-economic stratabound concentrations of gold and copper in the schistose rocks, 2) bull quartz veins from which all precious metal production was mined, and 3) other quartz veins containing scheelite and huebnerite, locally developed but never mined.

Today, all that remains of mining efforts in the area are some dumps and a few shallow adits.

It is believed that most of the activity in this area was promotional and that no ore was mined profitably. The limited outcrop of the Little Willow Formation, the impingement of residential areas, and the use of this area as a watershed practically eliminates this formation as a host for an economic mining development in the future.

METHODS OF STUDY

Fieldwork for this study was completed in July, 1979. Efforts were concentrated on three main objectives: 1) a review of the previous mapping, with special attention as to the nature of the questionable migmatitic unit, 2) gathering of further structural data to help delineate major contacts, characterize major units and to elucidate the relationship of structural development to metamorphism, and 3) careful and thorough sampling of the rocks with a view to the interpretation of the polymetamorphism.

Airphotos and the USGS topographic map (1975), both at the scale of 1:12,000, provided the base for mapping. Determination of the internal constitution of the Formation, by detailed pace and compass mapping, resulted in only moderate success due to limited exposure and the laterally discontinuous nature of the units. General lithologic trends were, however, traceable across the area and found to agree quite closely with the previous work.

180 samples were acquired for petrologic study. These include representative samples from each of the lithologies and extensive sampling of the pelitic schists and stock contact rocks, both of which contain much evidence for the metamorphic history.

Structural data were taken in as uniform a pattern as outcrop permitted. These data were plotted in equal area projections and map form.

All rocks were stained for calcium, using Amarand, and for potassium, using sodium cobaltnitrite.

Study of mineral assemblages and textures of 195 thin sections was completed to adequately describe the units, as well as to acquire the pertinent data from which to analyze the metamorphic history.

CENTRAL WASATCH GEOLOGY

The central Wasatch Mountains, located at the intersection of the Uinta and Wasatch tectonic trends, contain rocks which span two billion years of earth history.

The Little Willow Formation itself represents some of the oldest rocks in the area, the exact age being obscured by subsequent metamorphic events. Schneider (1925) correlated the Formation with the Farmington Complex, amphibolite facies rocks believed to be of sedimentary origin, located northeast of Salt Lake City and now known to be at least 1.6-1.8 billion years old. Crittenden (1977) believes the Little Willow rocks correlate better with 2.3 billion-year-old Red Creek Quartzite, of the Uinta Mountains, described by Hansen (1965). The USGS groups all three units as Precambrian X in age.

Unconformably overlying the Little Willow Formation is the Precambrian Big Cottonwood Formation (Crittenden, 1965). This 16,000-foot-thick sequence of quartzites and argillites was deposited in a shallow water-mudflat environment, as evidenced by basal and mudchip conglomerates, ripples and mudcracks. The fact that the Big Cottonwood is virtually free from regional metamorphic effects, unlike the Little Willow Formation, is used as a criterion to distinguish the units.

Also of Precambrian age, but missing the actual study area (Fig. 5), is the Mineral Fork Tillite, a glaciomarine sequence (Ojakangas and Matsch, 1976 and 1980 in print). This formation is part of a belt of similar distinctive dark diamictites and related rocks, stretching from Alaska to Southern Nevada, which has been extremely useful in correlating Precambrian rocks of northwestern North America.

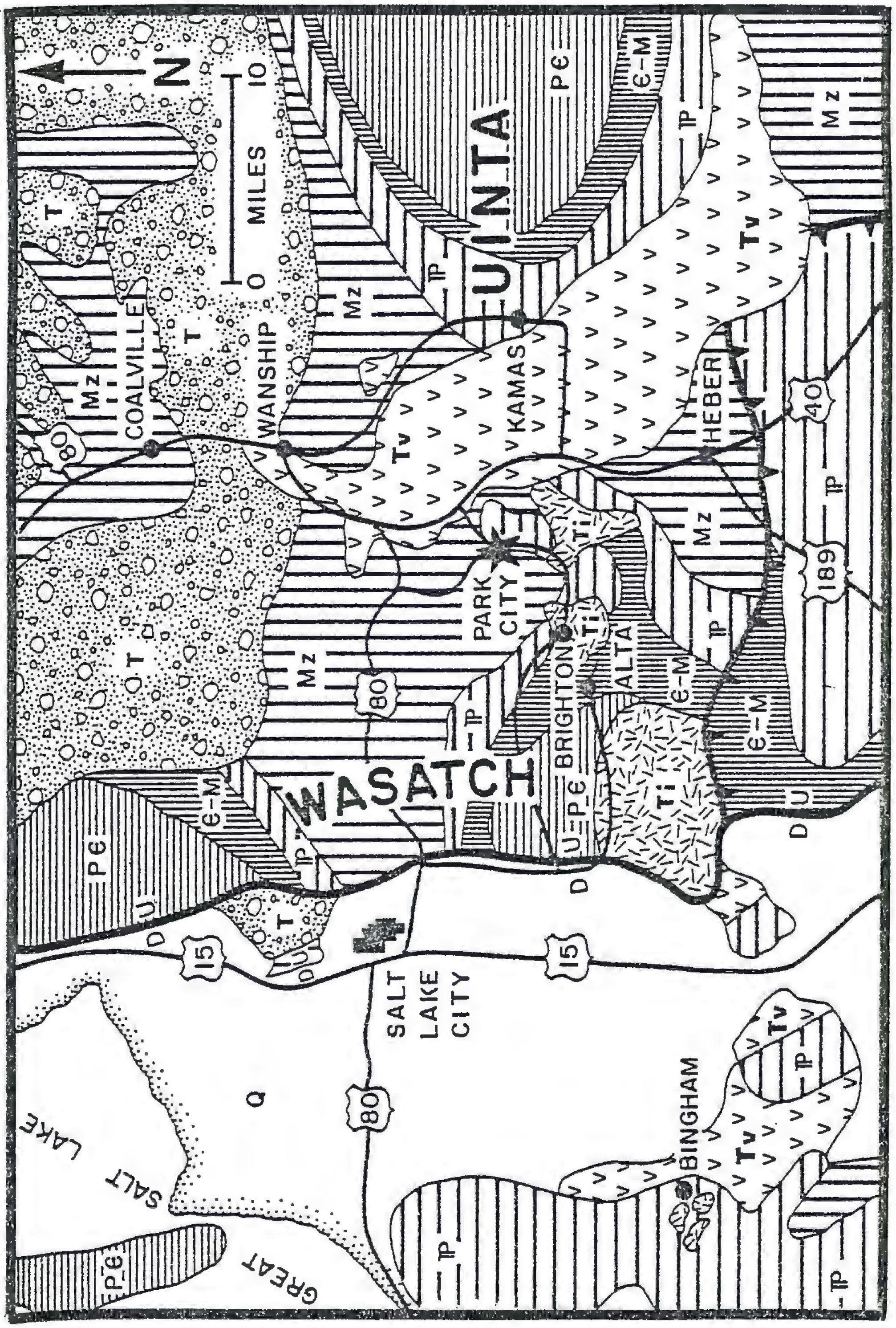


Figure 5: Generalized geologic map of the Central Wasatch Mountains. PЄ:Precambrian, ЄM:Cambrian-Mississippian, ℙ:Penn-Permian, Mz:Mesozoic, T:Tertiary (Tv:volcanic rocks, Ti:intrusive rocks) and Q:Quaternary.

This formation unconformably overlies the Big Cottonwood and is in turn unconformably overlain by the Mutual Formation, a late Precambrian shale-quartzite unit.

Recent Rb/Sr whole-rock age determinations confirm a similar age of Precambrian Y (800-1700 million years old) for the Big Cottonwood Formation, the Uinta Mountain Group, and the Belt-Purcell Supergroup of Montana and British Columbia (Crittenden and Peterman, 1975). Additional spatial and lithological similarities has prompted correlation of these units. The Mutual Formation and the Windermere Group, both "post-diamictite" in age are also regarded as equivalents (Crittenden and Peterman, 1975).

The base of the Paleozoic in the Central Wasatch Mountains is represented by a well-defined unconformity. Paleozoic rocks in the area are those of a geosynclinal environment, both transgressive and regressive in nature. A "hingeline" in the vicinity of the present Wasatch Front separated a rapidly subsiding trough of thick sediment accumulation on the west from an eastern stable shelf-shallow water environment (Crittenden, 1964). Deposition of phosphatic rock (Park City Formation) in the Permian indicates crustal stability during that time.

Brilliant red rocks seen in the Central Wasatch Mountains exemplify subaerial erosion and nonmarine deposition typical of Mesozoic time. Red shales, sandstones and conglomerates are the common rock types, with minor Jurassic marine limestone.

Late Cretaceous time saw the development of the Cottonwood Arch, the western extension of the Uinta Uplift. The axis of this anticline passes just north of Little Cottonwood Canyon, and the only recognized exposure of the Little Willow Formation forms the core of this east-west structure. An equally large, related syncline crosses the Wasatch Range near Parley's Canyon to the north (Fig. 5).

The Laramide Orogeny reached full force in the Central Wasatch area during latest Cretaceous and earliest Tertiary time, as evidenced by huge easterly-directed thrust faults. It has been theorized that the Cottonwood Arch served as a buttress, against which the thursts impinged (Eardley, 1968). The actual mechanics of the movement are still questioned; some postulate a great decollement (Roberts, 1965), others postulate gravity sliding (Eardley, 1968). The unconformity between the Little Willow Formation and the Big Cottonwood Series was apparently the locus of some of this thrust movement. Following the thrusting, the area was eroded to a peneplane.

The period from forty to twenty million years ago found the Central Wasatch involved in extensive igneous activity. Of major significance was the development of the Uinta-Cortez Uplift, an easterly-trending belt of intrusive and volcanic rocks which extends across the Wasatch range. The intrusions are of dioritic to quartz monzonitic compositions. Economically, the most important intrusions in this belt are the sub-volcanic porphyries associated with the Park City ores and the Bingham porphyry copper deposits. Of particular interest in this study is the Little Cottonwood quartz monzonite stock which abuts the Little Willow Formation and has effected thermal metamorphism on it and surrounding lithologies.

Formation of the present Wasatch topography began approximately twenty million years ago, coinciding with the advent of normal faulting and basin and range development. In general, the uplifted blocks tilt to the east and sediment eroded from these blocks fill in the adjacent basins. The Wasatch Fault Zone, with displacements as much as 15,000 feet (Crittenden, 1964), is the major normal fault zone of the area and is responsible for the uplift and exposure of the Little Willow Formation. The highest points within the Central Wasatch Mountains occur along the Wasatch Front, whereas the watershed divide is at a considerably lower altitude, generally ten to twelve miles east of the fault zone. This major fault system is still active and recent movement is evidenced by development of scarps in unconsolidated morainal and pluvial gravels.

Pleistocene glaciation has also left its mark on the Central Wasatch Mountains. Evidence for three epochs of ice advance is discernible (Marsell and Jones, 1955). The heads of most streams were occupied by ice and some of the larger canyons contained valley glaciers which modified valley shape and deposited large quantities of morainal and outwash material.

Associated with the glacial deposits in the Jordan Valley are interfingering deposits of the associated pluvial lakes. Little is known of the older Lake Alpine. Lake Bonneville is thought to have followed the last glacial advance. It exited at several levels, as evidenced by several terraces developed along the foothills. Spillover, followed by evaporation, shrank Lake Bonneville to the present day Great Salt Lake.

THE GEOLOGY AND PETROLOGY OF THE LITTLE WILLOW AREA

GENERAL STATEMENT

The Little Willow Formation is essentially a steeply dipping, north-northeasterly trending Precambrian metamorphic complex composed of schists, gneisses, amphibolites and migmatites (Plate 1). To the north and east, the Little Willow is unconformably overlain by the Precambrian Big Cottonwood Formation, of quartzites and argillites, which in general exhibits a north-northwesterly strike and moderate dip (Fig. 6). The southeastern border of the Little Willow Formation is the intrusive contact with the Tertiary Little Cottonwood quartz monzonite stock (Fig. 7). Unconsolidated Quaternary glacial and alluvial deposits form the remaining boundaries of the Little Willow, and apparently cover it on the floor of Little Cottonwood Canyon.

THE LITTLE WILLOW FORMATION

Four major lithologies are discernible within the Little Willow Formation. The most abundant rock in the area surrounding Deaf Smith Canyon is a quartzofeldspathic gneiss. Amphibolites and pelitic mica schists are found within this gneiss terrain. At the mouth of Little Cottonwood Canyon, adjacent to the Little Cottonwood Stock, a migmatitic unit is located beneath the Big Cottonwood Formation.

Quartzofeldspathic Gneiss

The majority of outcrops in the vicinity of Deaf Smith Canyon are of grey, fine-grained, slightly foliated, quartzofeldspathic gneiss, with varying amounts of micas and other minerals. This rock in the past has been referred to as a schist, but because of the poor schis-

Figure 6. A view north at the Little Willow (lower left on middle ridge) and Big Cottonwood (upper right) thrust contact. (See also Figure 4.)

Figure 7. A view south from site 2 at Little Willow (right) and Little Cottonwood Stock contact on south ridge of Little Cottonwood Canyon.

tosity, the term gneiss has been utilized for this study. Williams, Turner and Gilbert (1954) define gneiss as "rock in which the schistosity is rather poorly defined because of the preponderance of quartz and feldspar over micaceous minerals."

At first glance, outcrops of this rock appear to be of a uniform grey to tan color, but closer inspection reveals a dark mottled aggregation of micaceous minerals in an off-white quartz and feldspar matrix (Fig. 8). Darker lenses, which are richer in biotite, are characteristic and vary in length from several centimeters to a few tens of meters. Locally, limonite staining on the weathered surface gives the rock a tan to medium brown cast. Chlorite, a common retrograde product of biotite, gives the rock a greenish hue in some locations, especially in more brecciated zones.

Except for the biotite-rich lenses, which seem to be an original compositional feature, segregation layering is not evident. Development of foliation depends on the abundance of biotite, the mica-rich layers being somewhat schistose. These lenses may in fact be smaller scale equivalents of the mica schist units.

Quartz eyes were observed in some of the rocks and in some cases appeared to be stretched. Pyrite grains were seen in a few samples.

These rocks exhibit a granoblastic texture. Many of the rocks have a cataclastic fabric, especially those found immediately below the Little Willow-Big Cottonwood thrust contact and those flanking the Wasatch Fault Zone. This cataclastic fabric is characterized by fine-grained material surrounding 0.1-1 cm feldspar porphyroclasts, and by microfaulting.

Figure 8. Typical hand samples of quartzofeldspathic gneiss. Left - cut sample from site 7; note lighter colored portion and darker colored biotite rich portion. Right - weathered brecciated sample from site 24.

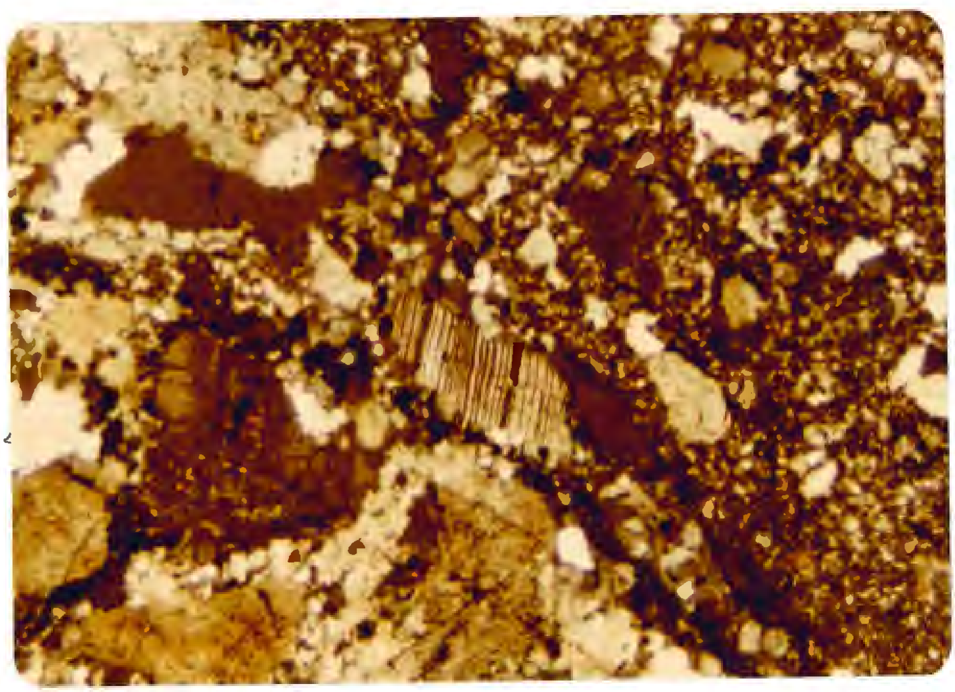


Figure 9. Mylonite (right) and mylonitic gneiss contact in quartzo-feldspathic gneiss, sample 24-2, width of field - 2.4mm., crossed polars.

Recrystallization is quite advanced, in most cases exceeding the degree of cataclasis. Therefore, according to Higgins' (1971) classification, these specimens would be termed mylonite gneiss. One sample shows, in addition to these features, an overprinting by a post-recrystallization cataclasis, forming zones of mylonite (fluxion structure, cataclasis greater than recrystallization) within the mylonite gneiss (Fig. 9).

Mineralogically, the quartzofeldspathic gneiss is composed of plagioclase (60-80%); K-feldspar (0-8%); quartz (10-30%); biotite (0-5% typical, as high as 30% in the mica-rich pods); muscovite (0-8%); and chlorite (2-10%). Accessory minerals include opaques (mainly magnetite and pyrite, 1-4%), zircon, apatite, and epidote.

Feldspar, the most common constituent of the gneiss, is almost totally plagioclase, much of which shows albite twinning. Grid-twinned microcline is found in a few samples, in amounts up to 8%. Staining of thin section heels verifies this preponderance of plagioclase over K-feldspar. The majority of plagioclase occurs as porphyroclasts. Sutured grain boundaries and development of mortar texture on the periphery of the porphyroclasts is not uncommon. In the more cataclastic specimens, kinked twins and fractured grains are commonplace (Fig. 10). A portion of the feldspar occurs as finer grained groundmass between the porphyroclasts. This material is usually well recrystallized and less altered. The degree of alteration may in part be related to the degree of brecciation, although all feldspars show some alteration. Both sericite and saussurite are common secondary products.

Quartz mainly occurs as small recrystallized pods and stringers within the groundmass of the rock (Fig. 11). Most of the grains show

Figure 10. Fractured plagioclase porphyroclast in quartzofeldspathic gneiss from near site 13, sample 10-Neff, width of field - 2.4mm., crossed polars.

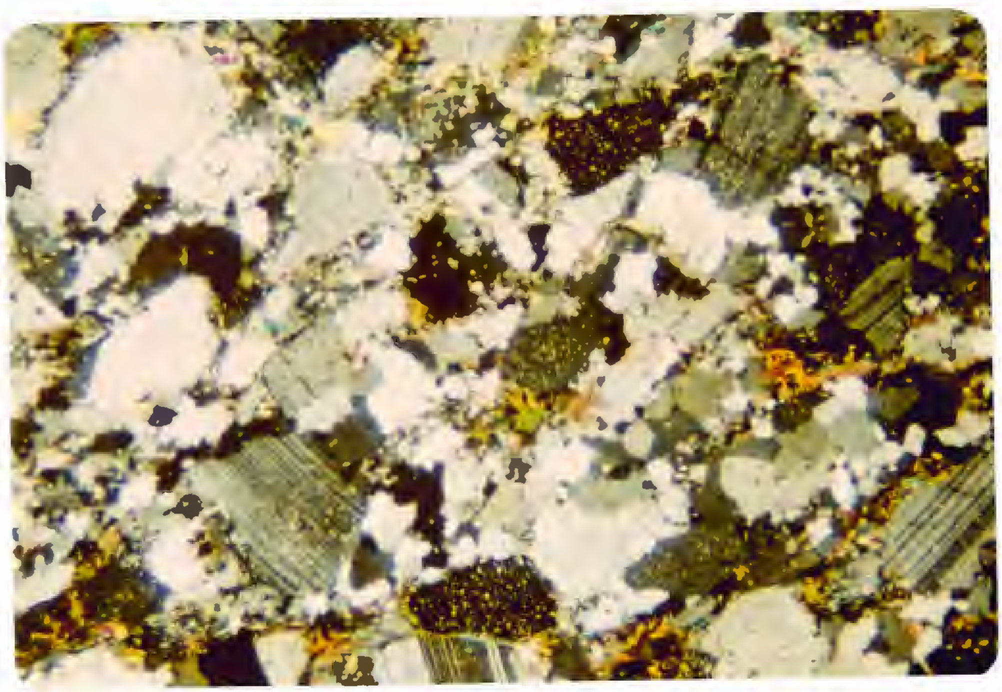


Figure 11. Typical quartzofeldspathic gneiss, note quartz stringers, feldspar porphyroclasts and well developed mortar texture, sample 15-20, width of field - 6mm., crossed polars.

little to no strain and somewhat simple, yet sutured, grain boundaries. Quartz also occurs as fine grained recrystallized groundmass and occasionally as inclusions in feldspar.

Green-brown pleochroic biotite is the most common mica. Amounts are quite variable, probably an effect of the original composition. Some of the flakes show a crude dimensional preferred orientation and in some cases they wrap porphyroclasts and quartz pods. Other biotites are secondary recrystallized grains which show no apparent relationship to foliation or other structures within the rock. Muscovite occurs much like biotite in amount and habit.

Chlorite is a very common secondary alteration mineral in the gneiss. It occurs as a replacement of biotite, in fractures and associated with the opaques.

Overgrowths on rounded zircons were seen in two samples (Fig. 12).

The assemblage quartz+plagioclase+biotite+muscovite±K-feldspar appears to have been fairly stable. The only changes reflected in the rock are the two cataclastic events, which were separated by the thermal recrystallization. Chlorite, saussurite and sericite are typical alteration products. The opaques may have been original, possibly remobilized during the metamorphism, or may have been added at the time of the stock intrusion. Zircon and apatite seem to be relict grains whose round shapes suggest a sedimentary origin.

Amphibolite

Within the gneiss terrain can be found dark green, fine to medium grained amphibolite. This rock crops out in somewhat lenticular units which vary in size from small (2x10 m) lenses on the northwest side of

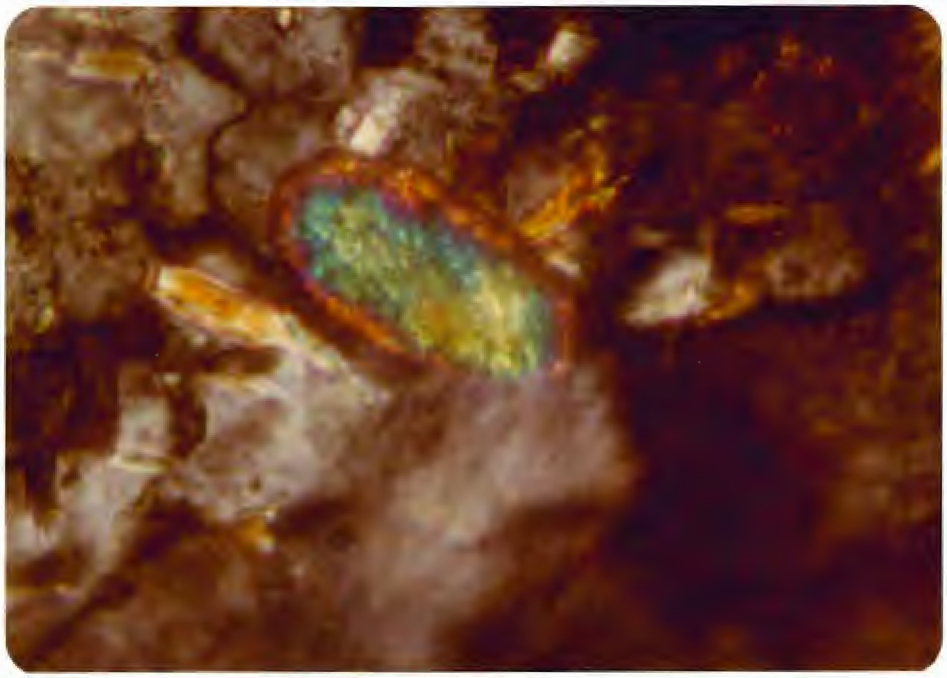


Figure 12. Overgrowths on rounded zircon in quartzofeldspathic gneiss, sample 7-3, width of field - 0.15mm., crossed polars.

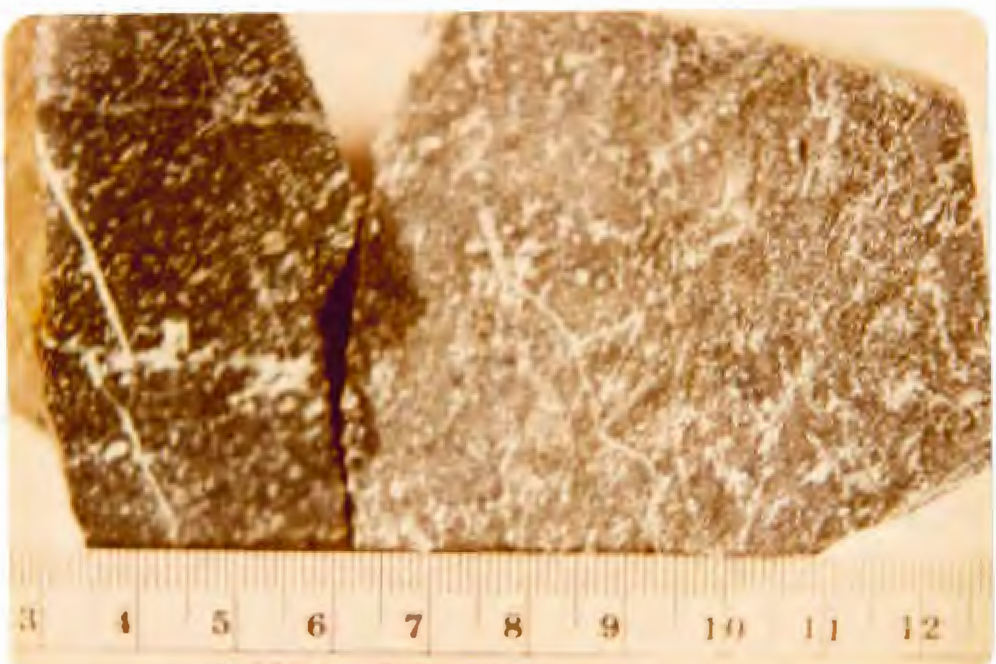


Figure 13. Typical hand samples of amphibolite; left - cut sample from site 41, right - sample from site 3 showing foliation surface.

Little Cottonwood Canyon, to a 60-250 m wide unit which can be traced across the entire Little Willow outcrop area. It is not known whether the smaller exposures represent the maximum extent of the amphibolite, or if they are continuous under the overburden. Actual outcrop can be quite rare, but the unit can be traced by means of float and the characteristic brown soil developed above this rock type.

The amphibolite is readily visible in the field, its dark color contrasting with the light tan of the quartzofeldspathic gneiss. The rock is dark brown on the weathered surface and black to dark green on the fresh surface (Fig. 13), the greener varieties being more chloritic.

The actual contact of this unit with the quartzofeldspathic gneiss is rarely observed, but where seen is usually quite sharp. Locally, the two units interfinger over a 4-6 m zone. Generally the grain size of the amphibolite is finer at the contact.

Compositional layering is not evident in this unit. The rock is moderately foliated and, in places, a crude mineral lineation is developed. Some of the rocks have a streaked, spotty appearance due to irregular 4-8 mm segregations of quartz and feldspar. A felty texture is apparent in a few of the samples.

Jointing is common and conjugate veins, filled with plagioclase, epidote and quartz can be found (Fig. 57).

The amphibolite consists primarily of hornblende (70-90%) and plagioclase (10-15% common, as high as 25%). Minor constituents include quartz (0-5%), epidote (2-15%), biotite (0-5%), and chlorite (2-3%). Accessory minerals are magnetite, illmenite, pyrite, sphene, leucoxene, apatite, and garnet.

The hornblende in the amphibolite is blue-green, recrystallized, and exhibits nematoblastic texture, with development of crude dimensional preferred orientation in some instances (Fig. 14). The mineral occurs in two habits: as larger (up to 5 mm) euhedral grains and as fine-grained amphibole in the matrix. The pleochroism scheme is α pale brown, β medium green, and γ blue-green; the maximum extinction angle (γ to c) is 15°, and the birefringence is 0.023.

Plagioclase occurs as fine-grained matrix material as well as larger grains, a few of which exhibit albite twinning. Mortar texture can be seen locally; polygonization, primary recrystallization, and complex grain boundaries are fairly common. Inclusions of quartz within the larger grains are not uncommon. In the majority of cases, the relief of the plagioclase is slightly greater than that of the quartz (Becke line method), so an anorthite content of greater than 20% is suspected (Deer, Howie and Zussman, 1966, p. 327). Both sericite and saussurite alteration of plagioclase are common, except in some of the finer-grained recrystallized portions. Stained thin-section heels revealed no K-feldspar.

Quartz is primarily restricted to secondary veins, although some does occur as inclusions in feldspar and as recrystallized aggregates in the matrix (2% maximum). Both strained and recrystallized grains are present.

Epidote is most common as a secondary vein mineral, although as much as 15% is present in the host rock, in the form of fine-grained matrix material. As a vein mineral, it is layeritic in habit, with quartz filling in between the euhedral grains.

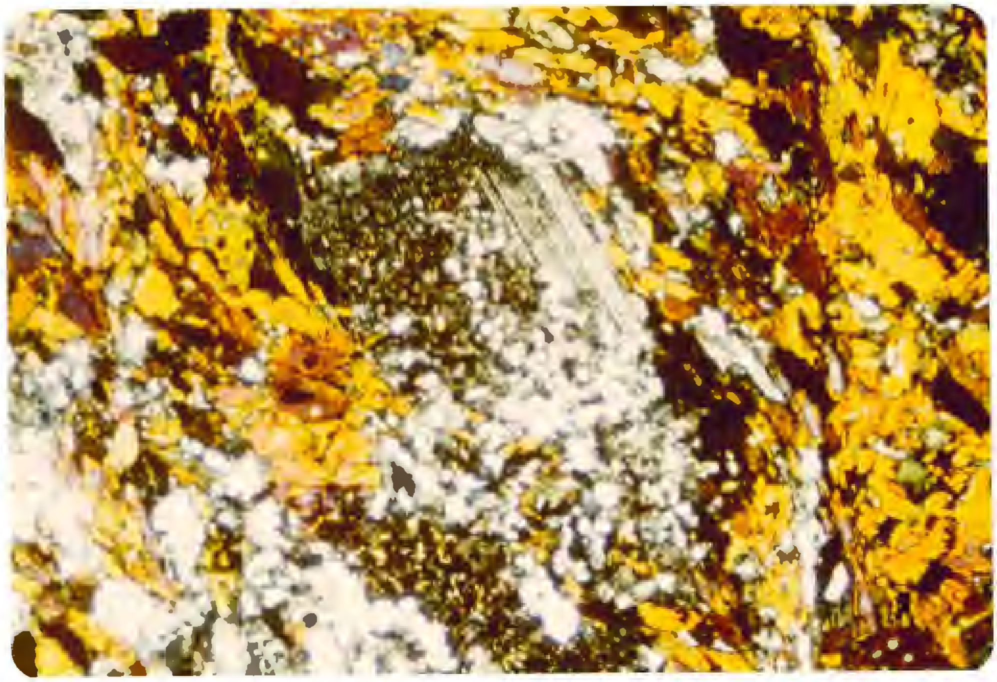


Figure 14. Typical amphibolite with plagioclase porphyroclast (?), sample 41-1, width of field - 2.4mm, crossed polars.

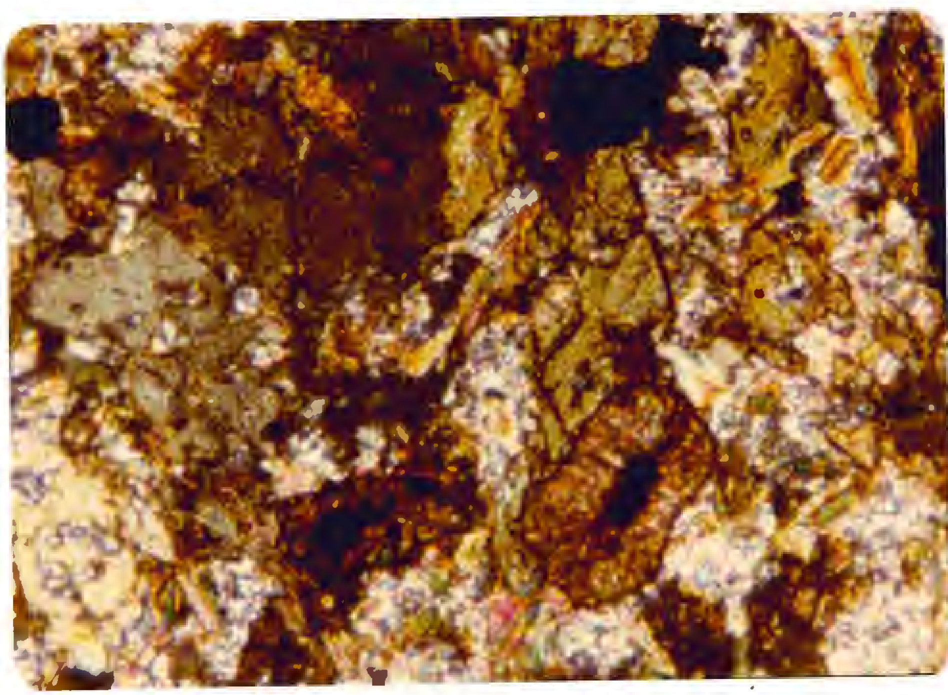


Figure 15. Sphene surrounding opaque, probably ilmenite, in amphibolite, sample 24-3, width of field - 0.6mm, crossed polars.

Most of the biotite present is secondary as evidenced by flakes oriented roughly perpendicular to the foliation. A common habit for biotite is in small decussate (2-5 mm) pods and clots associated with opaques.

Chlorite is a common secondary mineral in the amphibolite. In places it can be seen replacing biotite. Most typically it is associated with the opaques.

The opaques in some cases make up to 10% of certain layers, mainly as euhedral grains. Sphene is typically found surrounding opaques, probably ilmenite (Fig. 15). Garnet occurs occasionally as small subhedral to euhedral grains in the amphibolite.

Pertinent mineral assemblages in this unit include:

hornblende-plagioclase±quartz

hornblende-plagioclase-biotite±quartz

hornblende-plagioclase-epidote±quartz±biotite

hornblende-plagioclase-garnet±quartz

Chlorite, leucoxene and sphene appear to be secondary alteration minerals.

Pelitic Mica Schists

Also within the quartzofeldspathic gneiss terrain are mica schists, most of which are rich in aluminum. The units range in size from small (0.3-2 m) mica-rich pods (described as part of the quartzofeldspathic gneiss) to roughly tabular units which crop out in northeasterly trends across the Little Willow area.

Wherever observed, these schists appear to be conformable with the quartzofeldspathic gneiss. Foliations, lineations, and most structures are consistent throughout both lithologies. The contact is in places sharp, and in other locations gradational with one rock type becoming gradually more dominant as the other diminishes. Indeed, typical quartzofeldspathic gneiss lenses are not uncommon in the schists. In mapping, those areas which are predominantly schist were distinguished from areas of predominantly quartzofeldspathic gneiss (Plate 1).

Actual outcrop of the schist can be quite variable. In places the schists form the major portions of a cliff and the best exposures are on vertical faces. Perhaps more typical are small (0.5-10 m wide) outcrops disbursed along a brush and rubble covered surface. These can be traced to some extent with reasonable certainty of their continuity.

The rock is variable in color, including black, grey, green, tan and silver, depending on the principal mica in the rock.

The schists are medium-grained with the alternation of felsic and micaceous layers, often evident even in hand sample. An outstanding feature of many of these rocks is the growth of large andalusite and cordierite porphyroblasts. These crystals are commonly more resistant to weathering and stand out on the weathered surface (Fig. 16).

Foliation due to higher abundance of micas than in the quartzofeldspar gneiss is always well developed. The majority of the schists exhibit a prominent crenulation of the primary foliation (S_1), giving rise to an easily measurable lineation. A lepidoblastic texture is common, with subparallel micas apparently controlling the shapes of less anisotropic minerals. Porphyroblasts are both euhedral and anhedral, usually with sieve texture. Microfaults parallel to crenulation axial surfaces were noticed locally, usually rehealed or with recrystallized minerals.

Figure 16. Weather-resistant porphyroblasts in mica schist:

A : Cordierite, from site 6.

B : Andalusite, from site 21.

The majority of schistose rocks comprising this unit are somewhat pelitic and rich in micas, but otherwise monotonous. However, contained within these churlish rocks are some tantalizing variations of the pelitic mica schist. A distinctive sequence of garnet-cordierite schist, andalusite-mica schist, stretched-pebble metaconglomerate, and quartz-sericite schist was observed at a number of sites (sites 1, 12, 15, 22 and 44). The sequence could not be traced across the entire area, and seems to be a series of lenticular units, of variable thickness, never exceeding about two meters. No stratigraphic top or bottom indicators were recognized, but the sequence is fairly consistent at the various locations even though the units appear to be lenticular and interfingering. Contacts between these units appear to be conformable and usually sharp. At some locations, quartzofeldspathic gneiss bounds the sequence and can be found interlayered with it.

A most conspicuous unit is the stretched-pebble conglomerate. The recrystallized pebbles, which vary in size from 1.5 to 15 cm, are isolated and often stretched or flattened (Fig. 17). The majority of the pebbles are composed of quartz but one was found with quartz and feldspar and a graphic texture. Foliation wraps the pebbles. The original pebble outline can often be discerned. The matrix has a high proportion of micas, usually muscovite with minor biotite. Recrystallized quartz stringers in the matrix are also common.

This apparent metaconglomerate serves as an excellent marker unit, as well as lending further evidence of the sedimentary origin of these rocks.

Locally within the mica schist unit, some rocks originally thought to be metaconglomerate were determined to be quartz-mica $\pm$ andalusite rock containing stretched and boudinaged quartz veins (Fig. 18). Three lines of evidence were used to distinguish these pseudo-conglomerates: 1) even though stretched, these pods lacked the "rounded" nature common in the true pebbles; 2) several pods, lying on one horizon, could roughly be fitted back together; and 3) when the rocks were cut to see the third dimension, several pods were found to constitute a continuous sheet of quartz.

A distinctive andalusite-mica schist, with large (6 cm long) andalusite crystals, is found immediately adjacent to the conglomeratic unit (Fig. 19). Andalusite makes up as much as 35% of the rock as euhedral porphyroblasts. The foliated mica matrix is primarily biotite with minor muscovite. Recrystallized biotite surrounding the andalusite is very common (Figs. 19 and 24). Locally, samples are highly chloritized.

A garnet-cordierite-chlorite rock, associated with the above sequence, is found certainly at site 15. A pock-marked weathered surface is characteristic of this rock (Fig. 20). Visible chalcopyrite is particularly evident in some samples. Except for remnant folia in the garnet, foliation is not evident in this rock, suggesting late, post-deformational crystallization.

Euhedral to subhedral almandine garnets, some with S-shaped opaque inclusion patterns (Fig. 27) make up to 30% of the rock, with sector twinned cordierite (Fig. 25) as high as 35%. Chlorite is closely associated with the cordierite. High concentrations of zircon and tourmaline,

Figure 17. Stretched pebble conglomerate, within mica schist unit:
A. Field picture from site 11, note andalusite crystals.
B. Hand sample picture, sample 44-2.

Figure 18. Stretched and boudinaged quartz veins:
A. Field picture from site 11.
B. Hand sample 11-3, from site 11; the rock on the right is the cross section of the rock on the left. Although the pods appear to be discrete they are found to be a continuous layer of quartz in the third dimension.

Figure 19. Andalusite - biotite schist:

A. Hand samples: left, polished sample 15-3; middle, cut sample from site 22; right, uncut sample 15-6.

B. Closeup of float sample found near site 15; note biotite surrounding the resistant andalusite crystals.

believed to be original detrital grains, are found locally in small pods. Biotite commonly occurs in decussate clots (Fig. 22). At other locations, a similar looking rock, with pock-marked weathered surface and apparent lack of foliation, was found. The peculiar biotite clots were also seen, but no garnet or fresh cordierite was found.

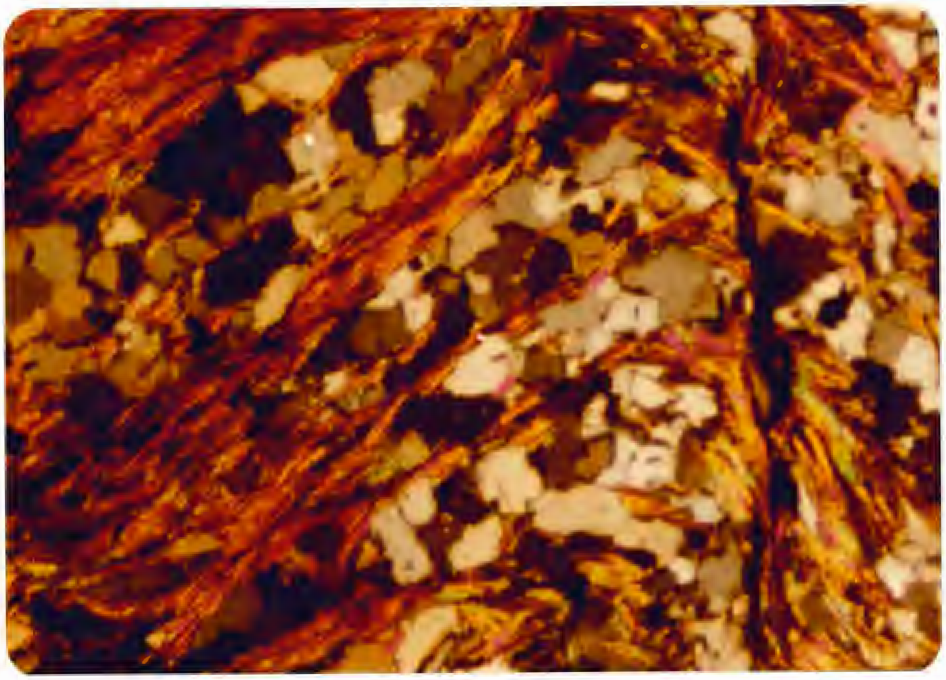
Also adjacent to the metaconglomerate is a very distinctive white quartz-sericite schist (Fig. 21). Locally this rock forms a resistant layer one to two meters thick. It is composed of 40-70% quartz as recrystallized stringers associated with 10-30% microcline grains. These layers alternate with fine-grained muscovite (20-40%). In places, this rock has well developed crenulation.

The dominant mica schist is composed of quartz (10-50%), biotite (10-45%), muscovite (10-55%), plagioclase (0-10%, usually) and rarely K-feldspar (0-3%). It is typical in rocks with very high concentrations of biotite to find very little to no muscovite, and vice versa. Andalusite is found locally and in variable amounts from 1-10%. Chlorite (1-15%) and sericite (1-40%) are common alteration products. Accessory minerals include pyrite, magnetite, chalcopyrite, hematite, zircon and tourmaline.

The most common habit for quartz is as segregated recrystallized stringers and pods. It also occurs as inclusions in andalusite and cordierite, secondary veins and as pebbles in the conglomerate. Most grains show some signs of recrystallization, although veins and pebbles, typically with larger grain size, often exhibit more strain. Sutured grain boundaries are common; polygonal ones are rare.

Figure 20. Hand sample 15-11, garnet-cordierite-chlorite-biotite rock. The rock on the left shows the pock-marked weathered surface.

Figure 21. Hand sample 15-5 of distinctive, white quartz-sericite schist. The bottom rock is the cross section of the top rock; note crenulations on the foliation surface.



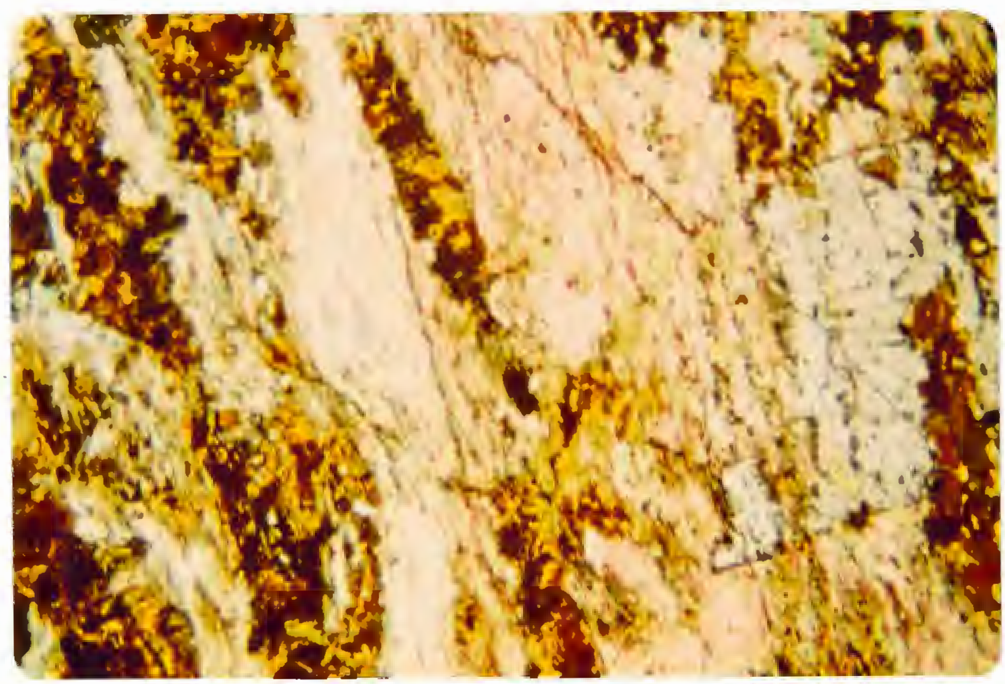


Figure 22. Biotite in the mica schists:

A. Foliated and crenulated flakes associated with muscovite; note well developed spaced schistosity. Sample 1-1, width of field 2.4mm., crossed polars.

B. Recrystallized biotite clots; also note andalusite prophyroblast and muscovite with (lattice) preferred orientation. Sample 22-1, width of field 6mm., crossed polars.

Both plagioclase and microcline were identified as minor constituents in the mica schist. The majority occur as recrystallized grains, associated with the quartz. Sutured grain boundaries are common. Microcline and albite twins are seen locally. Sericite alteration is variable from minor to moderate.

Biotite occurs in two distinct habits: 1) foliated and crenulated flakes associated with muscovite (Fig. 22), and 2) large unkinked flakes, often oriented subparallel to the crenulation axial surfaces (S_2) or rimming andalusite grains. It was also found as decussate grains in "clots" evidently replacing some other mineral (Fig. 22).

Muscovite also occurs in two orientations, as crenulated S_1 mica, exhibiting dimensional and lattice preferred orientation, and as unkinked S_2 grains, randomly to crudely dimensionally oriented. Muscovite and sericite are commonly found replacing andalusite and probably cordierite. Muscovite is found in all the rocks except the garnet-cordierite schist and some of the extremely biotite-rich rocks.

Andalusite is always found as porphyroblasts. The majority are euhedral (Fig. 23), some are subhedral and all exhibit spongy, sieve texture (poikiloblastic). Grains with 0.5-1.5 cm cross sections are typical. All andalusites studied in thin section are definitely post-S_1 because they overgrew that foliation (helicitic texture). Some thin sections display all Z-shaped or S-shaped inclusion patterns, suggesting rotation during growth, probably synchronous with crenulation development (S_2). Further evidence for grain rotation is indicated in examples showing internal foliation discontinuous with external foliation (Fig. 24). A second, post-S_2 generation of andalusite, which grew statically, is observed in some of the rocks (Fig. 23). Partial alteration to sericite and fine-grained muscovite, and rimming by recrystallized biotite is common. Andalusite is found associated with all minerals present, except garnet.

Fresh cordierite is rare except in the garnet-rich rocks where it usually exhibits sector twinning (Fig. 25). Commonly the mineral has a brownish corrosion and partial alteration associated with it (Fig. 26). Some highly sericitized megacrysts are suspected to have been cordierite. Cordierite grains tend to be subhedral to anhedral and poikiloblastic. Although not conclusive, limited evidence suggests a post-S_1 and a pre-S_2 growth for some of the cordierite, since the grains exhibit a faint internal foliation, yet appear to be wrapped by the S_2 micas. Cordierite in the garnet-cordierite rocks crystallized statically with a radial habit. Cordierite is found associated with garnet, biotite and chlorite and more rarely with andalusite and muscovite.

Garnet is found only very locally in the mica schist, but where found can be a major constituent of the rock. Grains are euhedral to subhedral, 1-5 mm in diameter, and commonly exhibit a pervasive fracturing. S-shaped inclusion patterns (Fig. 27) are present in some samples, implying rotation during growth. Since the garnet is the only mineral in the garnet-cordierite rock to show foliation, it may be an earlier, relict phase in a post-deformationally recrystallized matrix. Garnet is associated with biotite, cordierite and chlorite, but never with andalusite or muscovite.

Chlorite is a common alteration product, especially of biotite. It also occurs associated with cordierite. Sericite is a common alter-

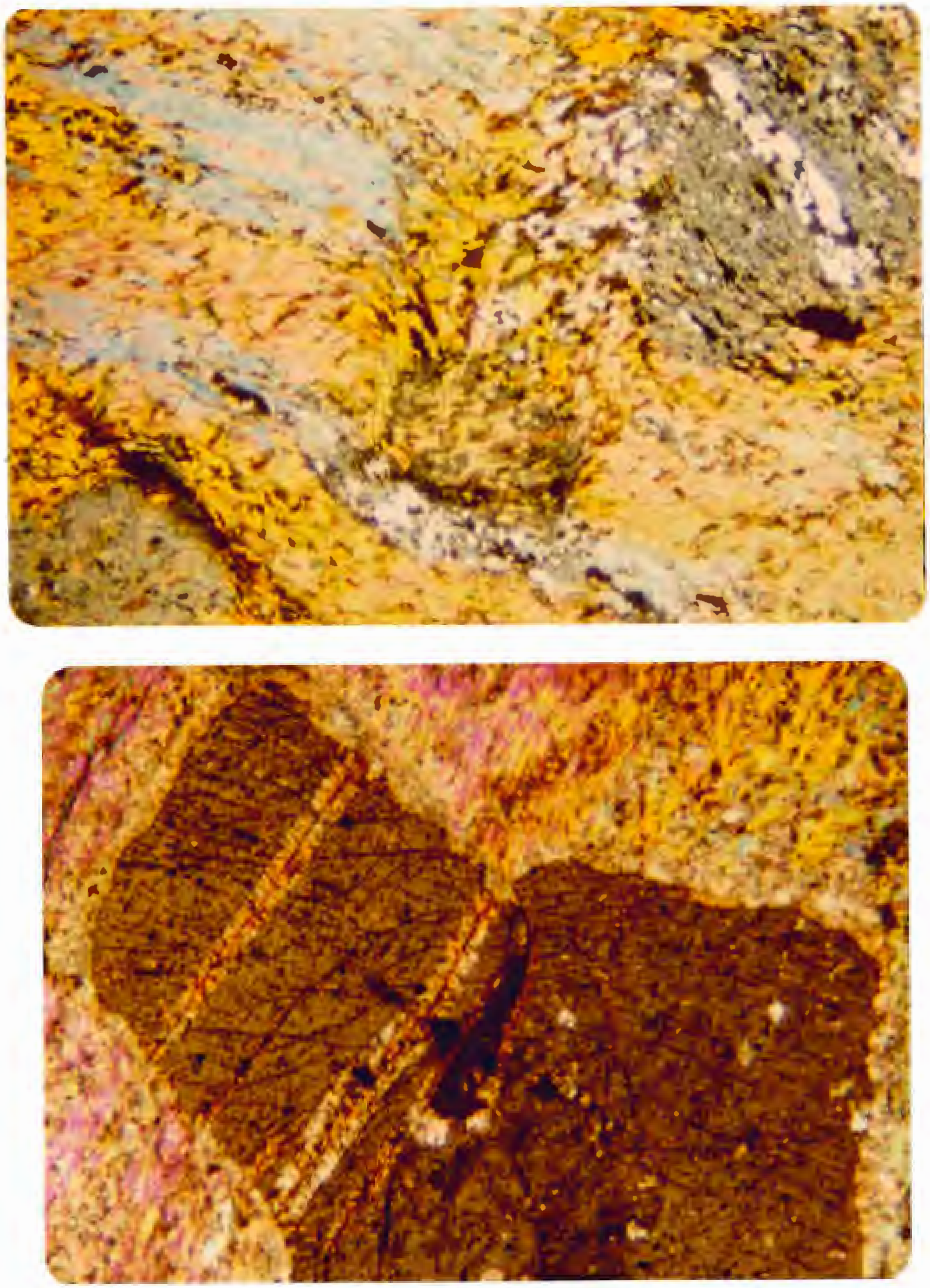


Figure 23. Andalusite in the mica schists:

A. Similar bent internal foliation patterns and inclusion of some S_2 biotite (center) suggest a post S_1 to S_2 growth, with possible rotation. Sample 15-3, width of field 6mm, crossed polars.

B. Euhedral porphyroblasts overgrowing both foliations indicate a second generation (post S_2) of andalusite. Sample 21-2, width of field 6mm, crossed polars.

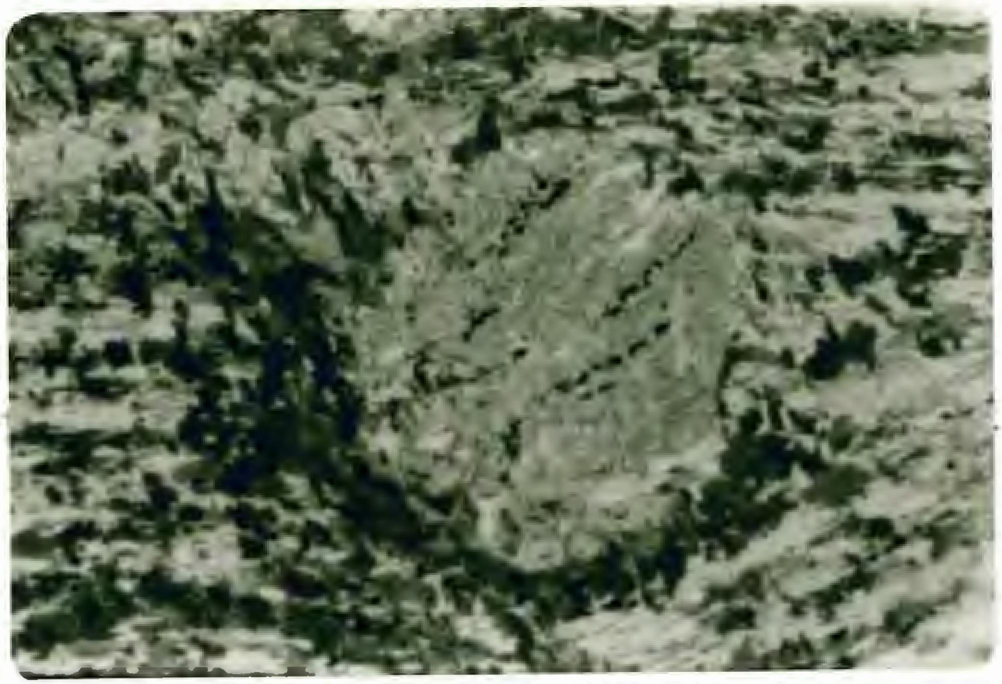


Figure 24. Rotated andalusite in the mica schist. Note discontinuous internal foliation and decussate recrystallized biotite surrounding the andalusite. Sample 4-5 (Grant), width of field 6mm., ordinary light.

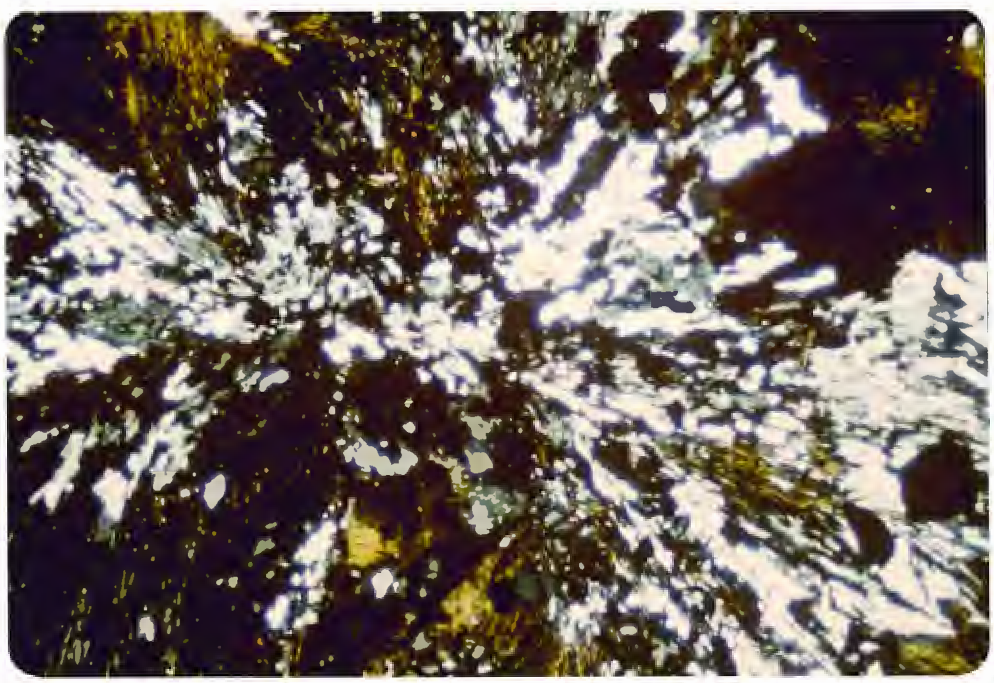


Figure 25. Sector twinned cordierite with included chlorite. Sample 15-11, width of field 2.4 mm., crossed polars.

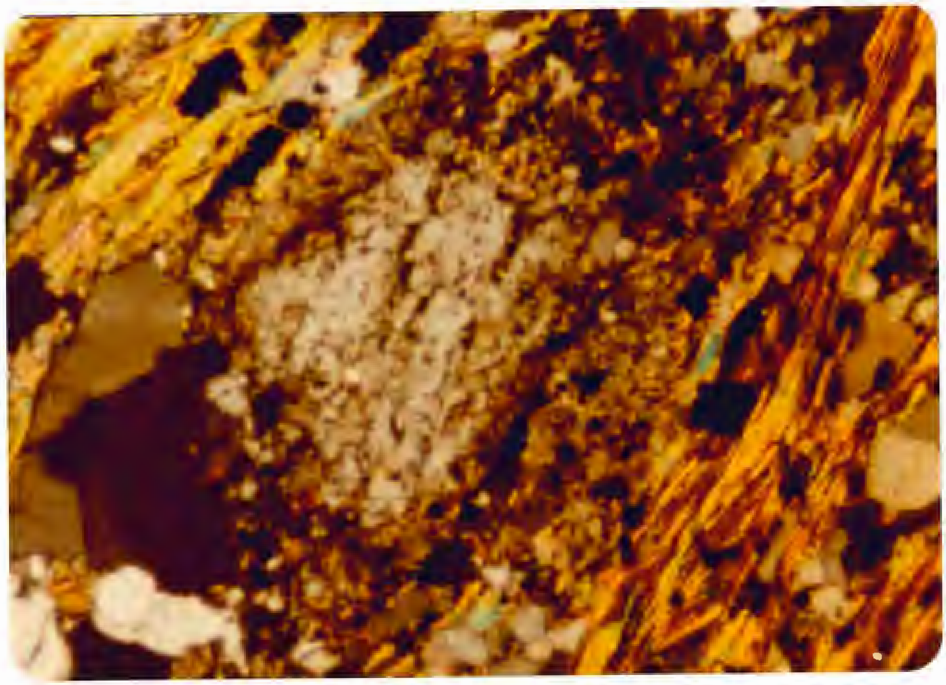


Figure 26. Partially altered cordierite with brownish corrosion; note faint included foliation, yet the crystal is wrapped by the (f_2) foliation. Sample 1-1, width of field 2.4mm., crossed polars.

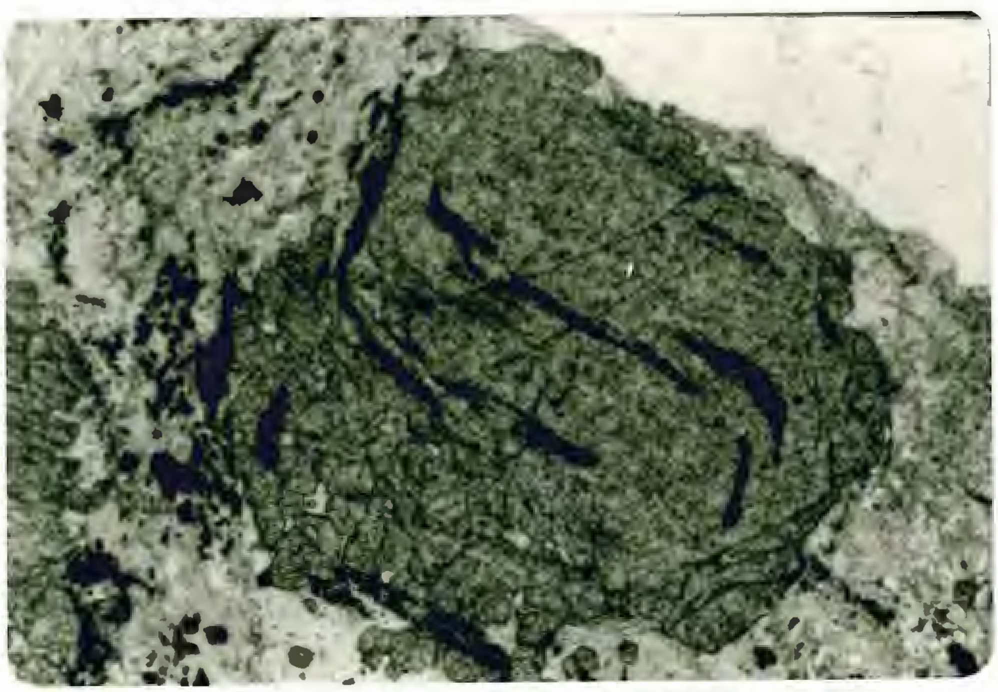


Figure 27. S-shaped opaque inclusions in garnet. Sample 15-11, width of field 6mm., ordinary light.

ation mineral of andalusite, cordierite and feldspar. Opaques tend to be euhedral grains which favor certain layers in the schists. High concentrations of small zircon and tourmaline grains were found locally in small micaceous pods and are probably detrital grains.

Mineral assemblages which are important in this unit include:

quartz-muscovite-biotite$\pm$plagioclase

quartz-muscovite-cordierite-biotite$\pm$andalusite

quartz-muscovite-cordierite-biotite$\pm$andalusite

quartz-muscovite-K-feldspar$\pm$plagioclase

quartz-biotite-garnet

garnet-cordierite-chlorite-biotite

cordierite-chlorite-biotite$\pm$muscovite

Migmatitic Gneiss

A migmatitic gneiss exists as a south-southeasterly extension of the main Little Willow body. This unit crops out mainly as steep, massive cliffs, adjacent to the Little Cottonwood Stock and beneath the Big Cottonwood thrust plate, on both sides of the mouth of Little Cottonwood Canyon.

Previously, this lithology had been mapped as Big Cottonwood Formation (Crittenden, 1965; James, 1979) and as part of the Little Willow Formation (Neff, 1962). In this study the unit was mapped separately because it could be distinguished from the other rocks, based on the following criteria:

1) the very heterogeneous nature of the unit;
2) presence of foliated sillimanite in the pelitic portions of the rock (Fig. 37);

3) presence of ductile deformation features (Fig. 30); and

4) development of localized, yet ubiquitous leucosomatic envelopes (rinds) around melanosomatic portions of the rock (Fig. 31).

Leucosome is defined as the light-colored part of a migmatite, usually rich in quartz and feldspar, and melanosome as the dark-colored portion, usually rich in mafic minerals (Mehnert, 1968).

The contact of this unit with the Little Cottonwood Stock is quite obvious (Fig. 7) and dips steeply to the west, as exposed in Little Cottonwood Canyon. The contact with the Big Cottonwood Formation is quite evident on the south side of Little Cottonwood Canyon, where it exists as a narrow thrust zone with a shallow dip to the west. On the north side of the canyon this contact is not evident, but rather inferred, being obscured by brush and talus. The remaining boundaries are with unconsolidated Quaternary deposits.

From a distance the migmatite has a medium brown color, due to weathering of iron minerals present in this rock. The effect is intensified by the contrast with the light colored intrusive and the whitish to tan Big Cottonwood quartzites. Close up, the unit is a melange of dark portions floating in a white to light grey quartz-rich matrix.

The vast majority of the migmatites are medium-grained. Numerous coarser-grained pods and veins are dispersed within the outcrops.

A traverse through the migmatite shows that variability in composition is the rule rather than the exception.

A large portion of the rock is a massive, recrystallized, grey to white quartzite. One quartzite unit was traceable on both walls of the canyon. Typically, quartzitic units are gradational into a more feldspar-rich quartzose rock. A chalky white appearance is a good first indication of this composition change.

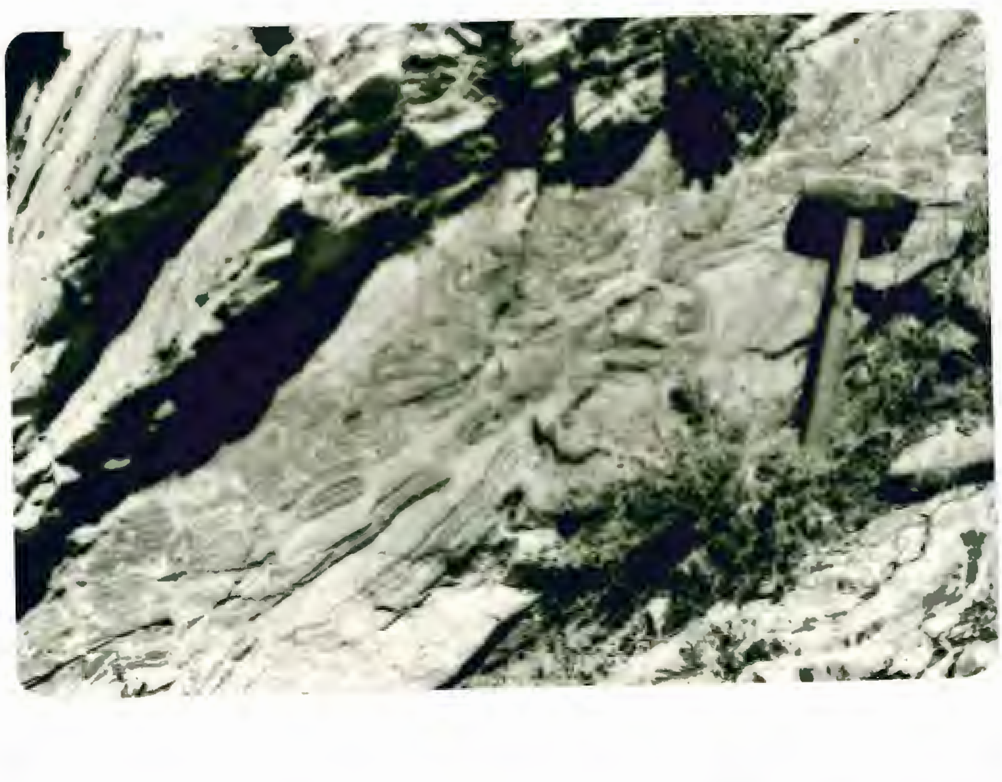
Alternation of the quartzofeldspathic, pelitic and quartzitic layers, averaging two to ten centimeters, but ranging up to one meter in thickness, is typical, interspersed between the more massive quartz-rich rocks. The contacts between the layered and massive portions tend to be gradational, whereas the individual layer contacts are usually sharp.

Folding and faulting locally obscure all original structure. Isoclinal folding was evident in several locations (Fig. 50).

A very common structure throughout much of the outcrop area are boudinaged pelitic layers within the lighter colored portions (Fig. 28). This type of migmatitic structure has been referred to as "schollen" (Mehnert, 1968). Often these brittle rafts, even though bent and rotated in a ductile matrix, appear to have once been continuous, and can be mentally reassembled into layers. The blocks were variable in size, between one centimeter and one meter in diameter, apparently dependent on the original thickness of the layer.

Westward, as the Little Willow contact is approached, the schollen structure became less apparent, grading first into boudinaged layering and then into more continuous layering (Fig. 29). This type of layering was noticed locally at a few other locations within the migmatite.

The quartzitic units appear to have deformed ductily in these rocks, instead of behaving brittley, which is more typical for this composition. A possible explanation for this is an increase in ductility due to a small amount of interstitial melt.



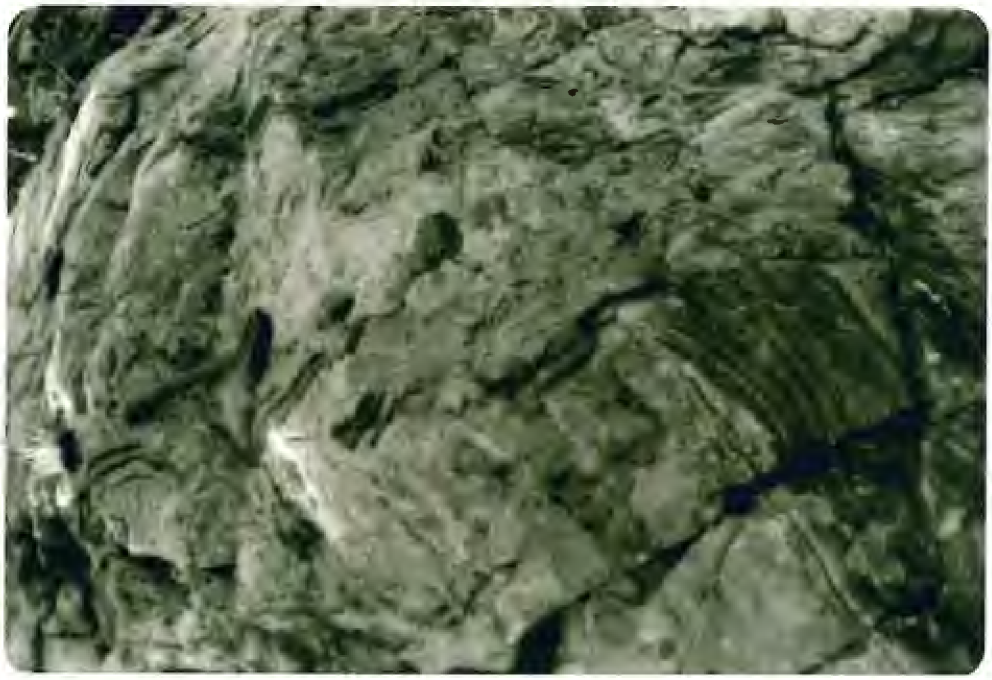


Figure 28. Structures in the migmatites:
A Left - Sheared layers found between sites 26 and 27.
B Left - Boudinaged and broken layers at site 27.

A Right - Isolated rafts near site 30.
B Right - Broken and rotated layer at site 43.

On a smaller scale, the rafts themselves exhibit variable structure. The most common are composed of simple parallel layers or are homogeneous. Although rarer, blocks with a swirly, schlieric pattern are also seen (Fig. 30). In a number of the rafts, a gradation seems to exist between the parallel and swirly patterns.

Also conspicuous in these rocks is the development of leucosomatic envelopes or rinds around dark colored portions of the rocks (Fig. 31). In the case of layers, this apparent remobilized "sweat" material is found parallel to and at the contact. In the rafts, this light-colored portion often rims the entire melanosomatic block, and usually surrounds individual pelitic areas of the block, in both schlieric and "stromatic" (layered) types. Rinds range in thickness from one to two grains to about one centimeter.

It is also common locally to find leucosomatic veinlets and pods within the more massive dark-colored portions (Fig. 32). These were found to trend both parallel and perpendicular to the foliations. Size is variable from minuscule veinlets to 15x40 cm pods. They are always coarser-grained than the sourrounding rocks and contacts are usually diffuse.

Found sporadically within the migmatite are occasional lenses and layers of pegmatite, generally light colored and with simple mineralogy. Although some pegmatitic veins near the stock appear to be derived from it, the majority of the pegmatites exhibit evidence which does not support such a source. This includes:

1) the majority of pegmatites are surrounded on all visible sides by migmatitic country rock, i.e., no feeder systems were observed;

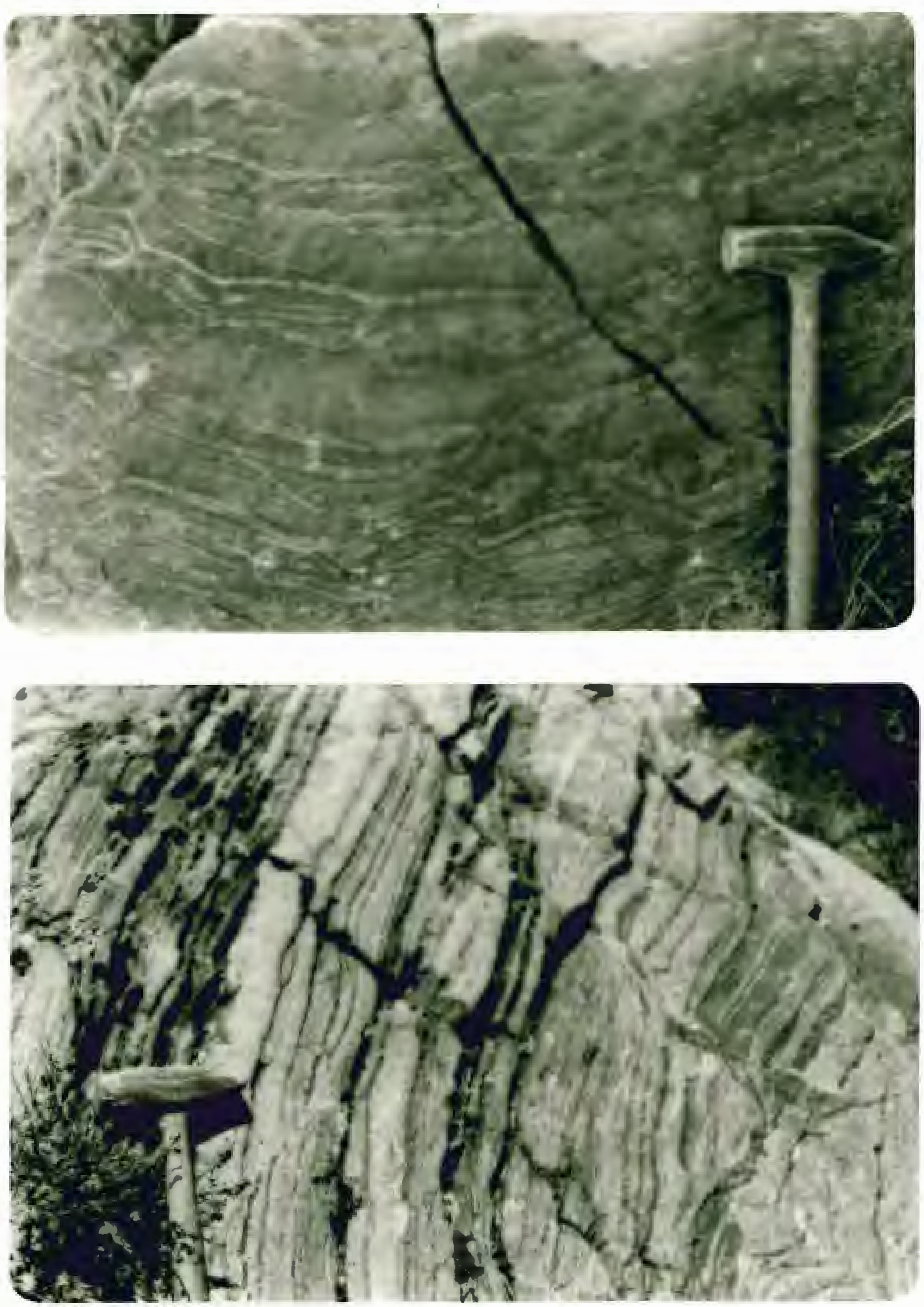


Figure 29. Migmatitic gneiss as typical Little Willow contact is approached.
A - near site 27.
B - near site 28.

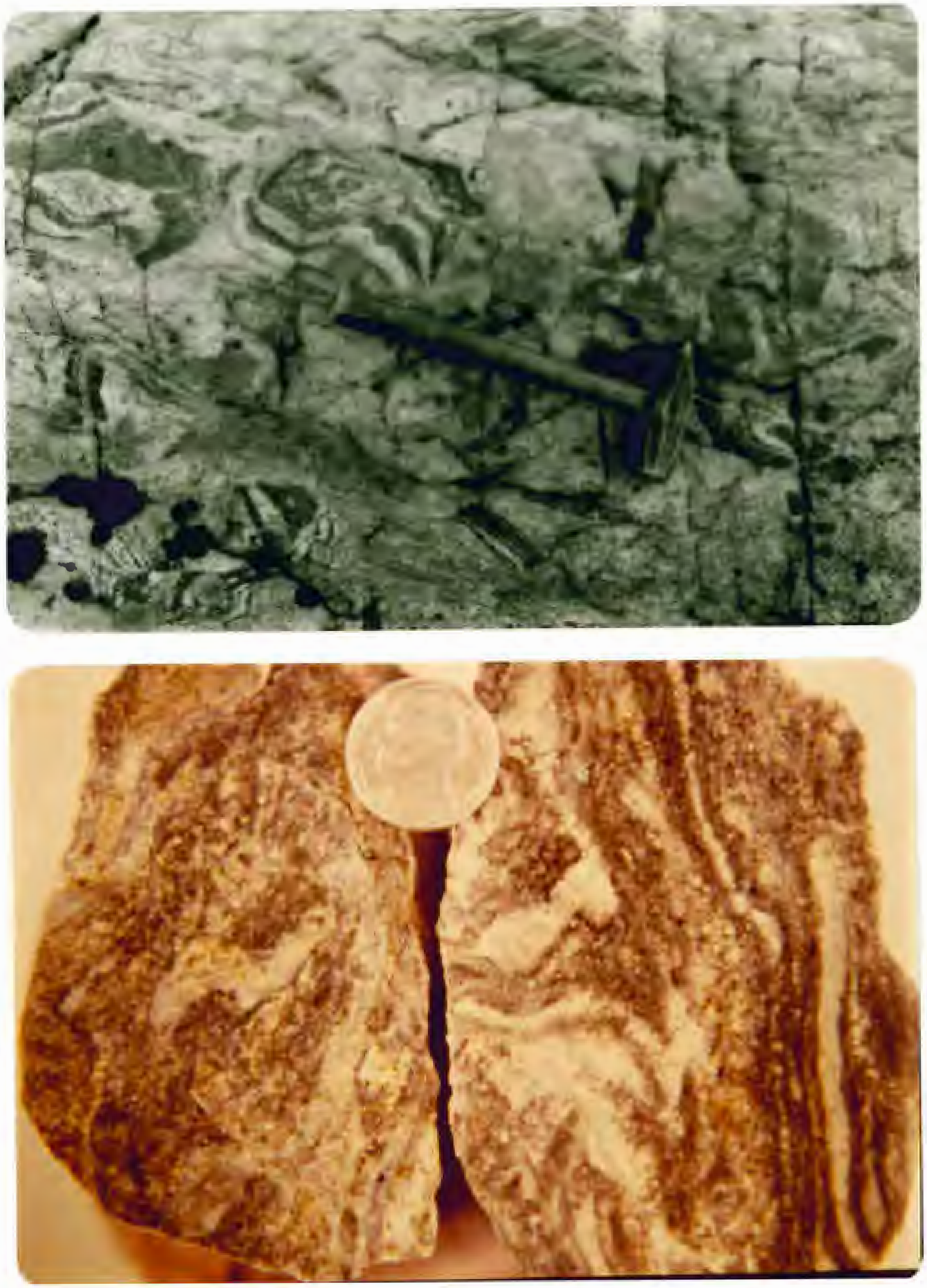


Figure 30. Swirly, schlieric patterned rafts from within the migmatite:
A - Field picture from near site 30.
B - Hand samples. Sample 45-10 on left, 31-1 on the right.

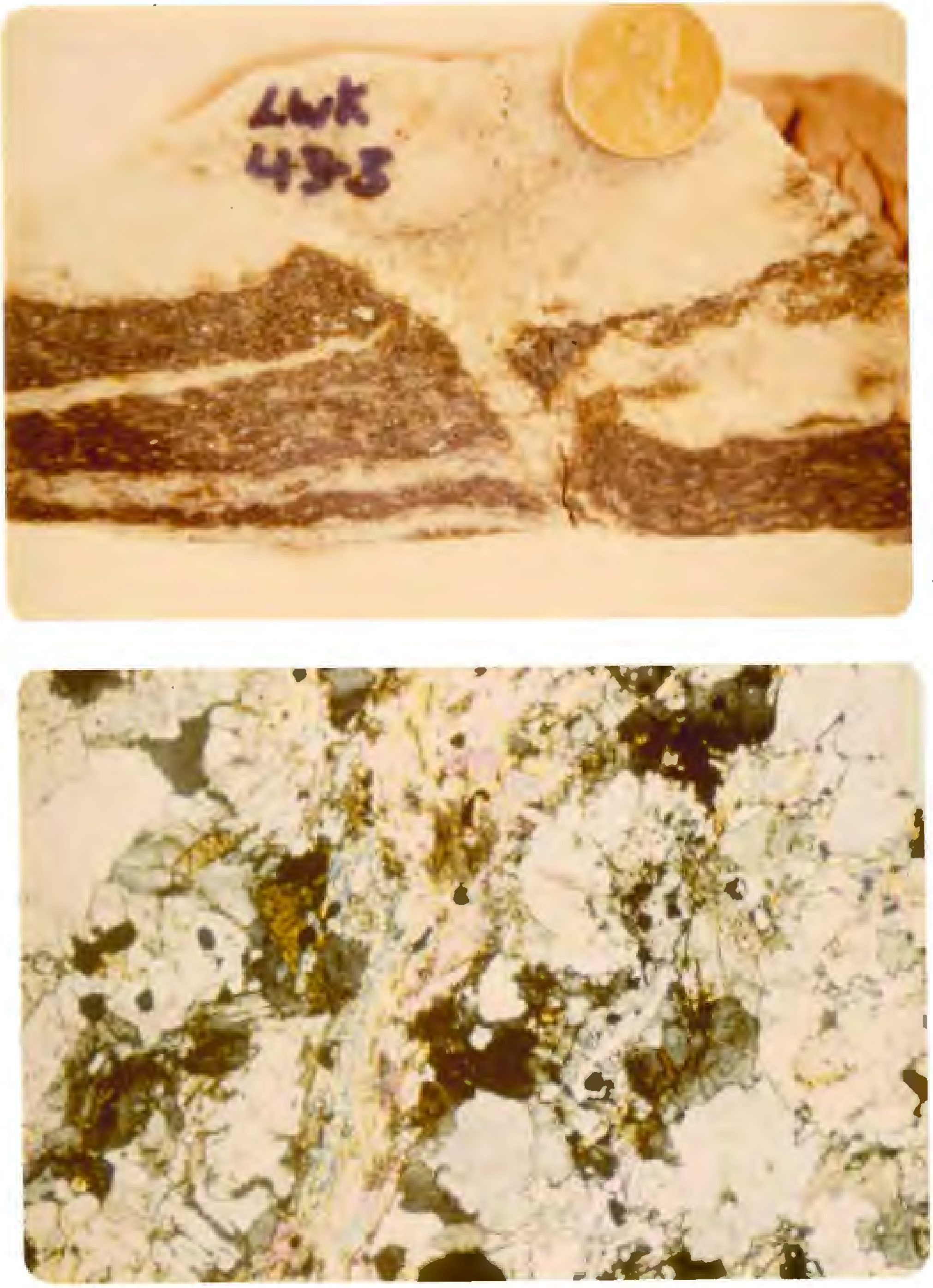


Figure 31. Leucosomatic rinds:

A - Cut hand sample (43-3) which shows light-colored rind around the dark portions;

B - Melanosome of mainly sillimanite, biotite and secondary muscovite, surrounded by a zone a few grains wide of mainly plagioclase and quartz. Sample LW-c, width of field 2.4mm., crossed polars.

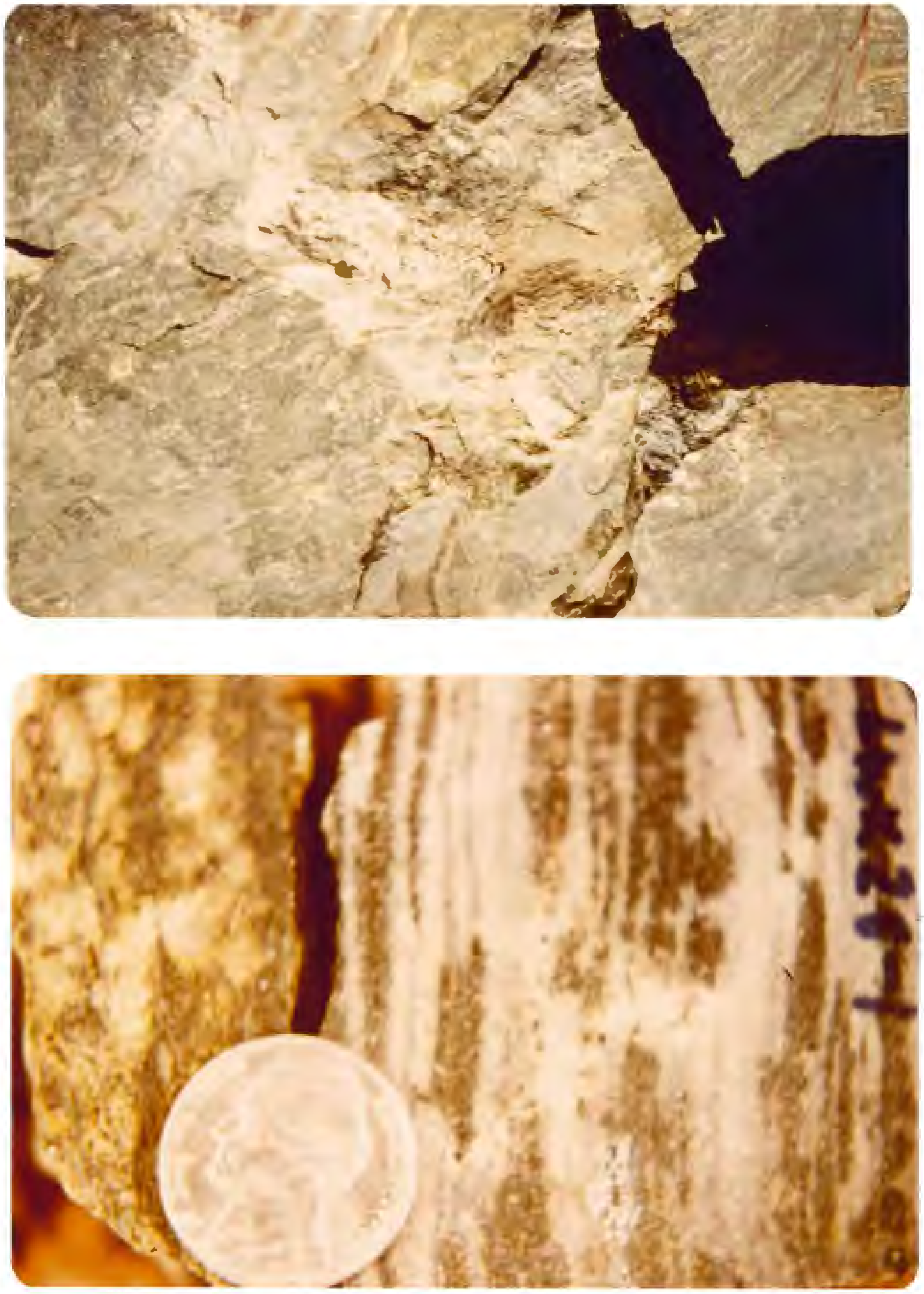


Figure 32. Leucosomatic veinlets and pods:

A - Field picture of pod perpendicular to foliation at site 45;

B - Blebs of quartz and feldspar in sample 36-5 on the left and cross-cutting diffuse veinlet in sample 28-1 on the right.

2) the contact between many pegmatites and the migmatites is not sharp, as would be expected if intrusion was into colder rock.

3) most pegmatites are deformed, with foliations and structures similar to those found in the country rock (Fig. 33); and

4) texturally the pegmatic rocks are granoblastic, not igneous, with sutured grain boundaries and kinked micas fairly common.

Massive quartzitic and quartzofeldspathic portions of the migmatite are granoblastic, usually saccharoidal, indicating the equigranular, recrystallized nature of the rock. Fine-grained recrystallized shear zones within the rock were found locally. Foliation was not evident, but a sub-parallel orientation of fine-grained muscovite was detected.

Leucosomatic rinds, veinlets and pods are always coarser-grained than the surrounding country rock. Foliation is rare, probably due to the preponderance of quartz and feldspar, although microscopically minor micas and sillimanite occasionally exhibit a preferred orientation. The primary texture is granoblastic, but in a few instances relict perthitic, myrmekitic and graphic (Fig. 34) textures were observed.

Melanosomatic layers and rafts always show foliation and often spaced schistosity and crenulation, although it is common for these features to be obscured by the recrystallization of micas. All exhibit a granoblastic texture; where recrystallization was extensive, a decussate texture is visible. Where present, andalusite is typically anhedral and poikiloblastic.

Mineralogically, the quartzitic portions are composed of quartz (90-95%), muscovite (3-6%), and a trace of feldspar. The massive quartzofeldspathic portions varied only in having a greater amount of feldspar (plagioclase 10-20%, K-feldspar 0-15% and sporadic) and slightly more muscovite (5-10%). Opaques, sphene, tourmaline and zircon were the common accessory minerals.

Figure 33. Contorted pegmatic units in the migmatite, near site 45; also note the intense jointing, possible due to the intrusive event.

Figure 34. Relict graphic texture in a leucosomatic pod. Sample 45-14, width of field 6mm., crossed polars.

Leucosomes are composed of quartz (35-65%), plagioclase (30-55%), K-feldspar (0-10% and sporadic), muscovite (1-10%, mostly secondary), biotite (2-5%) and sillimanite (0-2%) mainly as needles within quartz and feldspar. A few grains of cordierite (0-2%) were noticed in some samples.

Melanosomes are composed of sillimanite (30-70%) much of which has been replaced by muscovite (10-15%), quartz (2-20%), feldspar (2-15%), biotite (2-15%), andalusite (0-5%) and cordierite (2-4% locally). Plagioclase was the primary feldspar with K-feldspar occurring sporadically. Accessories include opaques, sphene, tourmaline and zircon.

Opaques (1-10%) were ubiquitious in all units, with magnetite, hematite, pyrite and ilmenite being the common ones. In one rock a two to three centimeter thick layer of magnetite was seen.

Of the pegmatites studied, the average composition was quartz (40-45%), microcline (30-35%), plagioclase (10-12%) and muscovite books (10-15%). Minor carbonate was also discerned.

In the melanosome quartz usually occurs as isolated grains; elsewhere it is intergranular with the feldspars. The majority exhibits a fair degree of recrystallization, with grain boundaries ranging from simple to complex and sutured. Occasionally mortar texture was seen. Recrystallization seemed to be most complete flanking the stock. Inclusions of fresh sillimanite needles in quartz are common.

K-feldspar is rather rare in the migmatites, occurring only sporadically throughout the area. It occurs in the leucosome only where also found in the surrounding rocks. Some highly sericitized grains are thought to have been K-feldspar, as relics of microcline are seen in some of them (Fig. 35). Samples from a one-meter-wide strip, immediately adjacent to the stock, contain as much as 40% poikiloblastic K-feldspar grains (Fig. 36). It was also found as overgrowths on plagioclase, rimming andalusite and in pethitic grains. Sillimanite needles within K-feldspar grains are fairly common. Secondary muscovite was found associated with it.

Plagioclase is the more common feldspar, often being the only one. Much of it has albite twinning. Recrystallization is common, especicially as the stock contact is approached. Locally myrmekitic and perthitic intergrowths and included sillimanite needles are seen.

Sillimanite is the most common aluminosilicate. The most typical habit is as foliated masses in the melanosome, with some occurring as needles in quartz and feldspar grains in the melanosomes and the leucosomes (Fig. 37). In all locations, except right at the stock contact, at least a portion of the foliated sillimanite is replaced by muscovite (Fig. 37), and rarely by biotite. In some places both sillimanite and andalusite are found, occasionally in contact. Both sillimanite crosscutting andalusite (Fig. 37) and rarely andalusite crosscutting sillimanite are seen. The intergrowth of sillimanite with biotite is quite common. Sillimanite is also associated with muscovite, microcline, plagioclase, cordierite and quartz.

Subhedral to anhedral, poikiloblastic andalusite occurs in the melanosomatic portions of the migmatite. Minor replacement by muscovite is seen.

Muscovite is ubiquitious throughout the area. By far the most common habit is as secondary decussate grains, although a small amount

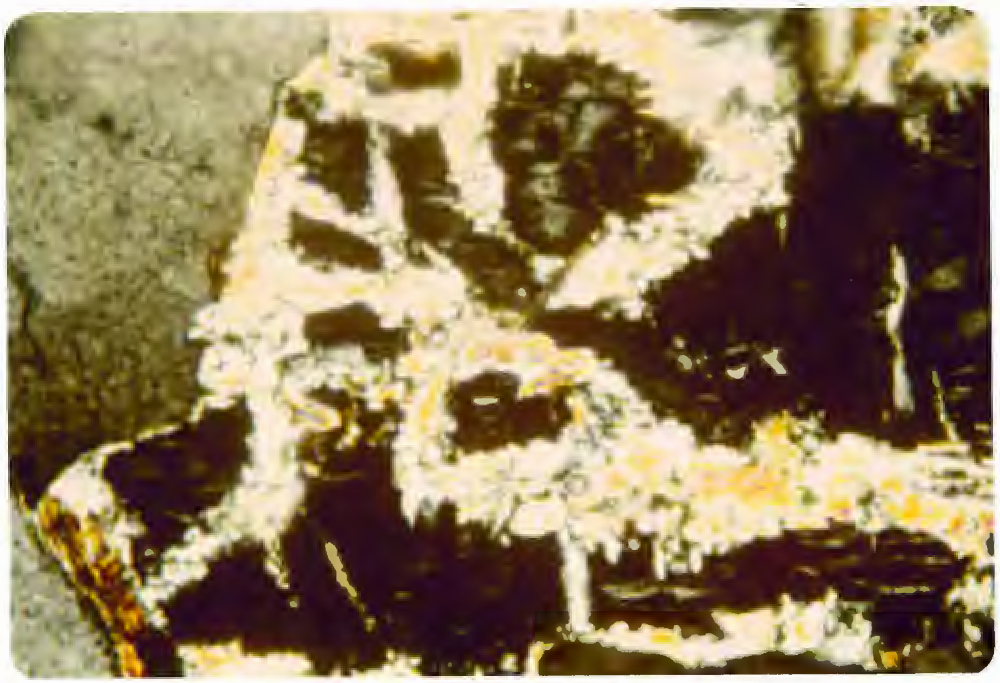


Figure 35. Microcline relics in sericitized grains. Sillimanite needles were found within the microcline. Sample 45-16, width of field 0.6mm., crossed polars.

Figure 36. Fresh microcline megacrysts found right at the stock contact, possibly due to potassium metasomatism. Sample 45-13-C, field of view 2.4mm., crossed polars.

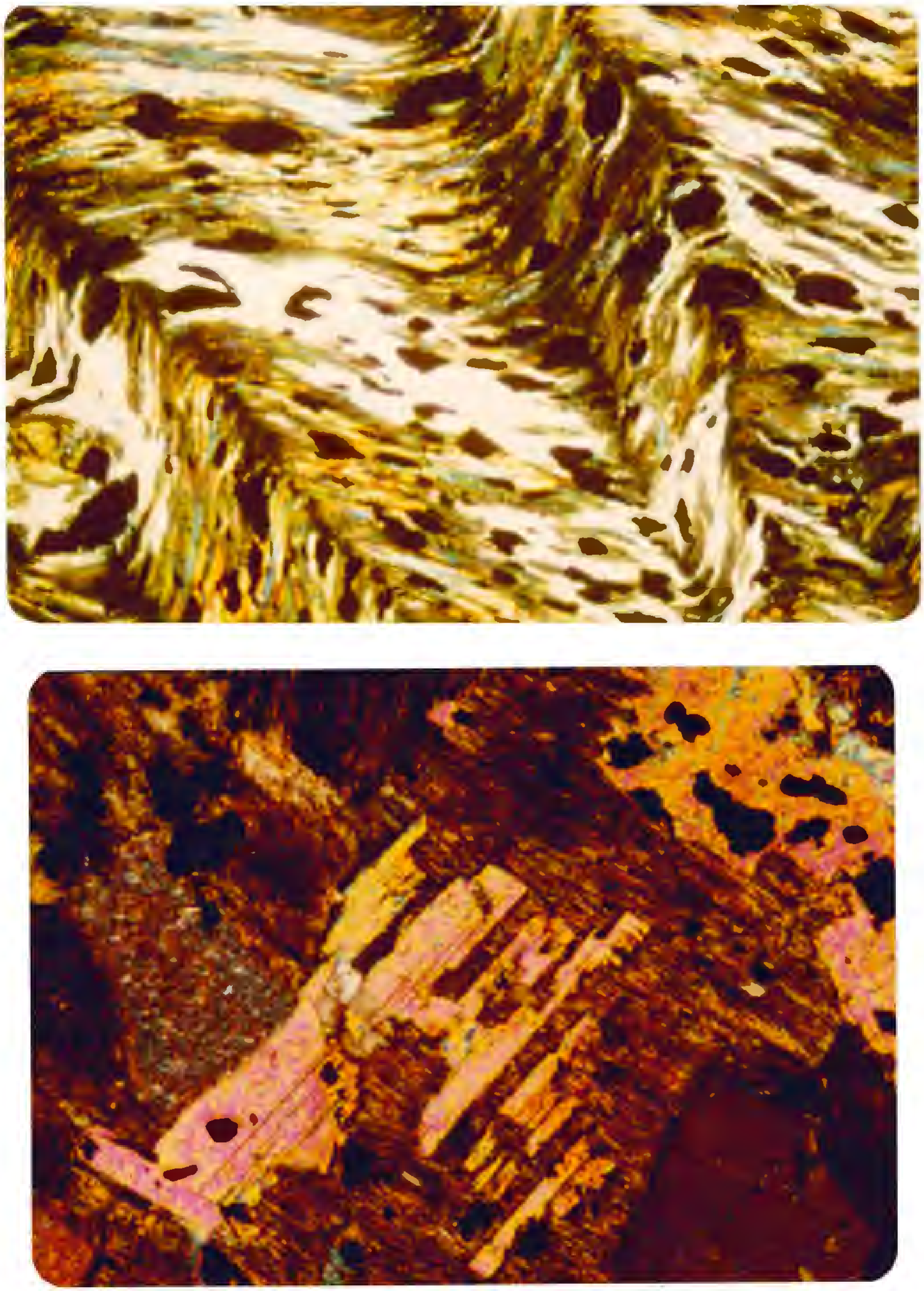
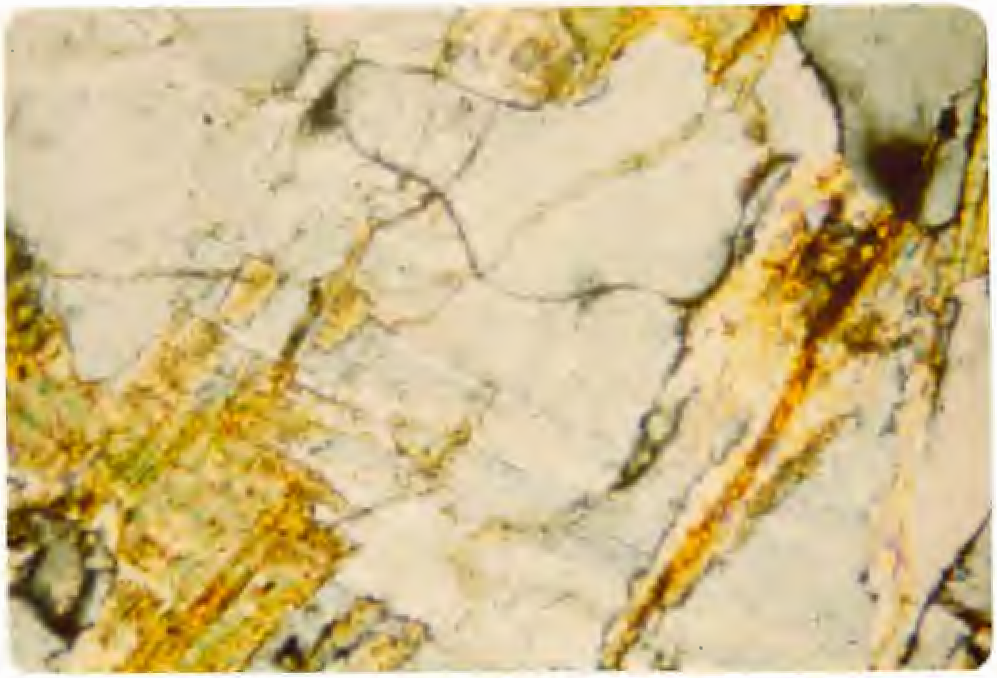
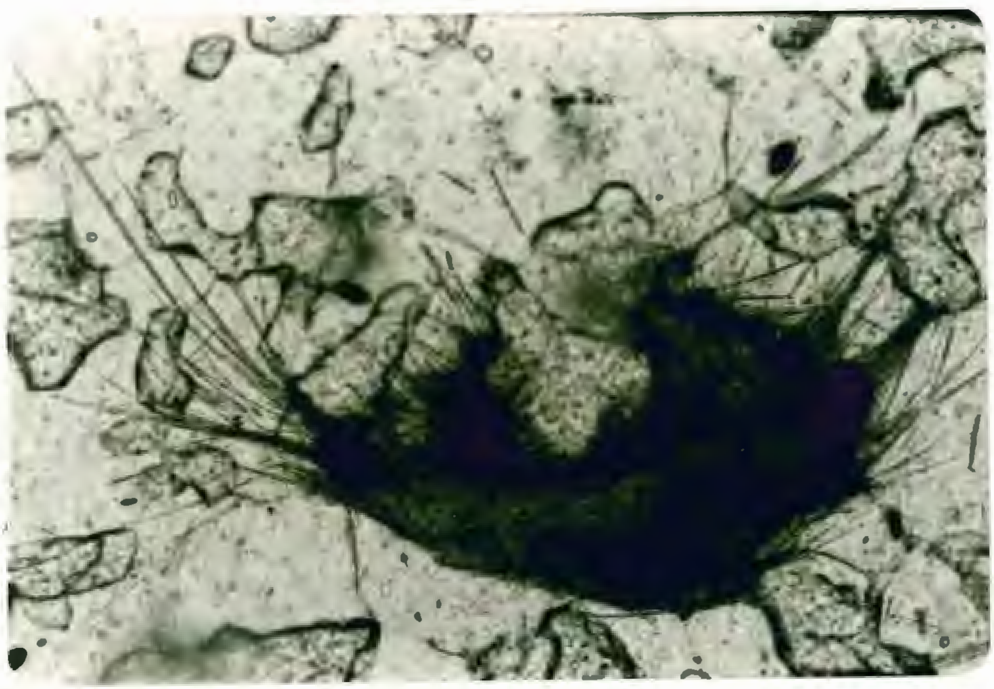


Figure 37. Sillimanite in the migmatites:
A - foliated and crenulated fibrolite. Sample 37-2, width of field 2.4mm., crossed polars;
B - muscovite replacing foliated sillimanite. Sample 45-6, width of field 2.4mm., crossed polars.





C - microcline with included sillimanite, partially replaced by biotite, andalusite is found right beside this grain. Float sample lw-B, width of field, 0.6mm., crossed polars.

D - sillimanite crosscutting andalusite. Sample 40-3, width of field 0.6mm., ordinary light.

is foliated and evidently of an earlier age. Generally it is finer-grained in the felsic rocks and coarser-grained in the pelitic rocks. It is a common replacement of sillimanite, where large flakes are found overgrowing foliated fibrolite. Sericite is also a minor but common replacement of andalusite and feldspars.

Biotite occurs much like muscovite in that the majority is recrystallized and decussate, with minor foliated grains. Biotite is rarer than muscovite, often being completely absent. Chlorite is a very common replacement of biotite (Fig. 38).

Cordierite was also found in some melanosomatic and leucosomatic portions as equidimensional grains (Fig. 39). It always had a slight reddish-brown corrosion associated with it. It is suspected that the more anhedral grains in the melanosome may be of an earlier age and the subhedral grains in the leucosome represent a second generation of cordierite; however no conclusive evidence was found to verify this.

Typical and pertinent assemblages found in the migmatites include:

Quartzofeldspathic portions:

quartz-plagioclase-sillimanite$\pm$K-feldspar$\pm$cordierite$\pm$biotite

Leucosomatic portions:

quartz-plagioclase-sillimanite$\pm$K-feldspar$\pm$biotite

Melanosomatic portions:

quartz-muscovite-sillimanite$\pm$plagioclase$\pm$cordierite$\pm$biotite

quartz-K-feldspar-sillimanite$\pm$plagioclase$\pm$cordierite$\pm$biotite

quartz-K-feldspar-andalusite$\pm$plagioclase$\pm$cordierite$\pm$biotite

quartz-sillimanite-andalusite$\pm$K-feldspar$\pm$plagioclase$\pm$biotite

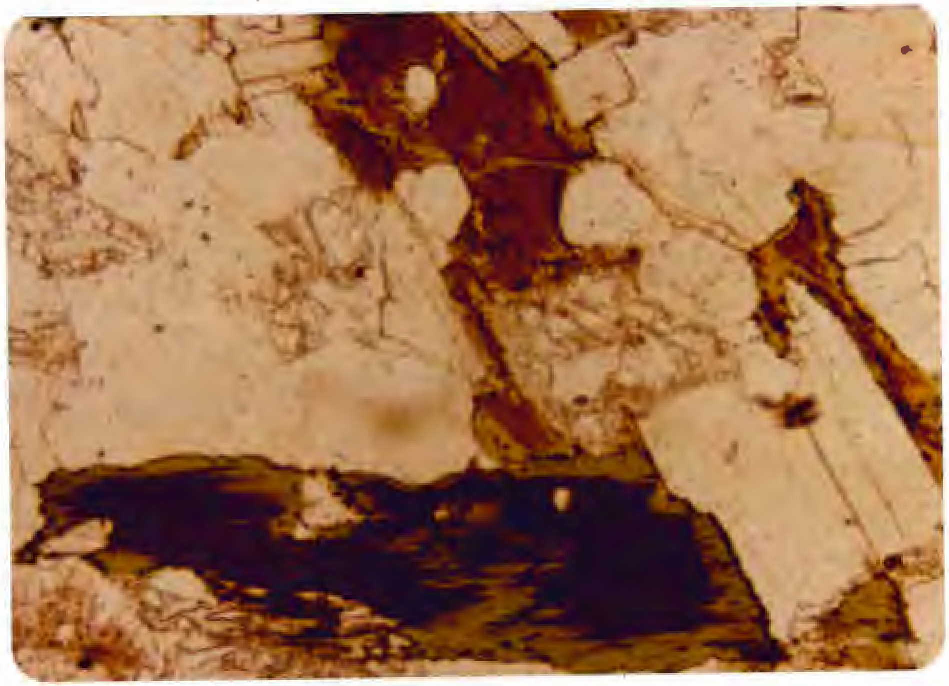


Figure 38. Chlorite replacing biotite. Sample 28-1, width of field 2.4mm., ordinary light.

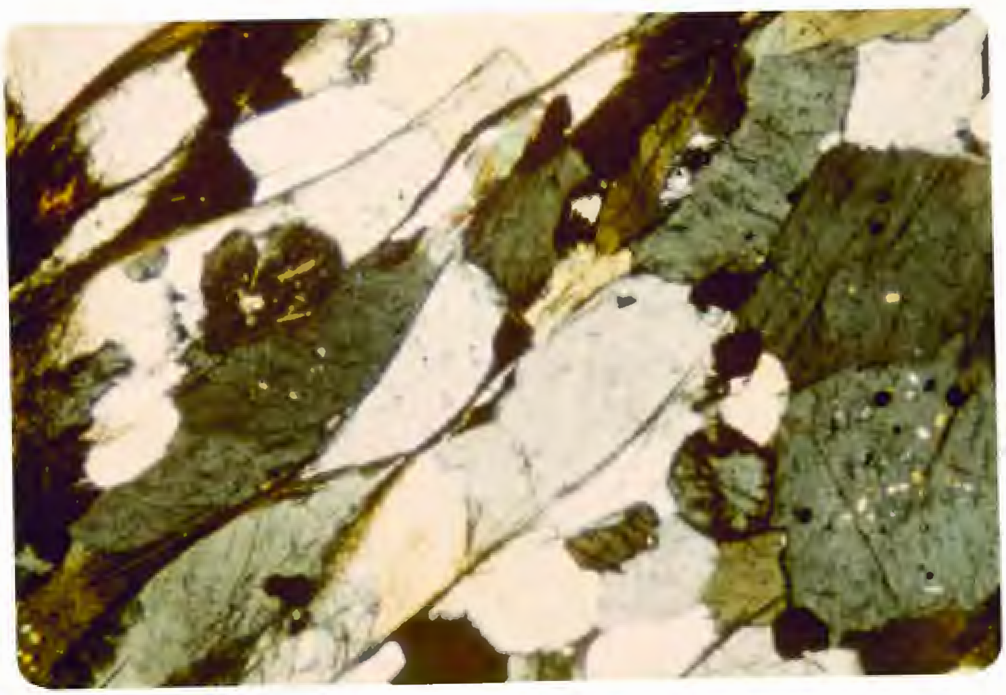


Figure 39. Fresh, subhedral cordierite with minor brownish corrosion, associated with quartz and K-feldspar with included sillimanite needles and foliated sillimanite. Sample 40-3, width of field 2.4mm., crossed polars.

THE BIG COTTONWOOD FORMATION

Immediately to the north and east of the Little Willow Formation is exposed a thick sequence of alternating quartzites and argillites. The ridge to the south of Little Cottonwood Canyon is also capped by rusty colored, intensely fractured quartzites. These rocks are correlated with the lowermost member of the Big Cottonwood Formation (Crittenden, et al., 1964).

Close to the Little Willow contact, these rocks are highly contorted and faulted and hence accurate orientations are difficult to ascertain. Within a short distance east of the contact, however, the moderately dipping strata can easily be recognized (Fig. 40).

The contact between the Little Willow and the Big Cottonwood is, in some locations, very easy to see (Fig. 6, p. 16), but in others can be quite obscure. The contact is believed to be an unconformity (Crittenden, 1952; Neff, 1962) because of the abrupt change in grade of metamorphism at the contact.

The contact is also the locus of some Laramide thrusting. This is verified by the presence of "diapthorite" zones of intense shearing and chloritization, brecciated quartzites, chevron-folded phyllites and slickenslides. Within the thrust zone, some cohesive breccias are found (Fig. 41). The angular, brittley fractured fragments are of Big Cottonwood quartzite and are cemented by carbonate.

The limonitic buff color of the Big Cottonwood quartzites serves to distinguish them from the mottled grey Little Willow quartzites (Fig. 41). The majority of the quartzites are fine-grained. They typically occur

Figure 40. Big Cottonwood Formation a few hundred meters east of site 14. Note alternating beds of quartzite and argillite, and ripples on argillaceous bedding surface.

in beds, 15 cm to 2 m in thickness. Locally, for instance on the south ridge of Little Cottonwood Canyon, the beds are massive.

The quartzites are for the most part, equigranular and locally recrystallized (saccharoidal texture). Secondary foliation is absent in these quartz-rich rocks. Grains are interlocking, with complex, irregular sutured grain boundaries, and subgrain polygonization is not uncommon.

The quartzites are composed of quartz (92-96%), muscovite (2-5%), and feldspar (0-5%).

Interbedded with the quartzites are bluish-grey argillites. Incipient metamorphism has developed a silvery, phyllitic sheen on some of these rocks. Bright green chlorite clots (Fig. 41) and limonitic pseudomorphs after pyrite are not uncommon in these portions of the Big Cottonwood Formation.

The argillaceous beds are two- to thirty cm in thickness. Actually, these layers are thin, interfingering, fine-grained arenaceous layers and pods interbedded with many fine layers of shaly material.

Mudcracks and ripple marks are found on some bedding surfaces (Fig. 40).

The argillaceous units are made up of subangular quartz grains, with minor feldspar, in a fine-grained muscovite and shaly matrix (Fig. 42). Interfingering pods and layers of different grain sizes are evident. In the majority of the samples, the micas are subparallel, indicating a slight development of foliation. Microscopic crenulation of this foliation is seen in some phyllites close to the thrust contact. An exaggerated version of this, a chevron folded phyllite (Fig. 56), is found locally right in the thrust-contact zone. Minor secondary flakes

Figure 41. Big Cottonwood hand samples:

A - Quartzites, cut sample from east of site 2 on the left, brecciated sample from near site 18 in middle and uncut sample from site 34 on the right.

B - Argillites, sample from near site 44 on left, note chlorite clot in top left corner, sample 14-1 in middle and sample 39-1 on the right, note prevalence of andalusite blebs in argillaceous portions.

of biotite are found subparallel to these crenulation axial surfaces. Clots of chlorite, possibly pseudomorphous after cordierite, are quite common throughout the argillite, imparting a hornfelsic appearance to the rock (Fig. 43). Locally, close to the stock, anhedral poikiloblastic andalusite can be seen along with sericite clots (Fig. 44).

The argillaceous units consist of quartz (40-60%), feldspar (trace), muscovite and sericite (20-40%), chlorite (10-15%) and biotite (2-4%). Close to the stock, andalusite (5-10%) and sillimanite (trace) are also found in the argillite. Magnetite, pyrite, limonite, hematite and zircon are all found as accessories in this rock.

Some strain recovery is apparent in the quartz, in the form of subgrain polygonization. Varying intensities of strain in the grains are evident, possibly a factor of proximity to the thrust zone. In the argillaceous compositions, quartz grains are subangular and fairly well sorted in the individual pods and layers.

Feldspar is a minor constituent in these rocks, occurring interstitially with quartz in the quartzites and as separate grains in the phyllites.

Muscovite usually flaunts random habits in the quartzites, but is in subparallel arrangement for the most part in the argillites. Bent flakes are evident in the crenulated portions. Biotite occurs as secondary grains, chlorite primarily as decussate clots.

Close to the stock, andalusite is seen as anhedral porphyroblasts, with sieve texture. Sillimanite, in the form of randomly oriented fibrolite, occurs locally, in contact with andalusite (Fig. 44). Cordierite relicts in chlorite clots were observed in a few cases, but small size prevents definite identification.

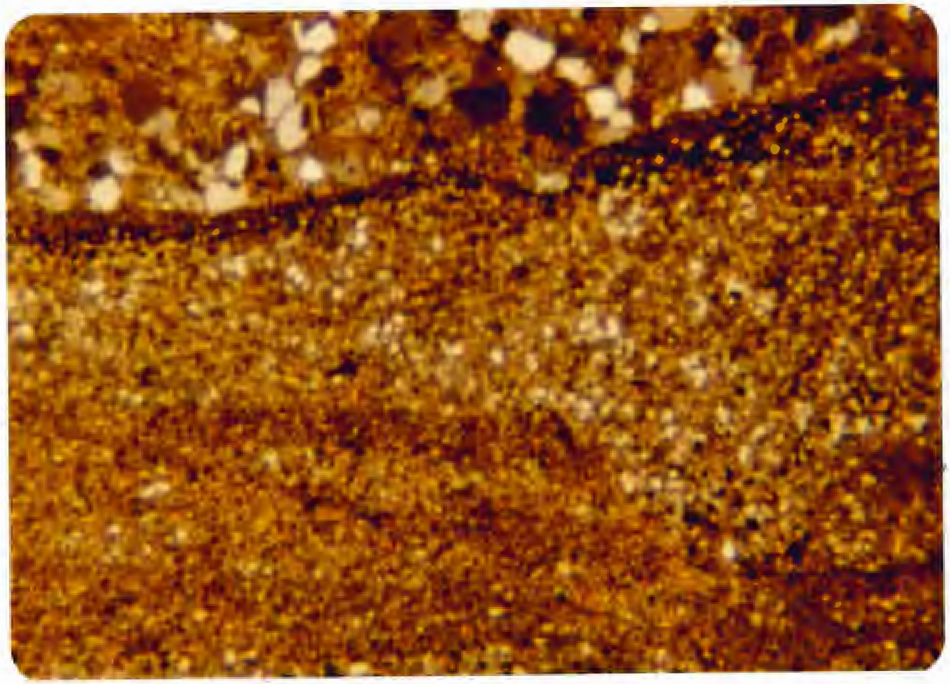


Figure 42. Big Cottonwood argillaceous quartzite, note coarser-grained pod at the top. Sample 14-1, width of field 6mm., crossed polars.

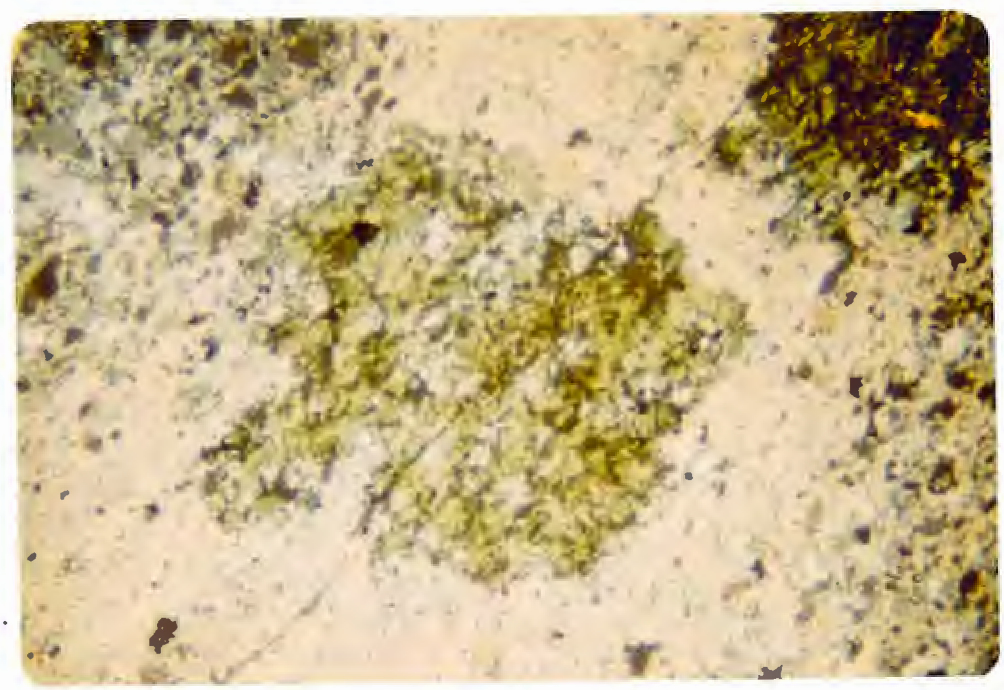


Figure 43. Chlorite clot in Big Cottonwood argillite, note crenulations in the fine-grained muscovite which also are evident in the decussate chlorite. Sample BC-LWC from north of site 44, width of field 6mm., crossed polars.

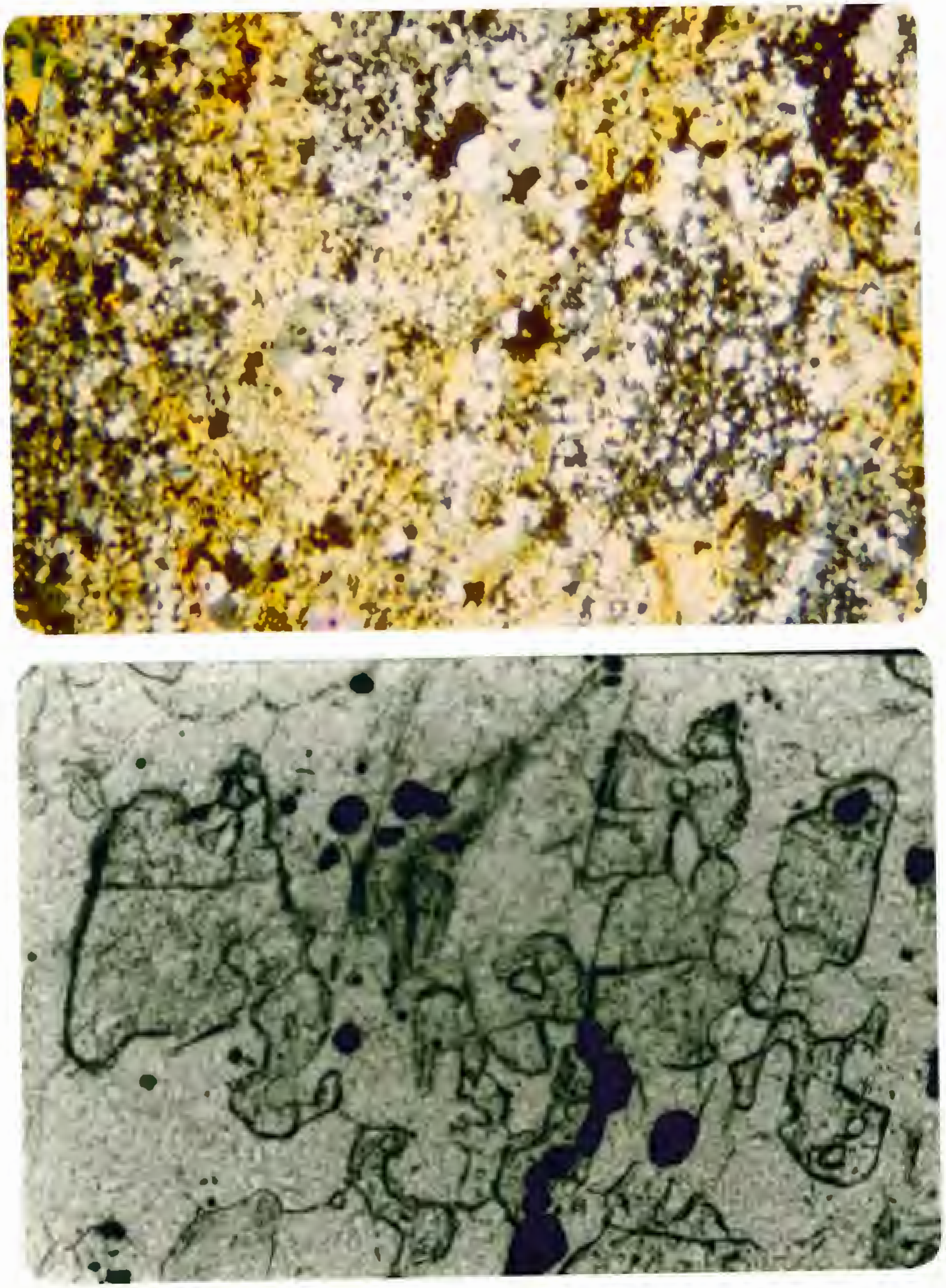


Figure 44. Porphyroblasts in the Big Cottonwood rocks:
A - Anhedral andalusite and randomly oriented sericite clots, possibly replacing cordierite. Sample 39-1, width of field 6mm., crossed polars.
B - Sillimanite and andalusite. Sample 39-1, width of field 0.6mm., ordinary light.

The assemblage quartz-muscovite-andalusite$\pm$sillimanite$\pm$cordierite(?) is of major significance when considering the metamorphism of this area.

THE LITTLE COTTONWOOD STOCK

The Little Cottonwood Stock, which abuts the Little Willow Formation in the vicinity of Little Cottonwood Canyon, is a quartz monzonite intrusion which has been dated at 24-31 million years (Crittenden, 1973). In outcrop, it is the most extensive intrusion of the Central Wasatch Mountains, having a roughly elliptical shape, approximately 14.5 km x 9.5 km. In this study, only intrusive rock from close to the Little Willow contact was examined.

The intrusive contact on a gross scale is nearly concordant with the country rock. This has also been reported from other parts of the contact (James, 1979). The contact dips very steeply to the west in Little Cottonwood Canyon (Fig. 7, p. 16). Where observed, the contact is sharp, sometimes brecciated, faulted and with limonitic stains. A few small xenoliths of Little Willow Formation can be seen in the plutonic rock, some appearing to be partially assimilated. Close to the stock, foliations and boudinage structures in the Little Willow are subparallel to the contact. Development of a strong pervasive joint system and veining in the Little Willow Formation was evident at the contact (Fig. 59).

The Little Cottonwood quartz monzonite is a very light-colored rock, which when viewed from a distance appears almost white. Distinction of this rock from the dark-colored Little Willow rocks is quite easy, both on airphoto and in the field. Where exposed, this massive granitic rock forms majestic steep walled ridges.

The quartz monzonite is coarse-grained and porphyritic, with microcline phenocrysts and quartz eyes conspicuous on the weathered surface (Fig. 45). Concentrations of mafic minerals are common in the quartz and feldspar matrix. Texturally, the rock is hypidiomorphic and holocrystalline.

This rock is composed of plagioclase (40-50%), K-feldspar (20-30%), quartz (20-35%), biotite (5-8%) and hornblende (1-5%). Sphene and zircon are the most common assessories, along with minor sulfides.

One sample, collected right on the faulted contact, shows signs of moderate alteration. Greater alteration of feldspars to sericite, development of minor muscovite and replacement of hornblende and biotite by chlorite are among the effects noted.

The majority of the microcline forms, with quartz, a coarse groundmass to subhedral plagioclase laths and microcline phenocrysts. The phenocrysts constitute 0-10% of total K-feldspar and may be up to 1.5 cm across. They are commonly zoned and occasionally rimmed by altered plagioclase (rapakivi texture) (Fig. 46). They poikilitically enclose euhedral biotite and hornblende and subhedral, often zoned plagioclase. Alteration is minor.

Plagioclase occurs as subhedral to euhedral laths in all samples studied. Zoning is common and albite twinning is also evident. Occasionally inclusions of partially resorbed hornblende can be found in the plagioclase grains. Again, alteration is minor.

Quartz is found as phenocrysts, quartz eyes and as finer-grained anhedral matrix. Quartz eyes are usually anhedral in outline. All quartz exhibits undulatory extinction.

Figure 45. Hand samples of Little Cottonwood quartz monzonite: Left - float sample with resistant microcline phenocrysts; Right - sample 45-13-H.

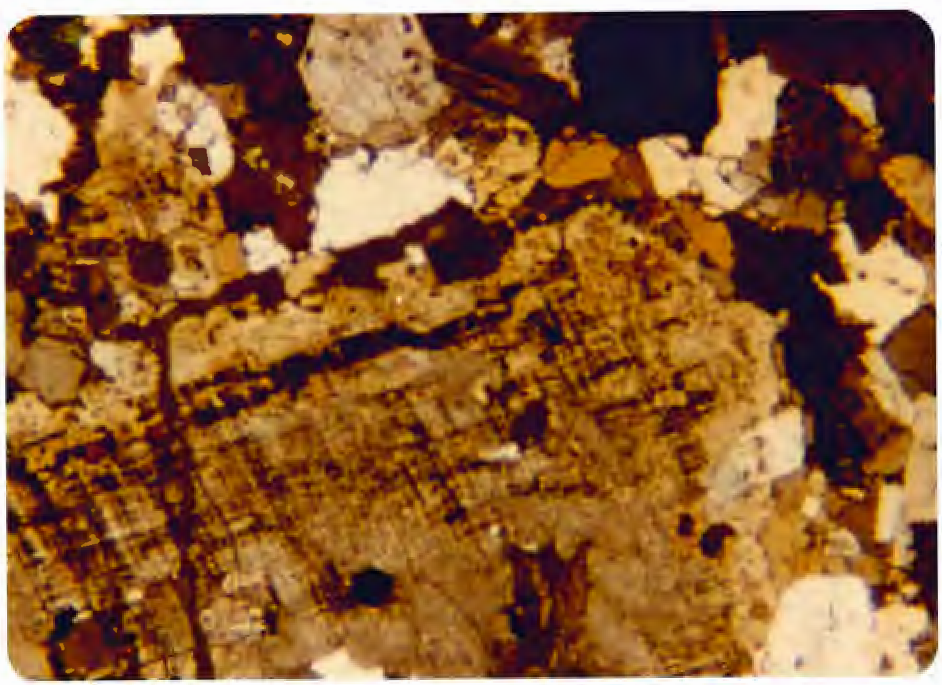


Figure 46. Quartz monzonite showing microcline phenocryst rimmed by altered plagioclase. Sample LCS-2, width of field 6mm, crossed polars.

Brown biotite occurs mainly as large euhedral grains, although some appears to be replacing hornblende. Blue-green hornblende is most common as euhedral grains, some of which are partially resorbed or replaced.

QUATERNARY DEPOSITS

Excellent effects of glaciation are seen at the mouth of Little Cottonwood Canyon; in fact, this canyon contained the longest (22 km) and largest (195 m thick at the terminus) glacier in the entire Central Wasatch area (Marsell, 1964).

The most obvious result of the glaciation is the symmetrical U-shape of The Little Cottonwood Canyon (Fig. 47).

Morainal deposits are found at the mouths of Little Cottonwood and Bell's Canyons. To the west of Little Cottonwood Canyon, a southern lateral moraine forms a ridge extending into Salt Lake Valley (Fig. 48). The northern moraine forms a bank of debris lodged on the valley wall; scattered boulders of Little Cottonwood quartz monzonite can be seen resting on Little Willow or migmatitic rock. In both cases, the moraine extends about 100 m above the canyon flood plain (Atwood, 1909). A terminal moraine is thought to have been eroded away from the mouth of this canyon by streams (Marsell, 1964).

Evidence of two earlier epochs of glaciation can be seen along the walls further up the canyon. The oldest glacier is believed to be mid-Pleistocene in age and is reported (Richmond, 1961) to have had its deposits lifted more than 760 meters by normal faulting. These deposits may correspond to the probable morainal deposit located within the Little Willow outcrop area, along the intersection of Sections 1 and 12 (Plate 1).

To the west of the Little Willow and the morainal deposits, in Salt Lake Valley, are interfingering and overlapping pluvial lacustrine and outwash deposits. Alpine deposits were laid down between the second and third glaciations; Bonneville sediments followed the last advance (Richmond, 1961).

The floor of Little Cottonwood Canyon is covered with talus and landslide debris, as well as sediment deposited by the contained river.

Deposited and eroded terraces of former Lake Bonneville, and scarplets from recent Wasatch Fault movements in these unconsolidated deposits (Fig. 48), are the most recent additions to the topography of this area.

SPECULATION ON THE ORIGIN OF THE ROCKS

Two logical origins for the quartzofeldspathic gneiss exist: an arkosic sediment or felsic igneous rock. Neff (1962) speculated that because of the rapid compositional change, a sedimentary origin would be more likely, since an igneous rock would tend to be more homogeneous. James (1979) writes that locally textures suggest intermediate to felsic ash flows and pyroclastic tuffs, although no specifics were offered.

The presence of overgrowths on rounded zircons and the interlayering with pelitic rocks and metaconglomerates would lend support to a sedimentary origin. High feldspar amounts may suggest an igneous origin.

The amphibolite units apparently were interlayered with or intruded into the quartzofeldspathic gneiss. Crittenden (1965) considered the amphibolites to be mafic sills because of supposed concordance with surrounding units, but in places (e.g., southeast of former Gold City) the rock seems to cut across metasedimentary units (James, 1979). The fining

Figure 47. Looking east into the U-shaped, glaciated Little Cottonwood Canyon. The southern lateral moraine can be seen on the right. (photo courtesy of Tim Holst)

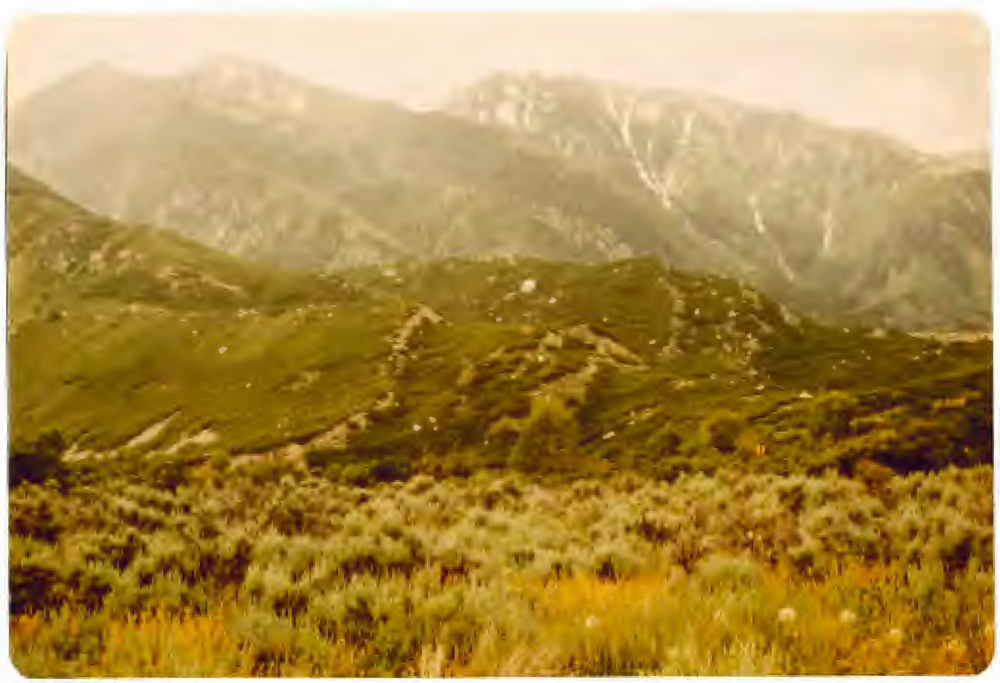


Figure 48. The southern lateral morainal ridge which extends into Salt Lake Valley. Note the recent fault scarps in the unconsolidated gravels. (photo courtesy of Tim Holst)

of the grain size toward the contacts may also indicate an intrusive origin, as dikes or sills. Mafic flows interbedded with quartzofeld-spathic units is another consideration.

A sedimentary origin for the mica schists is apparent. Evidence for this includes the pelitic compositions, association of conglomeratic units and the presence of concentrations of zircon and tourmaline grains. The quartz-sericite schist, with its higher percentage of K-feldspar, may represent a tuffaceous or ash layer, derived from a felsic source.

The interfingering and lensoidal nature of the various subunits, along with the conglomerates and the rapid variation in grain size, suggest a shallow water environment, probably fluviatile.

K-Ar radiometric dates of 27.3 and 19.1 million years for biotite and muscovite respectively in the mica schist (Whelan, 1969) implies that the most recent recrystallization of these rocks took place at this time, roughly coincident with the intrusion of the Little Cottonwood stock.

A sedimentary, near-shore origin is tenable for the migmatitic unit as well, because of the prevalence of quartzites, the heterogenity of the rock and the presence of pelitic portions.

The Big Cottonwood Formation is undoubtedly sedimentary as indicated by the interbedding of quartzites and argillites, deposited unconformably on the Little Willow metamorphic complex. Ripple marks and mudcracks further support a subaerial-shallow water environment.

Radiometric dating indicates intrusion of the Little Cottonwood Stock 24-31 million years ago (Crittenden, 1973). The sharp contact with the country rock, the lack of concurrent regional metamorphism, the rarity of related pegmatite dikes and the porphyritic nature of the quartz monzonite suggest an epizonal level of emplacement (Hyndman, 1972) for the stock.

STRUCTURAL GEOLOGY OF THE LITTLE WILLOW AREA

GENERAL STATEMENT

The structure of the Little Willow Formation and surrounding area is complex in that the rocks reflect several episodes of folding and faulting. This is evident in the highly contorted and fractured nature of much of the outcrop.

In dealing with the structural information, all data from the "typical" Little Willow Formation (including the quartzofeldspathic gneiss, the amphibolite and the mica schists) were treated as one set. Likewise, any data derived from the Big Cottonwood Formation were grouped. Information from the migmatitic unit was considered separately in an attempt to find relationships between it and the other units.

All stereo projections contained herein were first plotted on the Schmidt equal area net; some were then contoured with the aid of the Kalsbeek counting net (from Ragan, 1973).

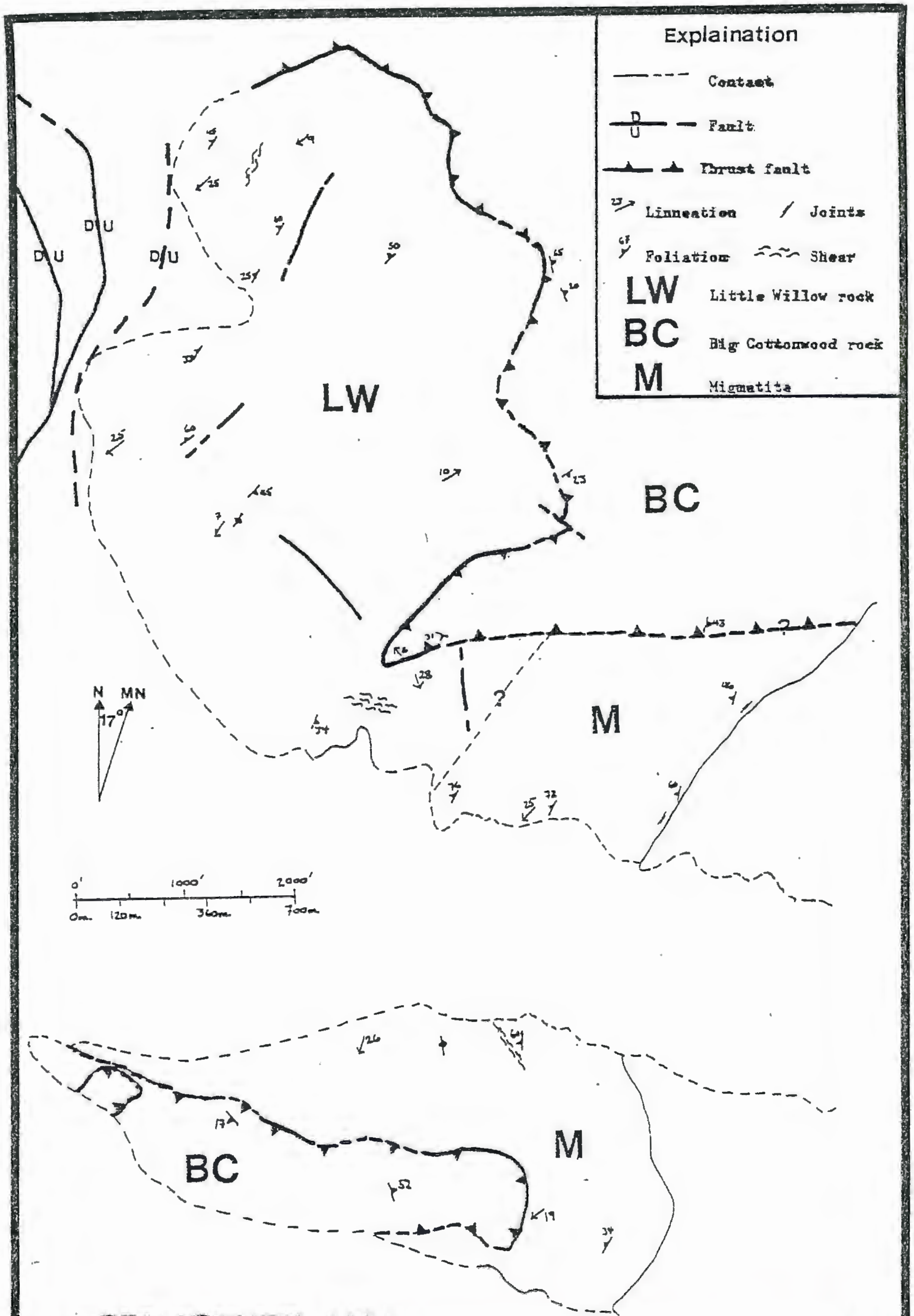
The included structural map (Fig. 49) is a compilation of data from this work, the work of Crittenden (1965), Neff (1962) and James (1979).

PENETRATIVE STRUCTURES

Folds

Folding is common in the Little Willow rocks and the migmatitic unit; it occurs rarely within the Big Cottonwood rocks, and is then confined to the vicinity of the thrust zone.

The most conspicuous folding seen in the mica schists is small crenulations of the primary (S_1) foliation. These folds range from microscopic size (Fig. 22, p. 36) to two centimeters in amplitude. The



F. 49 STRUCTURE MAP

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migmatite occasionally shows this crenulation, but it is often obscured because of extensive recrystallization. Along the Little Willow-Big Cottonwood thrust contact, highly accentuated chevron folds (Fig. 56) and minute crenulations on apparent bedding surfaces are seen in the more argillaceous units of the Big Cottonwood rocks. Crenulations, as well as most small scale folding, is rare in the rest of the Little Willow Formation and the Big Cottonwood rocks farther away from the thrust.

Within the migmatite, complex and contorted folding is common. Several upright, subhorizontal isoclinal folds, with average amplitudes in tens of feet, were also observed (Fig. 50). Larger scale folding was also seen in the Big Cottonwood rocks in the vicinity of the thrust.

The entire area is part of a large anticlinal fold (the Cottonwood Uplift) which trends east-west. The Little Willow and migmatitic rocks form the steeply dipping core of this structure; Big Cottonwood rocks "lap" over the older core and generally dip away from it.

Twenty-five poles to axial surfaces were measured and plotted (Fig. 51). They consist primarily of crenulation axes from the mica schists and isoclinal folds from the migmatite. Although a crude pattern exists in the migmatite folds, where a northeast-southwest girdle of poles is suggested, there are too few points and too much scatter for interpretation.

Foliation and Schistosity

Foliation planes within the mica-rich units of the Little Willow Formation are well defined by the preferred orientation of the micas,

Figure 50. Upright, subhorizontal isoclinal folds in migmatitic gneiss near site 27.

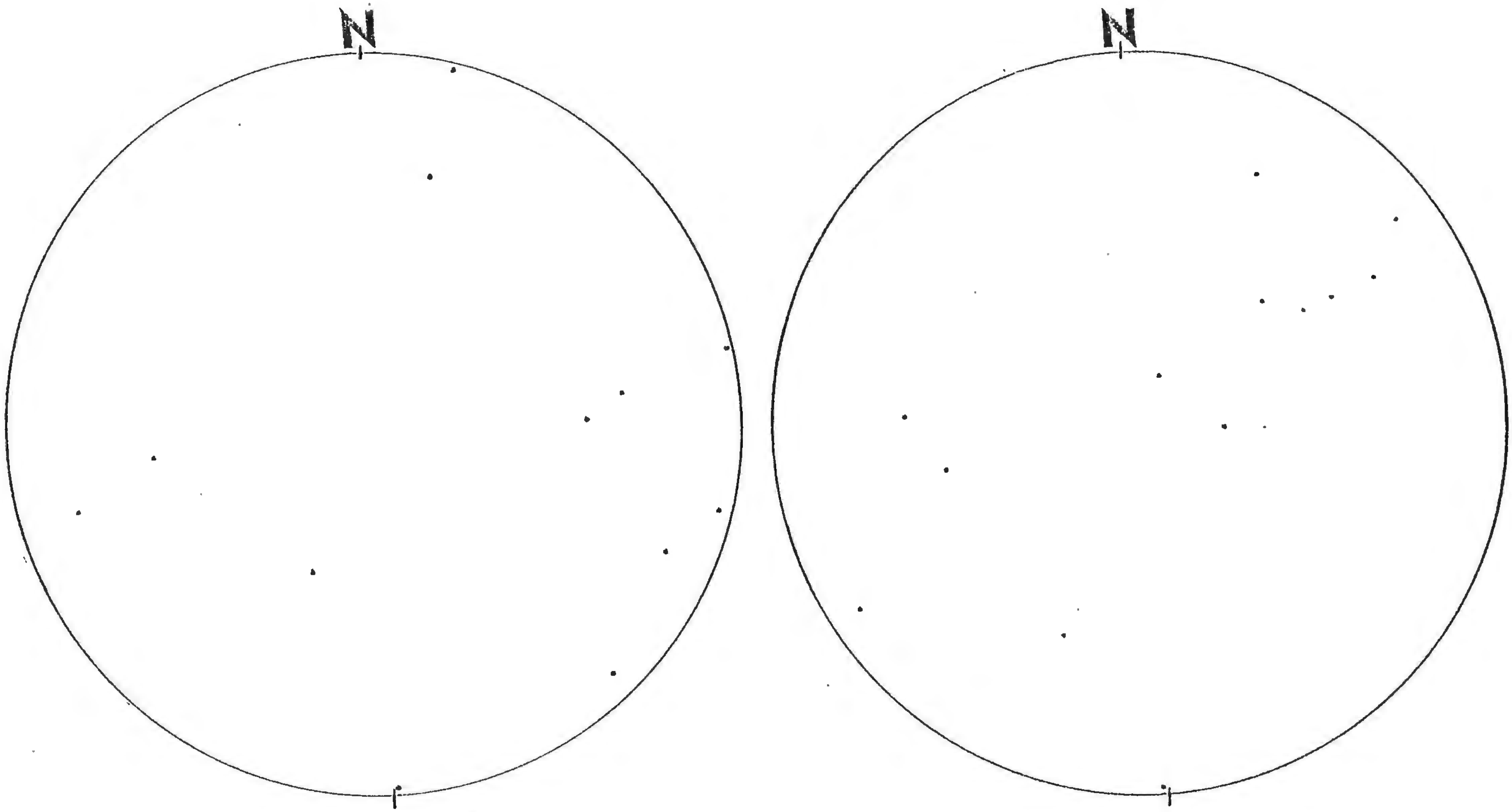

Figure 51. Poles to Axial Surfaces: A - Typical Little Willow rocks, 12 data points.
B - Migmatitic rocks, 13 data points.

and the definite layering developed between those layers and the quartz and feldspar rich layers (spaced schistosity). Both dimensional and lattice preferred orientations are common in these rocks. Foliation in the amphibolite and the quartzofeldspathic gneiss is not as well developed, but where observed is also defined by preferred mineral orientation.

The migmatitic unit also contains well developed foliation, also characterized by segregation and preferred orientation of layer silicates. Commonly within this unit, the foliation plan is defined by the preferred orientation of masses of sillimanite needles (fibrolite).

Spaced schistosity is absent in the Big Cottonwood rocks. Minor foliation does exist as alignment of fine-grained micas, subparallel to the bedding surface, imparting a phyllitic sheen to the more argillaceous rocks.

Within the Little Willow and migmatitic units, a second direction of foliation (S_2), characterized by recrystallization of micas subparallel to crenulation axial surfaces is observed. This phenomenon is particularly evident on the microscopic scale.

When plotted and contoured (Figs. 52 and 53) the S_1 foliations from the typical Little Willow appear to be nearly homoclinal, with average strike around North 25° East, and an average dip of 40° Northwest. Foliations in the migmatites have a concentration which is close to the high in the Little Willow foliations. The average strike for migmatite foliations is North 20° East, with an average dip of 60° Northwest. This suggests a common deformation for the two units.

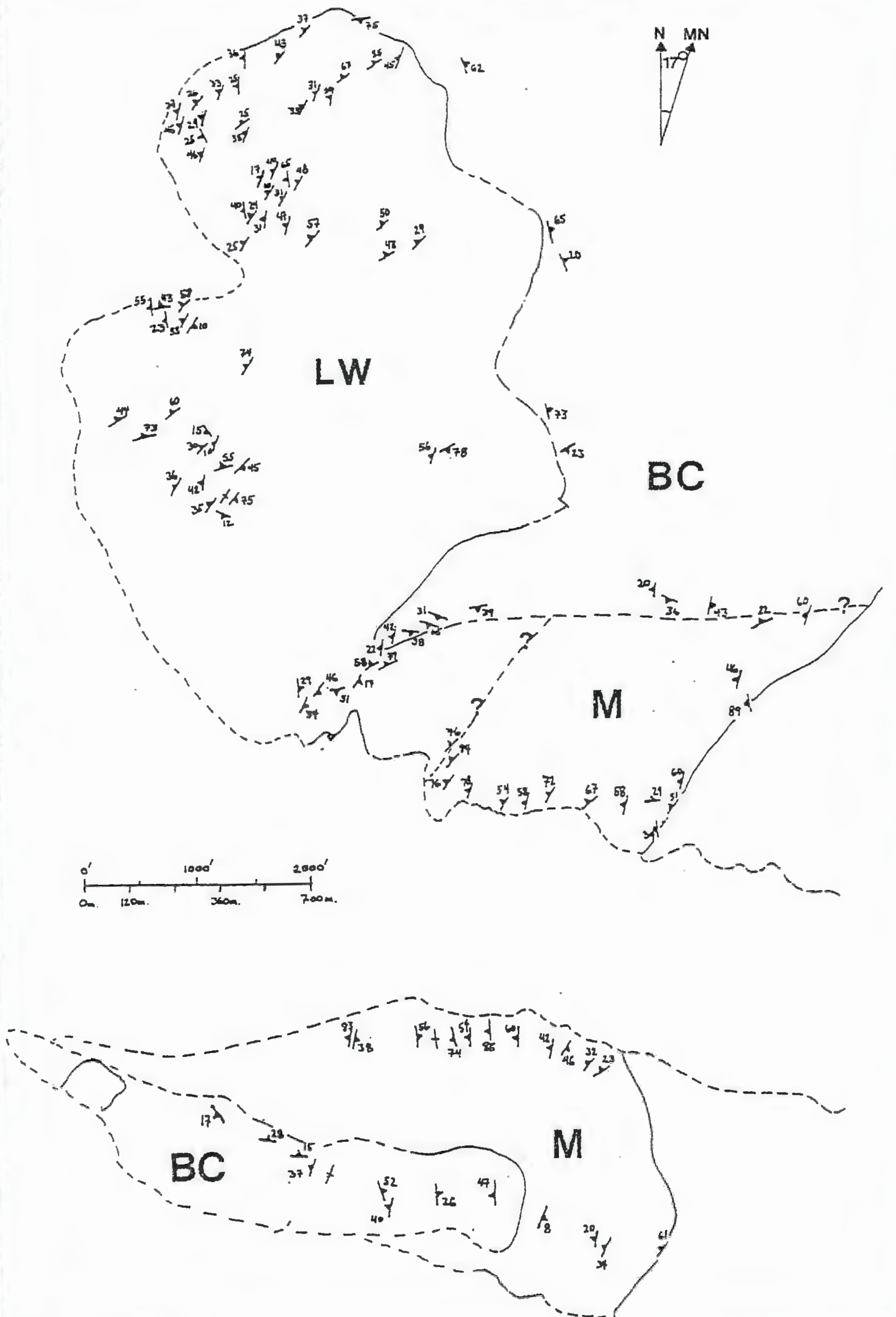

Figure 52. Foliations.

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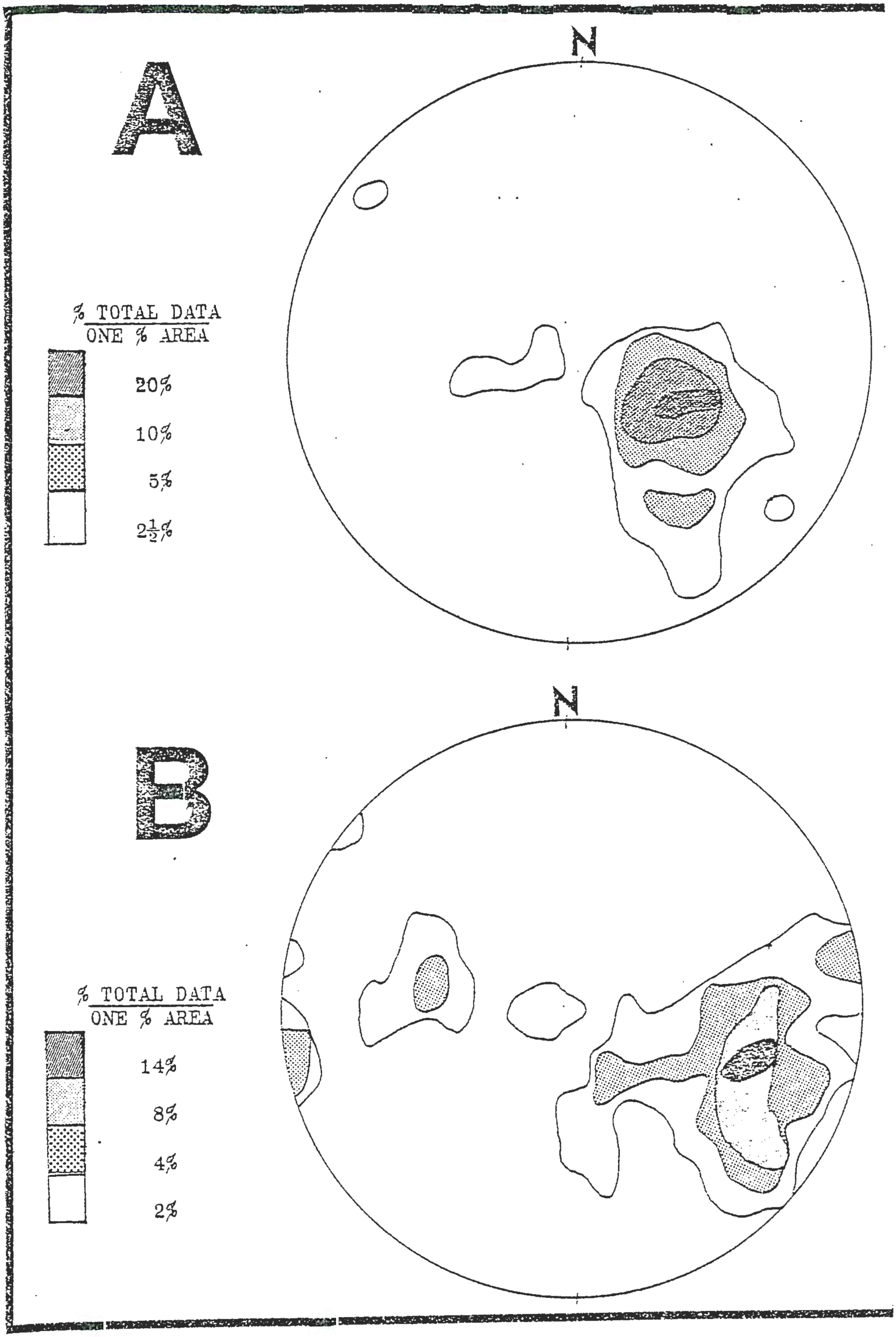
A
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% TOTAL DATA
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20%
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4%
2%

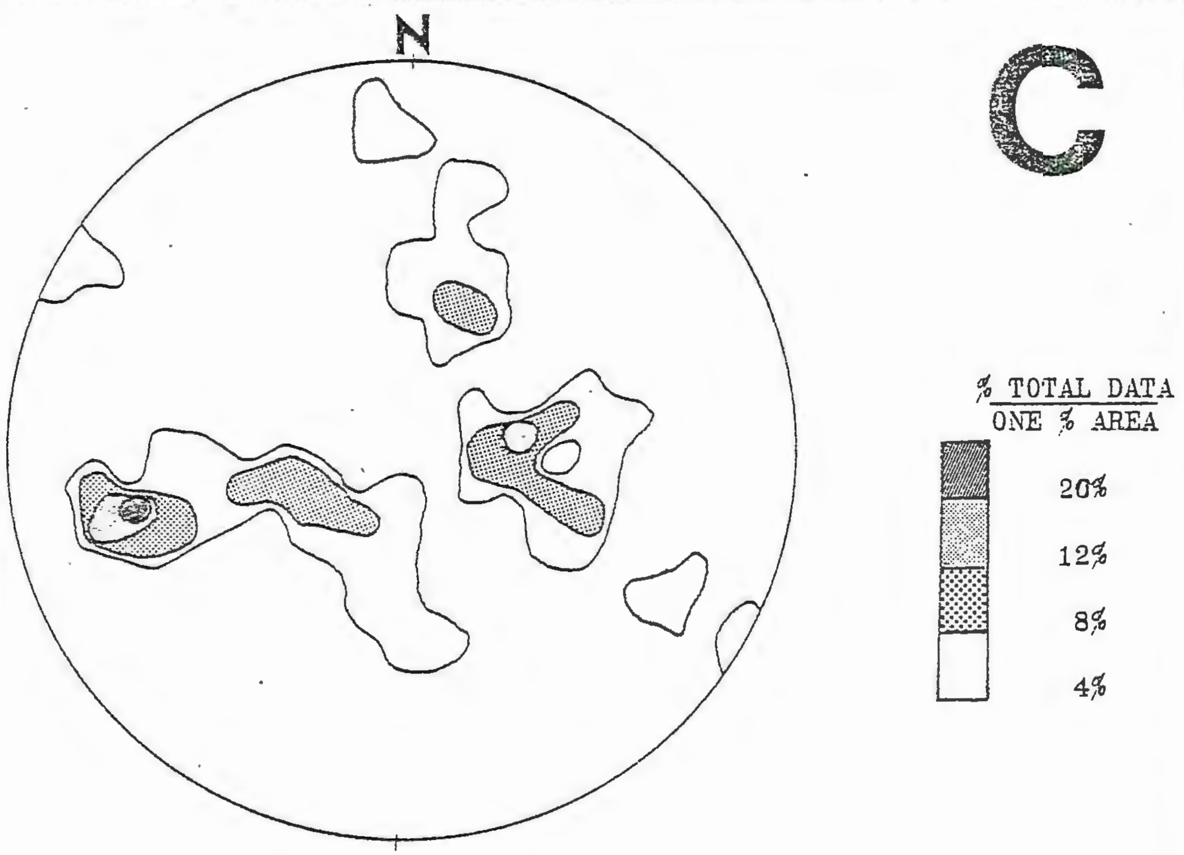

Figure 53. Plotted and contoured poles to S_1 foliations:
A - Typical Little Willow rocks, 82 data points, average high concentration value: North 25° East, 40° Northwest dip.
B - Migmatitic rocks, 49 data points, average high concentration value: North 20° East, 60° Northwest dip.
C - Big Cottonwood rocks, 24 data points, average high concentration value: North 10° West, 55° East dip.

The high for the Big Cottonwood foliation plot (North 10° West, 55° East dip) reflects the average bedding surface for these rocks away from the thrust fault. The dispersion of points around the plot may be due to the folded and fractured nature of the rocks at the thrust.

Lineations

Lineations obtained in the typical Little Willow rocks were defined by crenulation axes (S_1 and S_2 intersection). A look at the plotted lineations (Figs. 54 and 55) for the Little Willow rocks reveals a strong concentration at 210-240° with an average plunge of 10°.

Lineations in the migmatitic unit are very few and are defined by larger fold axes, boudins and mineral and slickenslide streaks on foliation surfaces. Boudins and fold axes are prevalent, but often their limited exposure prevented accurate measurement. The minimal data was found to be scattered.

Within the thrust fault zone a few readings were obtainable on apparent Big Cottonwood crenulation axes. The average value obtained was 330°, with a plunge of 5°. Apart from these, lineations were not found in the Big Cottonwood Formation.

NONPENETRATIVE STRUCTURES

Faults and Shear Zones

Cataclasis, observable both microscopically and megascopically, is quite common throughout the study area, especially within the typical Little Willow rocks.

Cataclastic texture, consisting of the development of porphyroclasts, mortar texture and microfaulting is rather common in the quartzofeldspathic gneiss, but rare in the meta schists.

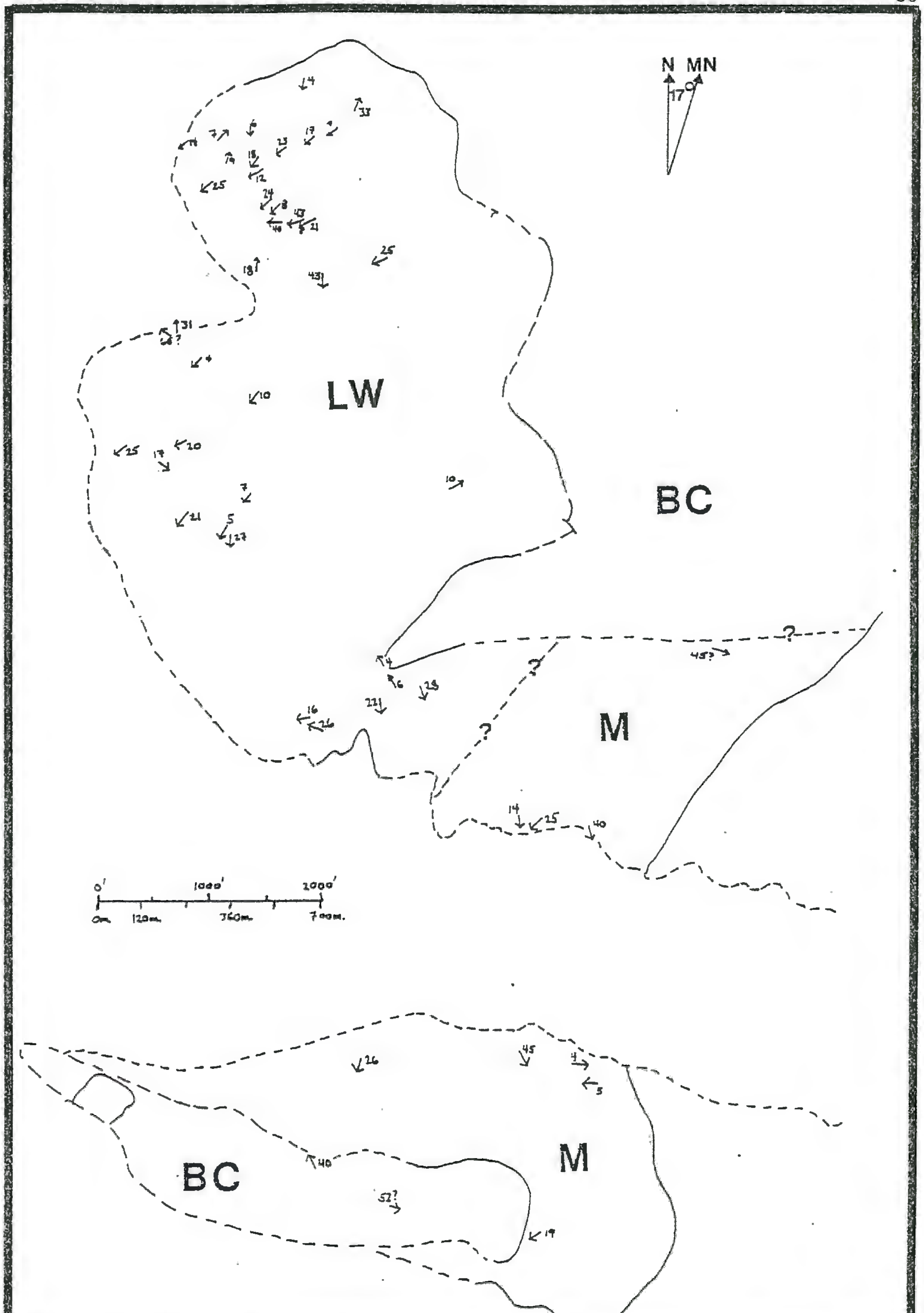

Figure 54. Lineations.

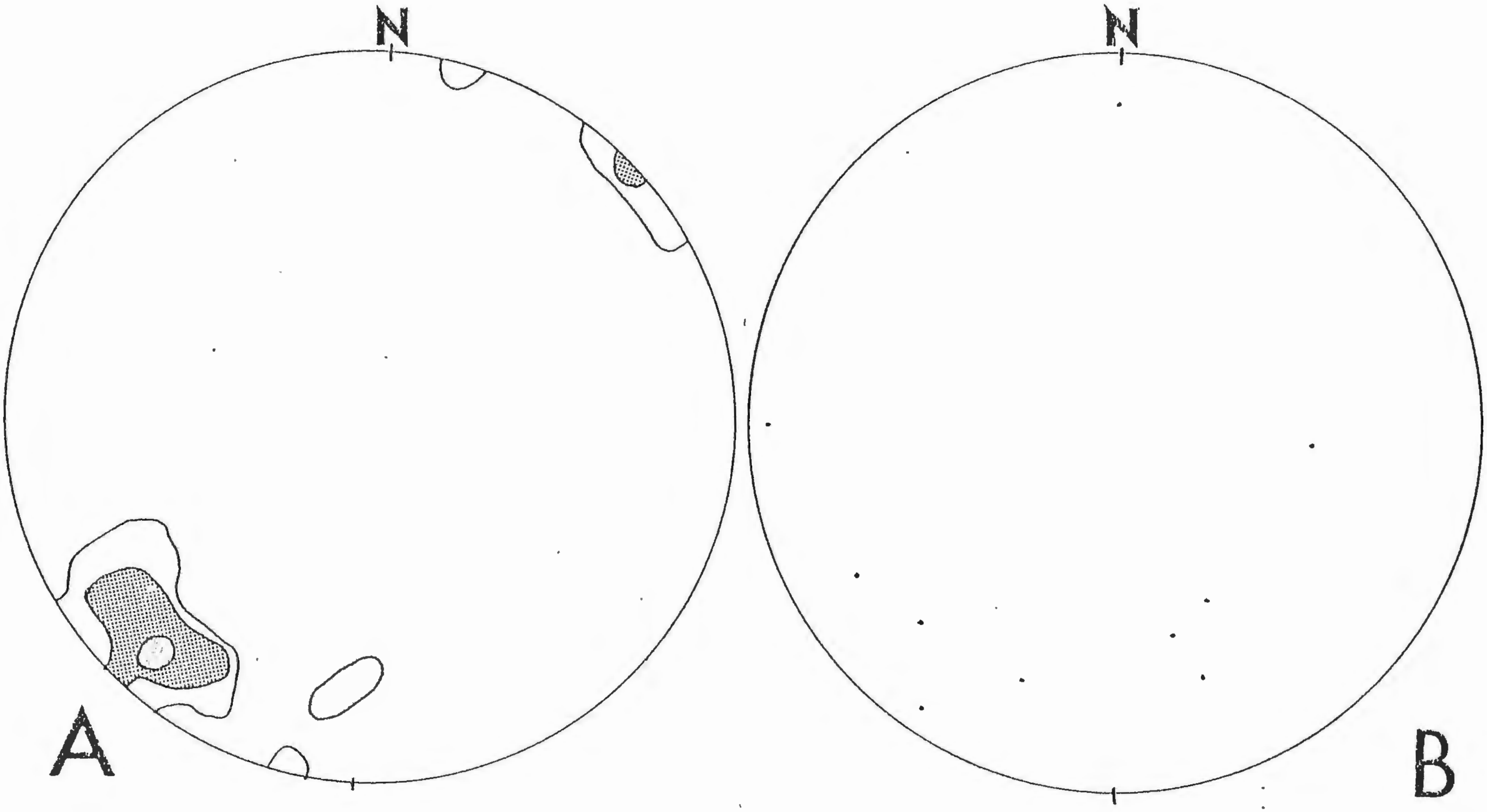

Figure 55. Lineations: A - Typical Little Willow rocks, 39 data points, contours: 20%, 10%, and 5%, average high concentration value - 224°, 10° plunge.
B - Migmatitic rocks, 10 data points.

Cataclasis is definitely gradational and sporadic. Within the same unit some rocks show little to no breakage, while others are extensively fractured.

A sequence of two cataclastic events separated by a thermal event can be ascertained from the study of the quartzofeldspathic gneisses. One rock in particular (Fig. 9, p. 18) shows an older microbrecciation which has been recrystallized (healed fractures and advanced mortar texture development). The rock was then rebrecciated, forming zones of mylonite within the mylonite gneiss.

Although more rare, cataclastic textures are seen in all the other units as well.

Megascopic evidence of faulting and shearing is also quite common in the area. Those structures that could be traced out were mapped; most were merely noted. Among the common features are small conjugate shears, slickensides and brecciation. A comparison of the orientation of shears reveals that they are quite variable.

Two major fault systems (described below) are recognized in the area. Additionally, a number of faults and zones of brecciation, shearing, chloritization or limonitization are found associated with the igneous contact.

The older faulting in the area is the thrusting which took place along the unconformable Little Willow-Big Cottonwood contact during the Laramide Orogeny. At places, this fault contact is easy to trace and pinpoint accurately; at others, extensive shearing and brecciation make this impossible. A diapthorite zone of chloritization is common along the fault.

Brecciated quartzites and chevron-folded phyllites, both Big Cottonwood rocks (Fig. 56), are two of the best indicators of the fault zone, and hence, the contact.

In the north the thrust contact dips to the east; to the south it flattens out and eventually dips to the west on the south side of Little Cottonwood Canyon. Displacement of the overlying plate was to the east, as known from regional study.

At the southwest corner of the study area, on the south ridge of Little Cottonwood Canyon, a second thrust sheet of Mississippian limestone rests on the Big Cottonwood thrust sheet. This can be explained as an imbricated slice of the main thrust sheet (Neff, 1962) or as an expression of a shingled pile of thrusts.

The Wasatch Fault Zone is an anastomosing group of normal faults which generally trend north-south on the west side of the field area. Some of these steep faults dip toward each other forming horst and graben structures, others lie en echelon, with blocks in a step-like arrangement. As this faulting is the most recent in the area, it is fairly easy to recognize the scarps in the rocks and the unconsolidated material (Fig. 49). They can usually be traced on airphotos. It has been suggested (Neff, 1962) that the several surface scarplets become one major fault at depth.

Joints

Jointing is commonly found in the Little Willow and adjacent rocks. It is common to find several joint orientations in one outcrop and this may reflect the multiple deformational history of the area. Neff (1962)

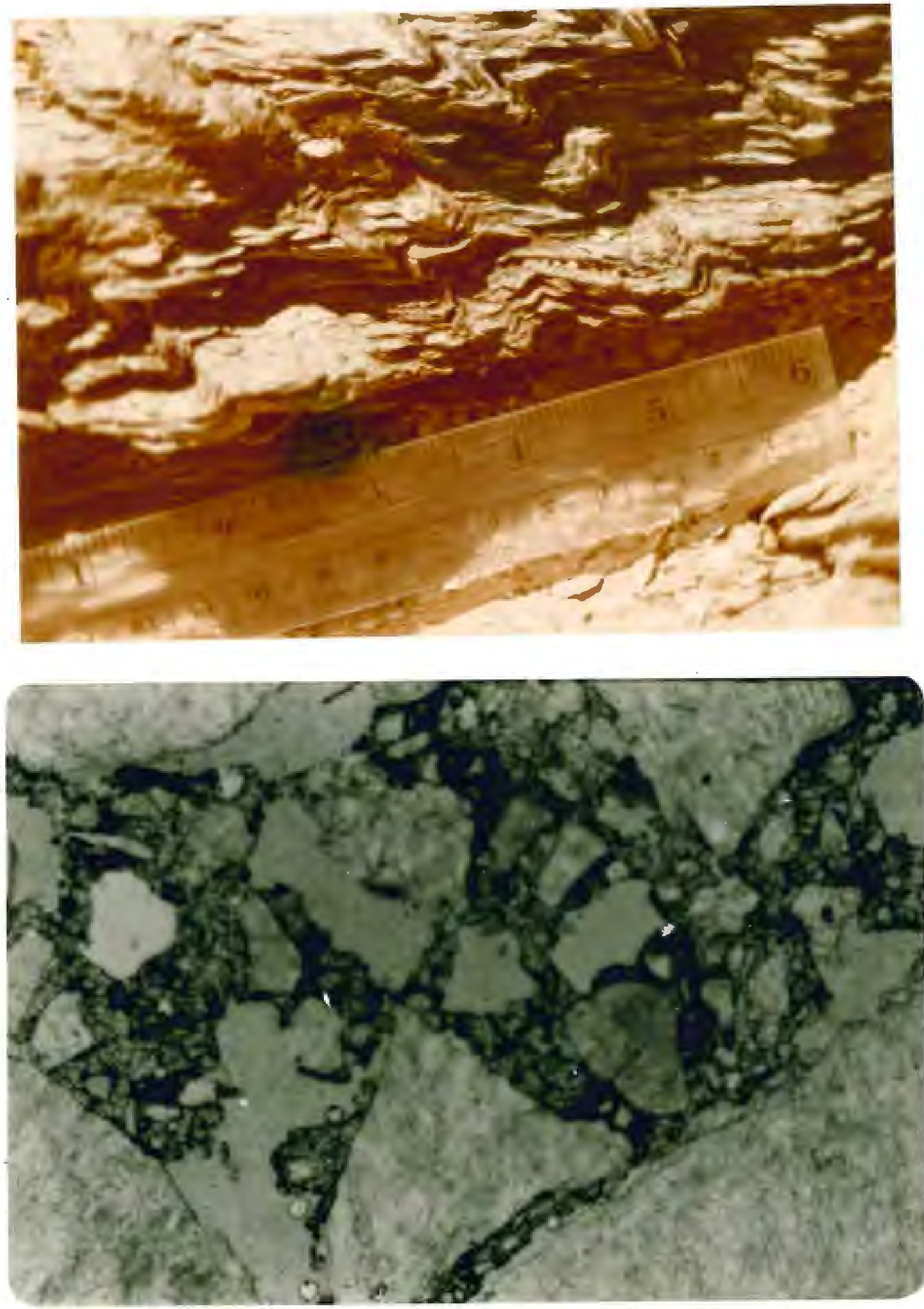

Figure 56. Rocks of the Little Willow-Big Cottonwood thrust zone contact:
A - Big Cottonwood phyllite with chevron folds near site 2;
B - Cohesive breccia of Big Cottonwood quartzite fragments, cemented by carbonate, near site 18 (width of field 6mm., ordinary light). (See Fig. 41 for hand sample picture.)

found no regional pattern in the orientation of the joints over the entire Little Willow area.

Locally some lithologies show excellent examples of conjugate fracture pairs, often rehealed or with vein fillings (Fig. 57).

Along the igneous contact, a very strong joint pattern is obvious (Fig. 58). Average values of the dominant joint orientation at ten locations near this contact were plotted (Fig. 60), but could not be interpreted.

STRUCTURAL INTERPRETATION AND CONCLUSIONS

The Little Willow area is one of complex structure. Effects of Precambrian regional deformation, Laramide thrusting, Tertiary igneous intrusion and Wasatch normal faulting can be recognized. These can be separated locally, but often the overprinting has made interpretation of the structural history difficult.

It would seem logical to attribute the main foliation (S_1) and spaced schistosity development in the Little Willow and migmatitic rocks to the earlier regional metamorphism, as these effects are absent in the later Precambrian Big Cottonwood Formation. Likewise, the evolution of crenulation (S_2) in the older rocks can be most simply explained as part of this regional event. Development of complex ductile deformation in the migmatite may also have taken place at this time.

The development of a slight phyllitic foliation, parallel to bedding in the Big Cottonwood rocks may simply be due to burial processes, while this thick sequence of sedimentary rock was lying unconformably on the older metamorphic complex.

Figure 57. (Above) Conjugate fracture fillings of plagioclase in amphibolite at site 41.

Figure 58. (Right) Extensive jointing in migmatitic rock along igneous stock contact at site 26. (See also figure 33 on page 53.)

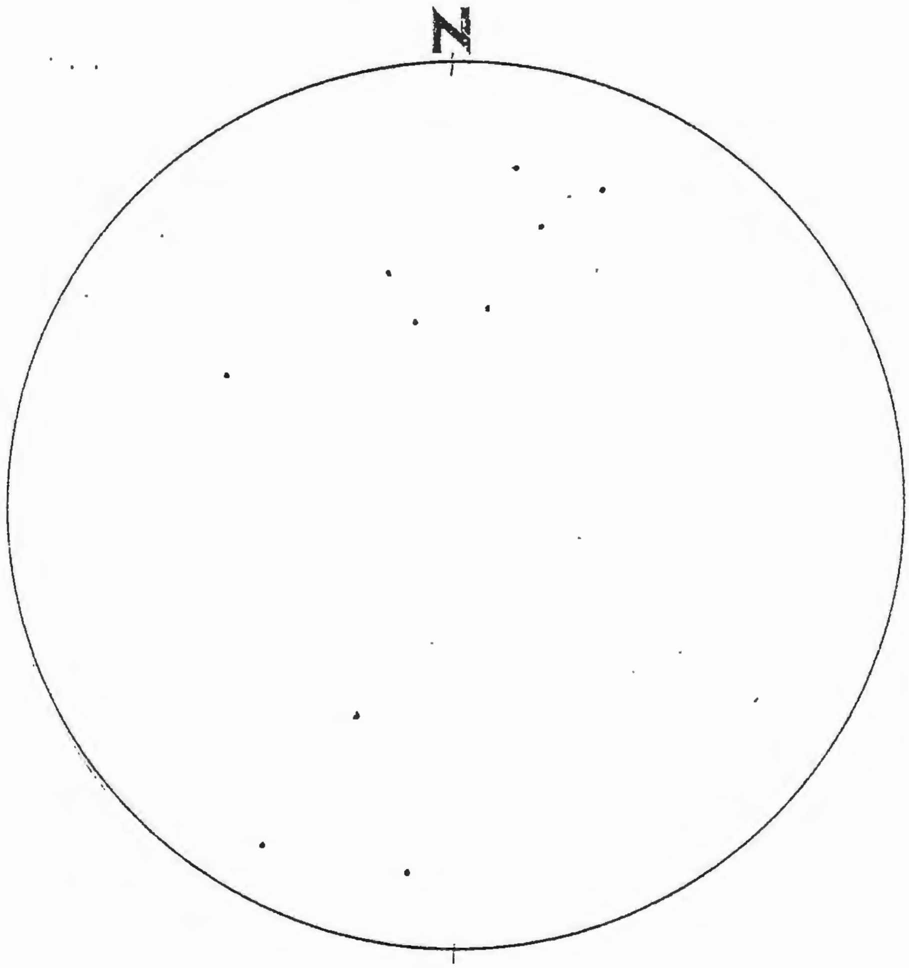

Figure 59. Dominant joint orientations near the igneous contact: 10 data points, values arrived at by taking the average of 10-15 readings at each location.

The Laramide Orogeny (thrusting) would be a plausible explanation for the first cataclasis and the thrust fault phenomena. Development of crenulations, folding and faulting in the Big Cottonwood Formation near the thrust contact probably happened at this time.

Next, the intrusion of the Little Cottonwood Stock caused faulting, shearing and jointing in the country rock near its contact, as well as provided the heat which led to the recrystallization of the first cataclastic effects. It is also geologically possible that the migmatitic metamorphism and structures were established during this event, as such phenomena have been noted in other thermal auroles (Okrusch, 1969). This does not seem to be the case, however, as migmatitic structures are not seen in the overlying Big Cottonwood rocks located equally close to the stock.

The final structural event was the normal faulting along the Wasatch Fault Zone. These faults crosscut all earlier structures and locally crushed and sheared portions of the rock.

Although not conclusive, it would seem that the Little Willow rocks and the magmatite are structurally equivalents. It would probably be more correct to associate these two units, than to correlate the migmatite and the Big Cottonwood rocks, as both the Little Willow and the migmatitic rocks share some common textural phenomena and structural trends.

THE METAMORPHISM OF THE LITTLE WILLOW AREA

GENERAL STATEMENT

There are a number of recognized regional events in the Wasatch Mountains which are pertinent to the understanding of the Little Willow metamorphic history.

A middle Precambrian regional metamorphism, prior to the deposition of the Upper Precambrian Big Cottonwood sediments, is indicated because of regional textures, e.g., foliation, developed in the Little Willow rocks, but not in the bordering Big Cottonwood rocks.

Laramide thrusting during latest Cretaceous time could be responsible for cataclastic textures found in the Little Willow rocks.

Intrusion of the Little Cottonwood stock 24-31 million years ago must have certainly caused some thermal metamorphic phenomena in the surrounding country rocks.

Wasatch normal faulting, which began approximately 20 million years ago, could also have effected cataclasis on localized zones within the Little Willow area.

A problem arises when attributing a particular metamorphic phenomenon to a particular event, as subsequent events have obscured the relationship. Close scrutiny of the textural evidence, combined with the observed mineral assemblage data, does, however, provide an indication as to the nature of the metamorphic history. For instance, foliated textures suggest regional development and decussate textures are typical of static thermal crystallization. Mineral assemblages found, especially close to the stock, should reflect the conditions of the latest thermal event; however, individual minerals could be either developed during the thermal event or preserved from the regional.

TEXTURAL ASPECTS

Textural evidence in metamorphic rocks is critical to ascertain relationships between minerals and in determining the sequence of development.

Hornblende in the amphibolite is well recrystallized and commonly foliated, lineated, or both (Fig. 14). The crystal shape is well defined, as would be expected because of the high anisotropy compared with the surrounding quartz and feldspar. The foliated and lineated nature of the hornblende suggests a regional recrystallization; however, the 560 million year old age determination (Whelan, 1969) obtained for the hornblende does not correspond to the recognized middle Precambrian regional event. This may be due to partial loss of Ar during the thermal event.

Muscovite and biotite are found in the Little Willow rocks in three distinct habits: 1) as S_1 folia, commonly crenulated (Fig. 22), 2) recrystallized parallel to the axial surfaces of the crenulations (S_2) (Fig. 24), and 3) recrystallized without preferred orientation, commonly in decussate aggregates. The third form is most common in melanosomes close to the stock, especially where muscovite is found replacing foliated sillimanite, and with biotite surrounding andalusite in the mica schist (Fig. 23). Thermal metamorphism associated with the igneous intrusion is the simplest explanation for this static recrystallization.

Helicitic euhedral andalusite in the Little Willow rocks (Fig. 23) indicates a post-S_1 growth. The prevalence of Z or S shaped inclusion patterns in a given section, as well as discontinuous internal foliation

(Fig. 24), strongly suggests rotation during growth. Inclusion of S_2 biotites in the outer portions of some of the andalusites implies continuation of growth until the onset of the S_2 event. Other andalusites, which crosscut all foliation and crenulation (Fig. 23), represent a second distinct generation of post-S_2, probably thermal andalusite. Ragged, poikiloblastic andalusites (Fig. 37), similar to those found in the Precambrian Big Cottonwood rocks (Fig. 44) which are often crosscut by sillimanite, are also attributable to the thermal metamorphism.

Folia of fibrolite weaving between individual grains of quartz and feldspar (Fig. 37) may represent a relict foliation from the regional metamorphism. Random needles of fibrolite penetrating quartz and feldspar (Fig. 39) may represent a second (thermal) generation of sillimanite: this texture is also found in the Big Cottonwood rocks close to the stock (Fig. 44).

The occurrence of sillimanite and andalusite together may indicate partial replacement of early andalusite by sillimanite (Fig. 37) or development of both simultaneously if conditions were near those for the polymorphic transition, or both.

Replacement of some sillimanite by decussate muscovite seems to be a late thermal phenomenon.

Opaque, bent inclusion trails in garnets found in the mica schists (Fig. 27) implies rotation during their post-S_1 evolution. Because the other minerals associated with the garnet imply a static environment of crystallization, the garnet may be a relict phase.

Cordierite in the northern Little Willow rocks is post-S_1 because of faint included foliation (Fig. 26). The cordierite in the garnet-cordierite rock is post-deformational, probably thermal, as evidenced by the radial configuration. Cordierite is found in the migmatite in both the leucosome and the melanosome. In the leucosome, cordierite grains are small, subhedral and relatively fresh looking (Fig. 39) and therefore a late development, possibly during anatexis, is suspected. Melanosomatic cordierites differ only in being larger and more anhedral and although an earlier growth is possible, it cannot be verified.

The majority of chlorite occurs as randomly oriented decussate grains showing no deformation (Fig. 39). This suggests thermal or late retrograde growth.

Quartz and feldspar in the Little Willow rocks exhibit a varying degree of cataclastic texture. The majority of these minerals show some signs of breakage, the best examples being fractured feldspar porphyroclasts (Fig. 10), although recyrstallization is usually quite well advanced. Recrystallization features range from subgrain polygonization (recovery) to well developed mortar texture (Fig. 11). Quartz and feldspar near the stock are usually well recrystallized, probably reflecting the thermal effects of the intrusion. Laramide thrusting may be responsible for this initial cataclasis.

Locally, evidence for rebrecciation, primarily finer grained zones of mylonite within the recrystallized mylonite gneiss (Fig. 9), can be seen. Normal faulting on the Wasatch Fault Zone is a logical cause for this localized rebrecciation.

In the Big Cottonwood rocks muscovite and biotite are commonly foliated parallel to bedding or randomly oriented. They are found crenulated only at the thrust contact (Fig. 44). Anhedral helicitic andalusite and minor sillimanite needles are believed to be thermal phases since they occur only near the stock. Chlorite clots (replacement of cordierite?) in Big Cottonwood argillites are pre-thrust, as evidenced by crenulations penetrative through them (Fig. 44).

MINERALOGICAL ASPECTS

In this section, sets of mineral assemblages, assumed to have developed near equilibrium, are analysed graphically and compared with known experimental data to determine the conditions of the late thermal metamorphism, which these assemblages should reflect. Then, removing the thermal minerals, an hypothesis is made concerning the earlier regional assemblages and what they may mean. Finally, uncertainties and problems regarding the metamorphic interpretation are outlined.

Mafic, quartzofeldspathic and pelitic compositions are found within the Little Willow rocks. Of these, the mafic and especially the pelitic rocks contain mineralogical information concerning the metamorphisms.

Mafic Compositions

The amphibolites within the northern Little Willow Formation are composed of the general mineral assemblage: hornblende-plagioclase-epidote±biotite±quartz±garnet, excluding chlorite and sphene since they are thought to be secondary. This assemblage is indicative of the amphibolite facies of metamorphism and suggests moderate temperatures and pressures (Vernon, 1975). The association of pelitic and quartzo-feldspathic schists and gneisses is common and expected at this grade of metamorphism (Turner, 1968).

Pelitic Compositions

Assemblages found in the pelitic rocks are listed in Table 1 (p. 101). Comparison of these assemblages, with the aid of sample locations (Plate 2), permits two isograds to be drawn within the field area (Fig. 60). The first appearance of sillimanite (the sillimanite isograd) is used to determine the lower limit of the sillimanite zone. The isograd cannot be located exactly in this study because of the scarcity of outcrop and the heterogeneity of the rock. Little Willow rocks northwest of the sillimanite isograd may contain andalusite, but not sillimanite; to the southeast of the isograd both andalusite and sillimanite are commonly present.

Closer in to the stock the second sillimanite isograd can be approximately positioned (Fig. 60). It separates higher-grade rocks in which the critical assemblage sillimanite and K-feldspar is found (see page 105).

Thompson (1957) developed a graphical analysis for metamorphosed pelites which shows the relations between compositions and mineralogies of pelitic rocks.

This method removes the ambiguity of the isograd system due to variations in bulk composition and provides a much more detailed view of the mineralogic changes involved in equilibrium metamorphism of pelitic rocks (Green, 1963). Thompson's projection reduces the significant components to three: Al_2O_3, MgO and FeO (hence, the name AFM projection), via the following restrictions and assumptions:

1) quartz is omnipresent, so the SiO_2 component is removed;

2) the analysis is projected from muscovite or K-feldspar;

SITE	MAP UNIT	ASSEMBLAGE +Q	POSSIBLE MINI-FACIES
1	LW	C+B+M+(H)	V,IV
		A+B+M+(H)	I,V
		C+A+M+(H)	III,IV,V
5	LW	B+G+M+(H)	ANY
6	LW	B+C+M	IV,V
8	LW	A+B+M+(H)	I,V
10	LW	A+B+M	I,V
12	LW	A+B+M	I,V
13	LW	B+M	ANY
15	LW	A+B+M+(H)	I,V
		G+C+B+H	–
		C+B+H+(M)	–
		G+C+H	–
21	LW	A+B+M+(H)	I,V
22	LW	A+B+M	I,V
44	LW	A+B+C+M+(H)	V
17	MG	S+M+B+(H)	I,V
27	MG	A+S+C+B+M+(H)	V
28	MG	B+M	ANY
		S+B+M+(H)	I,V
30	MG	S+B+M	I,V
31	MG	S+B+C+K+(M)+(H)	V
36	MG	S+K+(M)	ANY
		S+A+K+B+(M)+(H)	I,V
37	MG	S+B+K+(M)	I,V
		S+K+B+C+(M)	V
40	MG	A+B+K+C+(M)	V
		S+K+(M)	I,V
		A+S+B+C+K+(M)	V
43	MG	A+S+K+M+C+B	V
		A+S+B+K+M	I,V
		S+A+B+M	I,V
45	MG	S+A+K+B+(M)	I,V
		S+C+B+(M)	V
59N	MG	K+P+B+S+A+C+(M)	V
14	BC	M+(H)	ANY
2	BC	M+A+(H)	ANY
39	BC	M+A+S+B+(H)	I,V
40	BC	M+A+S+B+(H)	I,V

TABLE 1. PELITIC ASSEMBLAGES FOUND IN THE LITTLE WILLOW AREA

(Lw=Little Willow, MG=migmatite, BC=Big Cottonwood, A=andalusite, B=biotite, C=cordierite, G=garnet, H=chlorite, M=muscovite, K=K-feldspar, P=plagioclase, S=sillimanite, Q=quartz, ()=secondary phase. PT conditions of the mini-facies are given in Figure 62.

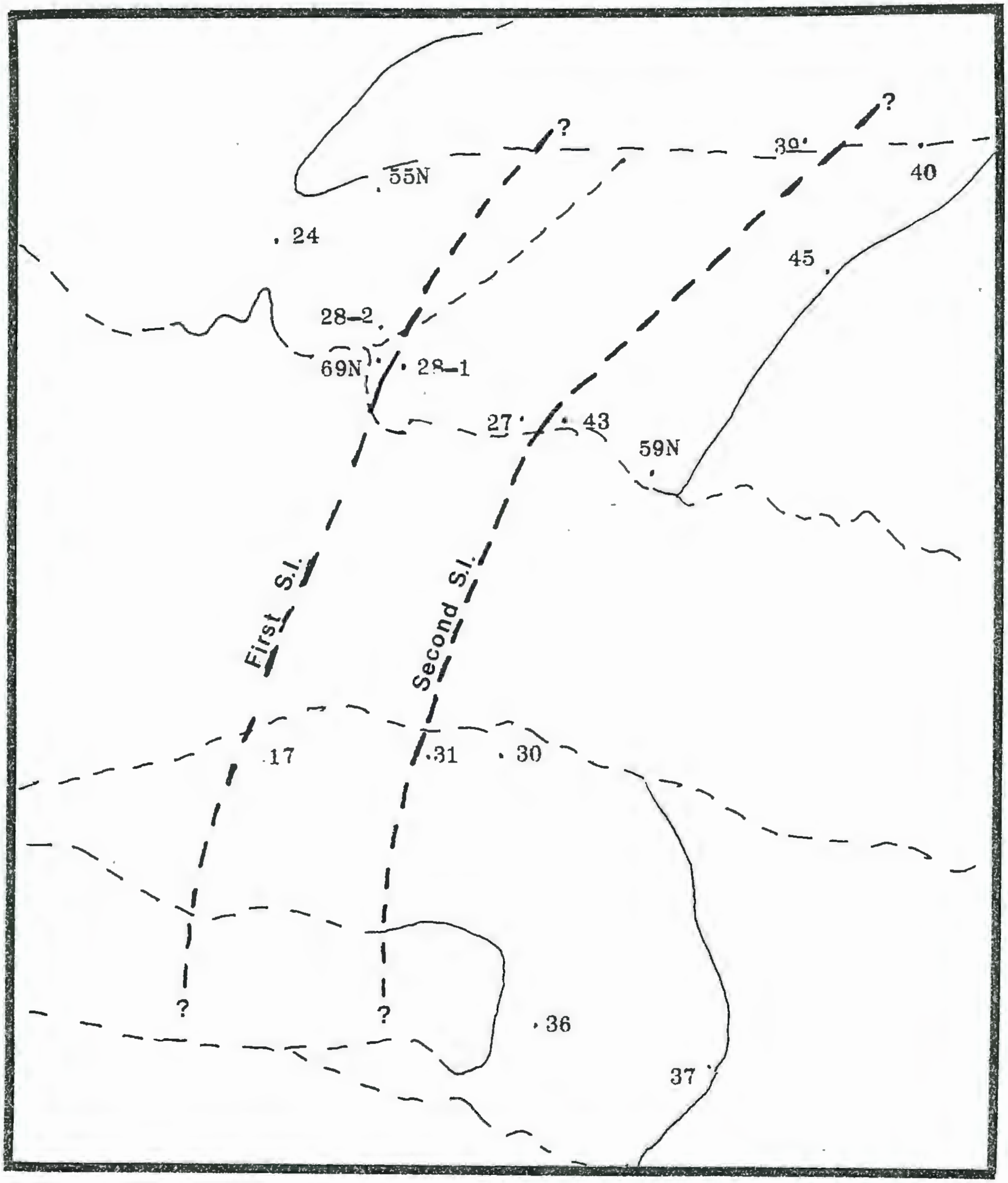

Figure 60. Sillimanite isograd map. Approximate locations of the first (andalusite→sillimanite) and the second (quartz+muscovite →K-feldspar+aluminosilicate+vapor) sillimanite isograds in the southeastern portion of the Little Willow area. Numbers are sample locations. See Table 1 for assemblage data.

3) trace components are omitted; and

4) H_2O is externally controlled.

Gibbs' phase rule ($f=c+2-p$) states that the maximum number of phases ($c+2$) exists when the variance (f) is zero and thus the temperature and pressure are fixed. However, if pressure and temperature have arbitrary values, the variance is at least two and $p \leq c$ (cf. Goldschmidt, 1911). Thus, under divarient conditions the maximum number of phases in an equilibrium assemblage shown in a Thompson projection is three.

By applying Schreinemaker's (1911) analysis to assemblages conforming to Thompson's model, a qualitative petrogenetic grid for pelitic rocks can be developed and made quantitative by application of experimental data, as shown in Fig. 61.

The most common assemblages in the quartz and muscovite bearing Little Willow rocks north of the first sillimanite isograd are the two-phase assemblages andalusite-biotite and cordierite-biotite and the three phase assemblage andalusite-biotite-cordierite is also found (Table 1). This last assemblage is only possible in mini-facies V on Figure 61. This facies is limited by the following univariant reactions:

1) andalusite+biotite+cordierite=muscovite+chlorite

2) andalusite+biotite=muscovite+cordierite+garnet

3) quartz+muscovite=K-feldspar+aluminosilicate+vapor

The assemblage garnet-cordierite-biotite-chlorite cannot be shown in the Thompson diagram because the rock contains no muscovite. In the absence of muscovite, this assemblage can also be stable under the conditions for mini-facies V.

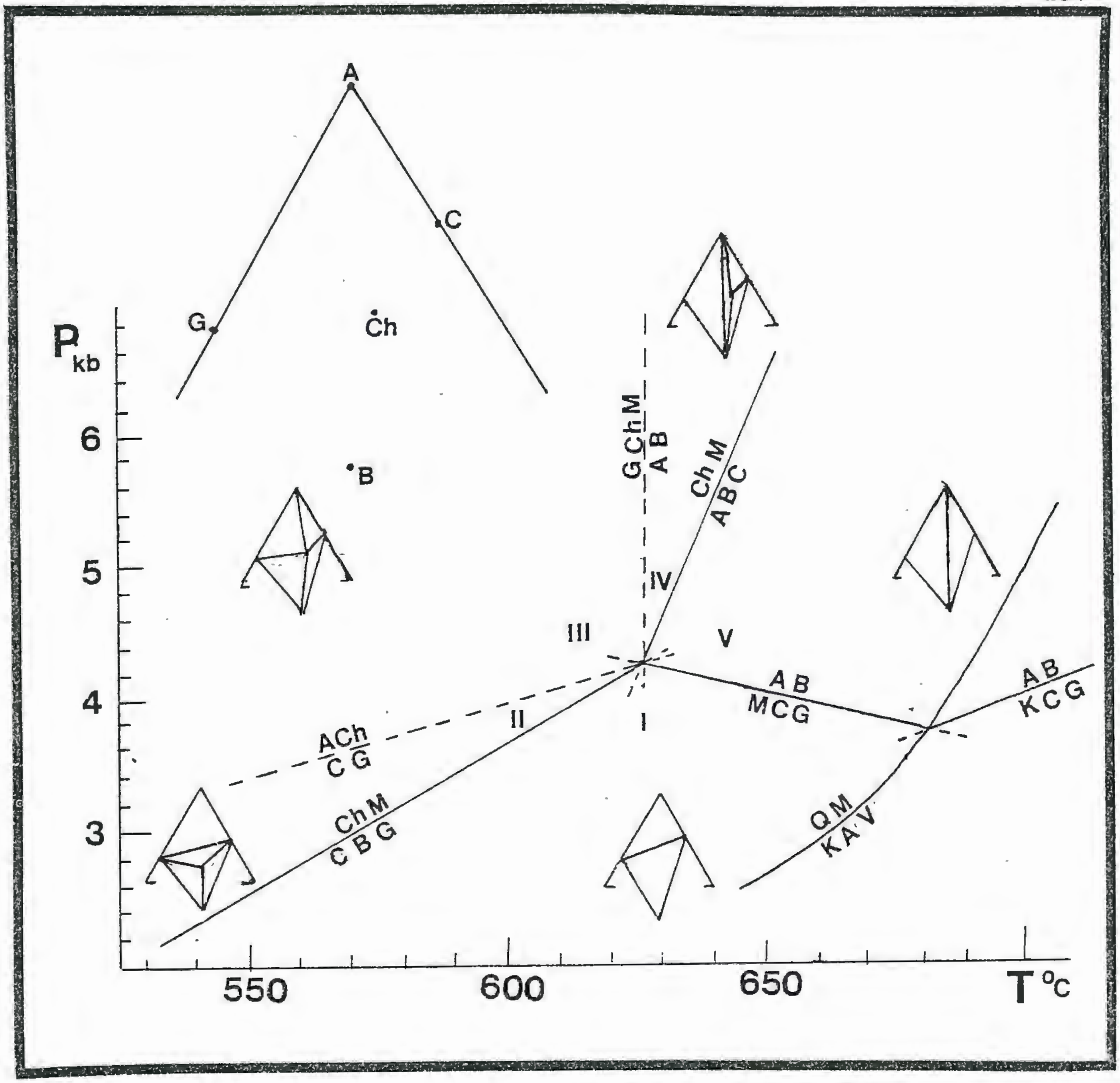

Figure 61. Petrogenetic grid for quartz-bearing pelites.

The reactions (Ch+G=A+B and (Ch+A=C+G) are shown qualitatively. The invariant assemblage (ABCChG) is metastable with respect to staurolite-bearing assemblages. A=aluminosilicate, B=biotite, C=cordierite, Ch=chlorite, G=garnet, K=K-feldspar, M=muscovite, Q=quartz, V=vapor. (adapted from Thompson, 1976)

Rocks within the sillimanite-muscovite zone have assemblages similar to those found to the northwest, differing only in the addition of sillimanite (Table 1). Thus, they also conform to mini-facies V (Fig. 61), with the general divariant three-phase assemblage of aluminosilicate-cordierite-biotite, plus quartz and muscovite.

The appearance of sillimanite is important in that it implies that the metamorphism must have taken place at temperatures greater than 625°C for pressures less than 5 kb; in general, to the right of the andalusite-sillimanite reaction line (Fig. 62).

Since the reaction :

4) andalusite $\rightleftharpoons$ sillimanite

involves only a simple polymorphic transition, these two minerals theoretically should never be found in the same rock at random pressure and temperature. However, it is well known that due to the low Gibbs free energy involved, the transition between andalusite and sillimanite is quite sluggish. Thus, in nature it is not uncommon to find relict andalusite and new sillimanite together, and precise experimental determination of the transition is notoriously difficult.

Close to the igneous contact some of the rocks lack muscovite and in others the only muscovite is secondary. K-feldsapr, on the other hand, is not uncommon and the critical assemblage quartz-K-feldspar-sillimanite is found. This suggests that these rocks underwent a reaction of the form:

5) muscovite+quartz=K-feldspar+aluminosilicate+vapor

which is commonly referred to as the second sillimanite isograd (Evans and Guidotti, 1966). This reaction, and those for the Al_2SiO_5 polymorphic transitions are shown on Figure 62.

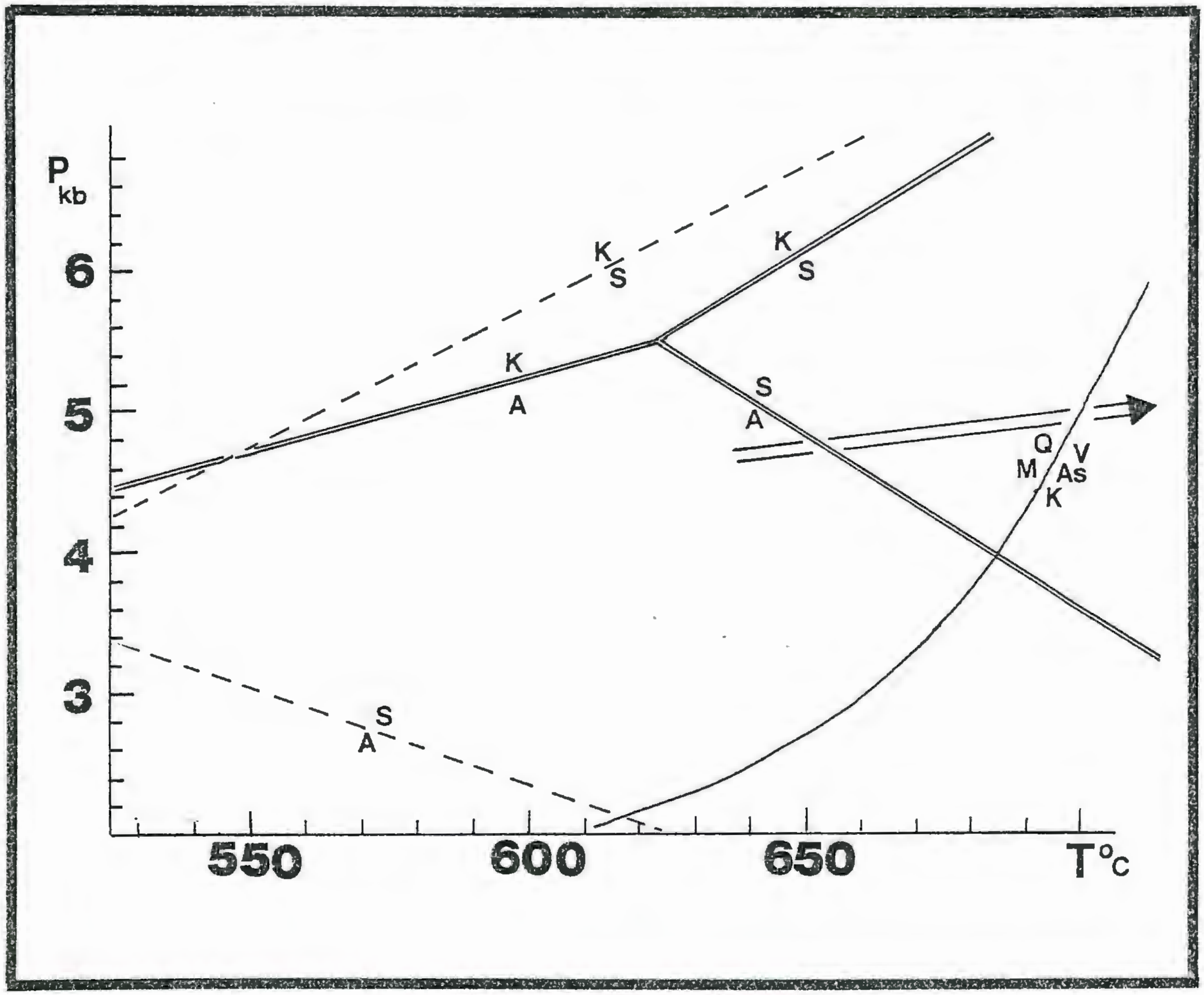

Figure 62. Petrogenetic grid. Shown are the Al_2SiO_5 relations of Holdaway, 1971 (dotted line), and of Richardson, Gilbert and Bell, 1969 (double line). Also shown is the Q+M=K+Al+V reaction line of Chatterjee and Johannes, 1974. A=andalusite, K=kyanite, S=sillimanite, Q=quartz, M=muscovite, K=K-feldspar, As=alumino-silicate and V=vapor. The arrow represents a possible metamorphic gradient for the Little Willow rocks.

Traversing into higher-grade rocks in this area, the first sillimanite isograd, producing sillimanite-muscovite, is encountered first, and then the second, yielding sillimanite-K-feldspar. This indicates a PT gradient as shown by the arrow on Figure 62 (around 4 Kb, 625-650^{o}C). If the pressures were higher, kyanite-muscovite would be expected; if lower, andalusite-K-feldspar.

Again, because of the rarity of exposures of suitable pelitic compositions across the area, placement of the second sillimanite isograd is approximate.

The conditions set forth above are adequate to initiate anatexis (partial melting) in rocks with near granitic compositions (Winkler, 1967; Grant, 1973), thus providing a possible explanation for the development of the leucosomatic envelopes. But, the structures of the migmatite are compatible with those of the Little Willow to the northwest, rather than with those of the Big Cottonwood Formation, which lacks evidence of anatexis even adjacent to the stock. So anatexis here was probably associated with the regional event. The sporadic occurrence of the leucosome can be explained by the heterogeneous composition of the rock, as initial melting temperatures are drastically dependent on bulk composition.

The conditions of the regional metamorphism are more difficlut to assess, especially since what may have been the higher grade regional portion is also the higher grade portion of the thermal aureole.

In the northwestern Little Willow, the foliated amphibolites indicate amphibolite facies metamorphism (moderate temperatures and pressures) for the regional event.

These conditions are further substantiated by regional assemblages in the pelitic rocks. As indicated by their textures, there is little doubt that andalusite, biotite, garnet and probably cordierite existed as regional minerals, along with quartz and muscovite. Referring again to Figure 61, this set of minerals is compatible with mini-facies V once again.

Foliated sillimanite found in the migmatites is readily explained by the regional event; however, development of foliated sillimanite has been reported from other thermal metamorphisms (e.g., Okrusch, 1969). Thus it is uncertain whether there are indeed two generations of sillimanite.

As noted above, the regional event is considered responsible for the development of the migmatitic structures and the leucosomes.

Relatively few samples of the Big Cottonwood pelites were studied. However, randomly oriented needles of sillimanite and anhedral, spongy andalusite seem to be thermal because of their textures and because of their close proximity to the stock contact. The close association of the two minerals suggests conditions close to the andalusite-sillimanite transition, which agrees with the observations in the Little Willow rocks. Removing these thermal effects, a low-grade biotite-chlorite-muscovite-quartz assemblage can be established for the lowermost units of the Big Cottonwood Formation.

Problems and uncertainties remain regarding the metamorphic history and interpretation of the Little Willow area. A most obvious concern is the exact relationships which exist at the hidden contact, as well as the accurate placement of the isograd and reaction lines for the thermal metamorphism. Some doubt exists for some minerals as to which were newly formed during the thermal metamorphism, which were merely preserved from the regional event and which developed during both. Likewise, some question remains as to if the migmatitzation and anatextic features are definitely regional metamorphic phenomena. The seemingly anamalous situation in the migmatitzation, in which typically competent quartzose units apparently deformed ductilly, also deserves further consideration. Finally, there remain minor discrepancies such as the ambiguous radiometric date for hornblende in the amphibolite and the uncertain petrogenetic relationship between andalusite and sillimanite.

SUMMARY AND CONCLUSIONS

A major concern of this thesis is the determination of the nature of the questionable migmatitic unit, flanking the Little Cottonwood Stock, referred to both the Big Cottonwood Formation and the Little Willow Formation in the past. Both the migmatite and the Big Cottonwood are intruded by the Little Cottonwood Stock, which has thermally metamorphosed these rocks. An unconformable-thrust fault contact exists between the Little Willow and the Big Cottonwood formations, as well as between the Big Cottonwood and the migmatite on the south side of Little Cottonwood Canyon. The remaining contacts between the Big Cottonwood and the migmatite and the migmatite and Little Willow are not observed because they are obscured by overburden.

The Little Willow Formation to the northwest is a metamorphic complex comprised of quartzofeldspathic gneisses, pelitic mica schists and amphibolites. The quartzofeldspathic gneiss is a grey to tan, fine-grained, granoblastic rock with local lenses of biotite rich schist. The rock is composed primarily of plagioclase with lesser amounts of quartz and biotite and minor K-feldspar, muscovite and chlorite. The rock exhibits cataclastic textures with advanced recrystallization.

The mica schist unit is a composite of several interfingering lensoidal compositions, including an andalusite-biotite schist, a garnet-cordierite rock and a quartz-sericite schist, all within a predominantly pelitic muscovite schist. A pelitic metaconglomerate is also found within the unit and serves as a marker bed. This entire unit appears to be a larger scale equivalent of the schistose biotite lenses found within the gneiss. The schists are medium-grained with development of foliation

(S_1), spaced schistosity, and prominent crenulation (S_2). Porphyroblasts of cordierite, garnet and andalusite are locally common. A regional metamorphic mineral assemblage of cordierite, rotated garnet, rotated andalusite, biotite, muscovite and quartz is recognized and implies metamorphism of moderate temperatures and low to moderate pressures. Effects of a later thermal metamorphism are also seen, such as a second, static generation of andalusite and randomly oriented recrystallized micas.

The amphibolite, which is probably of igneous origin, is a dark green, fine- to medium-grained rock which can be traced through the quartzofeldspathic gneiss. The rock is moderately foliated (S_1) and in places lineated. It consists of mainly hornblende with lesser amounts of plagioclase and minor quartz, epidote and biotite. This foliated assemblage also indicates moderate conditions (amphibolite facies) for the regional metamorphic event.

Foliations in the forementioned Little Willow Formation are homoclinal with an average strike of North 25° East with an average dip of 40° Northwest. Lineations, defined primarily by crenulation axes ($S_1 \cap S_2$), are concentrated at 210-240° with an average plunge of 10°.

The Big Cottonwood Formation which unconformably overlies the Little Willow Formation consists of alternating quartzites and argillites of unquestionable sedimentary origin. The limonitic to buff colored quartzites are commonly in beds up to 2 m thick, although some are massive. They are fine-grained and slightly recrystallized. The argillite occurs as 2-30 cm thick beds of interlayered fine-grained arenaceous pods and fine shaly layers. These rocks commonly contain chlorite clots and have a slight phyllitic foliation developed parallel to bedding. This foliation has an average strike of North 10^{o} West with an average dip of 55^{o} Northeast. Except for a few chevron fold and crenulation axes observed at the thrust contact, lineations were not found in the Big Cottonwood Formation. Except for thermal andalusite and sillimanite found close to the igneous contact, the rocks are characterized by the low grade mineral assemblage.

The migmatitic gneiss, located southeast of the main Little Willow Formation, is a medium-grained, heterogeneous metamorphic unit comprised of primarily quartzites and quartzofeldspathic rocks. Within these are dark colored rafts and boudinaged layers composed of sillimanite, andalusite, biotite, cordierite, with minor muscovite, quartz and feldspar, which are characterized by crenulated foliation and spaced schistosity. This regional texture is obscured by randomly oriented recrystallized minerals. Many of these dark portions are surrounded by leucosomatic envelopes of possible anatextic material. Migmatite S_1 foliations are centered on a strike of North 20^{o} East with an average dip of 60^{o} Northwest. A few variable lineations, e.g., isoclinal fold axes, are observed. The first and second sillimanite isograds can be approximately located within this unit. The isograd relationships and the associated critical mineral assemblages, when compared with experimental data, suggest metamorphic pressures of about 4 kb and temperatures in excess of $600^{o}C$. Although these assemblages reflect the thermal conditions, similar assemblages and conditions are indicated for the earlier regional metamorphism.

The conclusion of this thesis is that the migmatite should be considered part of the Little Willow Formation for the following reasons:

1) Both the migmatite and the Little Willow are characterized by crenulated foliation and spaced schistosity. The Big Cottonwood has only phyllitic foliation; crenulation is found only in the vicinity of the thrust zone;

2) Foliation in both the Little Willow and the migmatite exhibit the same approximate orientation, whereas the Big Cottonwood rocks show only a slight foliation subparallel to bedding which has a distinctly different orientation;

3) Both the migmatite and the Little Willow are medium to coarse grained rocks with granoblastic textures, whereas the Big Cottonwood, although locally recrystallized, is basically a fine-grained rock in which individual quartz grains can still be discerned within a matrix in the argillaceous porttions; and

4) Aside from thermal minerals, the Big Cottonwood Formation is characterized by a low-grade metamorphic mineral assemblage, whereas the Little Willow and the migmatite have higher grade pelitic metamorphic minerals.

It has been suggested that the migmatite is the equivalent of the Big Cottonwood, subjected to the contact metamorphism. This is unlikely as Big Cottonwood rocks found elsewhere near the stock contact do not display migmatitic characteristics such as schollen structures, leucosomatic veins and rinds and foliated sillimanite.

It is surmised that the migmatite is the equivalent of the Little Willow and that if the migmatitic structures and foliated sillimanite are indeed regional in origin, they represent higher grade metamorphism. It is also possible that the migmatite is yet an older metamorphic complex and that an unconformity or fault contact exists between it and the Little Willow, although no evidence was found for this.

Crittenden (1965) redefined the Little Willow Formation "to include all of the older metamorphic rocks of the Little Willow area" (James, 1979). The migmatite is the highest grade metamorphic rock of those studied. It has as long and complex of a history as any rock in the area and therefore should be part of the Little Willow Formation.

From the petrologic, structural and field studies undertaken in the Little Willow area, a comprehensive history can be deduced. This chronology includes the following events from oldest to youngest. Details of the regional events are referred to in the Central Wasatch Geology chapter (p. 15):

1) Middle Precambrian(?) - Deposition of what was to become the Little Willow Formation, including pelitic sediments, quartz sandstones, conglomerates, and probable felsic tuffs, with mafic dikes or flows.

2) Middle Precambrian(?) - Regional metamorphism of these rocks to amphibolite facies, prior to the deposition of the Big Cottonwood Formation. Characteristics include crenulated foliation, spaced schistosity and growth of the first generation of andalusite. Migmatization and foliated sillimanite found in the southeastern portion of the Little Willow are most readily explained by this event.

3) Middle-Late Precambrian - Uplift and erosion of these rocks forming a peneplain surface.

4) Late Precambrian - Transgression of the Beltian Sea and deposi-

tion of more than 16,000 feet of Big Cottonwood quartz sandstones and shales unconformably on the Little Willow.

5) Late Cretaceous - Development and uplift of the Cottonwood Arch.

6) Latest Cretaceous-Early Paleocene - Thrusting along the unconformity between the Little Willow Formation and Big Cottonwood rocks, related to the Laramide Orogeny, and resulting in an early cataclastic texture.

7) Oligocene - Intrusion of the Little Cottonwood quartz monzonite stock, effecting thermal metamorphism on the surrounding rocks. Pressures of about 4 kb and temperatures in excess of 600^{o}C appear to have preserved some of the regional characteristics and developed some new ones. Features of the contact metamorphism include recrystallization of the Laramide cataclasis, growth of a second generation of andalusite, growth of sillimanite (possibly second generation), and replacement of foliated sillimanite by muscovite. In general, the mineral assemblages seen today in the rocks, except for late retrograde phases, reflect the conditions of this thermal event.

8) Miocene-Recent - Uplift and exposure of the Little Willow Formation due to normal block faulting along the Wasatch Fault Zone. Localized zones of rebrecciated mylonite in the mylonite gneiss was the main textural evidence.

9) Pleistocene - Glaciation and the development of pluvial lakes.

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