STRUCTURAL ANALYSIS OF DEFORMED CARBONIFEROUS STRATA, MISPEC BEACH, SOUTHERN NEW BRUNSWICK

by

Lauren Magin Bradley

A Thesis Submitted to the Faculty
of the State University of New York
at Albany
In Partial Fulfillment of the Requirements
for the Degree of
Master of Science

College of Science and Mathematics
Department of Geological Sciences

1984

STRUCTURAL ANALYSIS OF DEFORMED CARBONIFEROUS STRATA, MISPEC BEACH, SOUTHERN NEW BRUNSWICK

by

Lauren Magin Bradley

Abstract of

A Thesis Submitted to the Faculty

of the State University of New York

at Albany

In Partial Fulfillment of the Requirements

for the Degree of

Master of Science

College of Science and Mathematics

Department of Geological Sciences

1984

ABSTRACT

Mispec Beach, New Brunswick is a coastal exposure of deformed Carboniferous sedimentary rocks. Late Paleozoic deformation in the region is thought to have occured in a restraining bend in the Cobiquid - Chedabucto fault. Within this exposure, two main generations of deformation are recognized. The first deformation is associated with slaty cleavage development, pebble elongation, and the formation of quartz veins. The second generation structure is a moderately developed crenulation cleavage. The diversity of structures present at this outcrop allows for a qualitative determination of the principal axes of stress and strain associated with the first generation structures.

ACKNOWLEDGMENTS

This project originated out of work being done in the region by Dwight Bradley. It was he who recognised Mispec Beach as a potentially rewarding field area. Win Means, Janos Urai and Bill Kidd served as formal advisers. Dwight Bradley assisted with mapping during the summer of 1982. Janos Urai has been a great help in interpreting the field data and in understanding the significance of the project. Fellow graduate students Paul Mann, Andy Bobyarchick, Bruce Idleman, Steve Tanski, Jane Kozinski, Peter Hall, Tom Ray, and Mark Jessell each made valuable contributions toward the completion of the thesis. I wish to thank all of the above people for their efforts on my behalf.

I also owe thanks to my mother Jean Ryan, and my family and friends for their encouragement throughout the completion of this project. My gratitude to my husband, Dwight Bradley, is beyond words, for without his support, encouragment and help this thesis would not be done.

This research was funded in part by grants from the SUNYA Benevolent Society (1982) and Sigma Xi (1982).

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

LIST OF PLATES

CHAPTER 1

INTRODUCTION

There are two large areas in the Northern Appalachians where late Paleozoic (post-Acadian) rocks underwent penetrative deformation during the destrution of Carboniferous sedimentary basins.

The Narragansett Basin in Rhode Island was deformed during the Permian Alleghanian Orogeny. Rocks within the basin underwent several phases of deformation, and metamorphism locally reached sillimanite grade. Although many aspects of its deformation have been studied (e.g. Mosher 1983), the tectonic setting of the basin and the relationships between Alleghanian structures and regional tectonics are not well established.

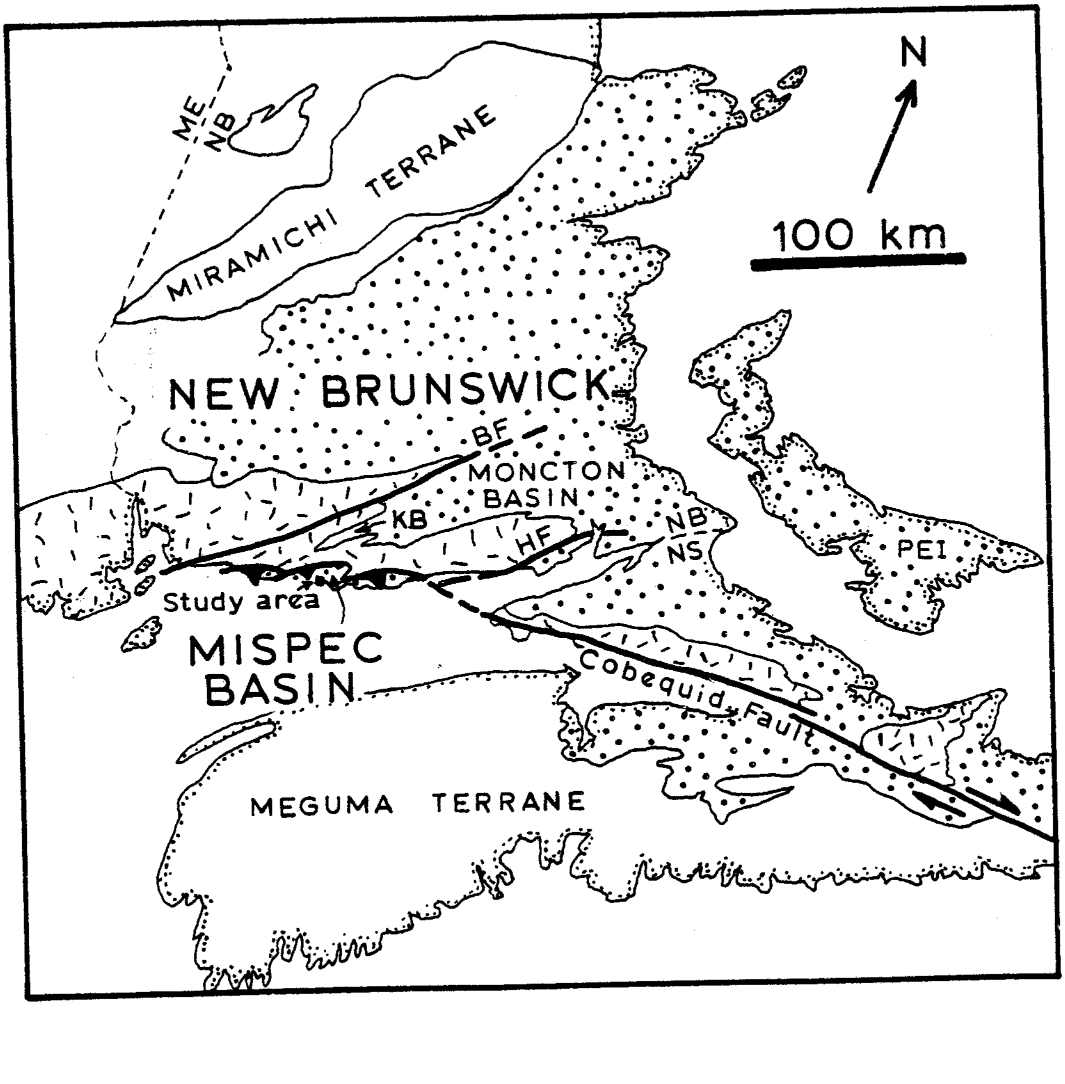
The second zone of late Paleozoic penetrative deformation with many similar characteristics is the Mispec Basin in southern New Brunswick. Here, the Carboniferous sedimentary rocks underwent polyphase deformation, probably in late Carboniferous. Metamorphism locally reached garnet grade.

What sets the Mispec and Narragansett Basins apart is that the Alleghanian Orogeny in the Mispec Basin can be confidently related to strike slip faulting. The basin is situated at what appears to be a restraining bend in the major east-trending Cobequid - Chedabucto fault system (fig. 1.1). But while the regional tectonic setting is reasonably well understood, very little detailed structural geology has been done.

This study focuses on a single outcrop within the Mispec Basin on the coast of New Brunswick. The outcrop was mapped at a scale of 1:50 in an effort to determine the structural style and orientation.

Although this is a small study area, the structures mapped may be typical of strike slip restraining bend deformation. In other zones of transpression that have been studied, as in southern California (Sylvester and Smith 1977) and the mountains of Hispaniola (Mann 1984), only the upper structural levels are exposed. In southern New Brunswick, we may be seeing deeper structural levels in the same tectonic environment.

Figure 1.1. - Location map of New Brunswick, Northern Appalachians, showing major structural features mentioned in text. Random dashes = pre-Carboniferous basement areas assigned to the Avalon Terrane; coarse stipple = Carboniferous sedimentary rocks; KB = Kennebecasis Subbasin; BF = Belleisle fault; HF = Harvey - Hopewell fault. Thrusts in the Mispec Basin are shown schematically. Geology adapted from Williams (1979).



CHAPTER 2

GEOLOGIC SETTING

GEOLOGY OF THE AVALON TERRANE, SOUTHERN NEW BRUNSWICK

New Brunswick is located in the Northern Appalachians, midway between New York and Newfoundland. Within the New Brunswick Appalachians, two zones of older basement rocks are recognized. The Miramichi Terrane, in the northwestern part of the Province, is widely interpreted as an Ordovician island arc that collided with North America during the Taconic Orogeny (e.g. Rast and Stringer 1980). The Avalon Terrane, along the southeastern coast of New Brunswick, is regarded as an island arc underlain by continental crust, which collided with the Miramichi Terrane (by then part of North America) in Devonian times, resulting in the Acadian Orogeny (e.g. D. Bradley, 1983). This study is concerned with deformation resulting from extensive strike slip faulting within the Avalon zone, following the Acadian Orogeny. The stratigraphy of southern New Brunswick is summarized in table 2.1, and a more detailed description of the succession in the area south of the Belleisle fault follows.

BROOKVILLE GNEISS. - The oldest rocks exposed in southern New Brunswick have traditionally been thought to be the Proterozoic Greenhead Group (Alcock 1938; Rodgers 1970).

TABLE 2.1. - Major rock units of the Avalon Terrane, St. John area, southern New Brunswick.

UNIT	AGE	LITHOLOGY
Lepreau Fm.	Triassic	Coarse, nonmarine clastic sedimentary rocks
Lancaster Fm.	Wesphalian B (Pennsylvanian)	Nonmarine sandstone
Mispec Group (West Beach and Ball Lake Fms.)	Early Carboniferous (?)	Coarse, nonmarine clastic sedimentary rocks
Small plutons	Devonian	Granite
Unnamed	M. Ordovician thru Silurian (?)	Marine clastics
St. John Group	Cambrian thru M. Ordovician	Marine clastics and limestones, overlain by turbidites
Coldbrook Group	Late Proterozoic (750 ± 80 Ma)	Mainly volcanics and metavolcanics; minor sedimentary rocks
Greenhead Group	Neohelikian (1000-1200 Ma)	Shelf carbonates and clastics, metamorphosed in part
Brookville Gneiss	Mid-Proterozoic (1641 ± 60 Ma)	Quartzo-feldspathic gneiss

More recent work has suggested the presence of an older unit, the Brookville Gneiss, exposed near Saint John (Currie et al. 1981; Olszewski and Gaudette 1982). This unit consists of quartz-feldspar-biotite and hornblende-quartz-feldspar-biotite gneisses (Olszewski and Gaudette 1982). Zircons from the Brookville Gneiss have yielded a uranium-lead age of 1641 $\pm$ 60 Ma (Olszewski and Gaudette 1982); this is considered to be the age of the source of the zircons in the Brookville Gneiss. It is possible that the Brookville Gneiss represents the basement flooring the Greenhead Group. Currie et al. (1981) suggested that the Brookville rocks have undergone two major deformation events that the Greenhead rocks have not experienced. However, the contact between the the two units is tectonic, and an unconformity has not been observed (O'Brien 1977; Giles and Ruitenberg 1977).

GREENHEAD GROUP. - The Greenhead Group consists mainly of sediments and metasediments with some interspersed intrusive rocks (Alcock 1938; Rast et al. 1976). Hoffman (1974, quoted by Ruitenberg et al. 1977) suggested a general Neohelikian age, based on the presence of the stromatolite *Archaeozoan Acadiense*, for the Greenhead Group. These rocks occur mainly in and around the Saint John area, and are entirely contained within the Avalon Terrane. The intense deformation of these rocks probably occurred during the assembly of Avalonia during late Precambrian.

COLDBROOK GROUP. - Somewhat younger Proterozoic volcanic rocks of the Coldbrook Group outcrop along most of the southern coast of New Brunswick. Although the contact between the Coldbrook and the older Greenhead is everywhere tectonic (O'Brien 1977), most past workers have regarded the Greenhead Group as the basement on which the Coldbrook Group was deposited (O'Brien 1977; Hayes and Howell 1937; Rast et al. 1976). Cormier (1969, as quoted by Giles and Ruitenberg 1977) reported an age of 750 $\pm$ 80 Ma, based on a Rb-Sr whole rock isochron. The Coldbrook Group consists of volcanic and less abundant sedimentary rocks, and their metamorphic equivalents (Alcock 1938; Potter et al. 1979). It can be divided into three distinct belts from east to west (Giles and Ruitenberg 1977). The eastern belt, located along the coast of the Bay of Fundy, consists mainly of intensely deformed mafic and felsic flows and tuffs and related sediments. The central volcanic belt is made up of weakly deformed, felsic and lesser mafic flows and coarse, lithic tuffs. Giles and Ruitenberg (1977) regarded the central and eastern belts are facies equivalents. The western belt is made up of the same volcanic lithologies as the central belt, but the rocks of the western belt are intensely deformed (Giles and Ruitenberg 1977). The Coldbrook Group probably represents what remains of a long chain of volcanic islands (e.g. Rodgers 1970).

CAMBRO - ORDOVICIAN SEDIMENTARY ROCKS - Cambrian sedimentary rocks, mapped as the Saint John Group, overlie the Coldbrook

Group with angular unconformity (Hayes and Howell 1937; Alcock 1938). They outcrop in what are either isolated fault bounded basins or synclines, all within 50 kilometers of Saint John (Alcock 1938; Rodgers 1970). The Cambrian rocks are dominantly platformal sediments with a few isolated beds of lapilli tuff (Ruitenberg et al. 1977; O'Brien 1977); The lower part of the unit consists of approximately 600 meters of purplish, white, and gray sandstones and conglomerates (Hayes and Howell 1937; Alcock 1938; Rodgers 1970; Ruitenberg et al. 1977). These sediments are overlain by 200-300 meters of fossiliferous, dark gray shales with minor interbedded limestones (Rast and Stringer 1977; Rodgers 1970). These rocks in turn grade upward into a turbidite sequence of interbedded graywackes and shales that range in age from the medial Cambrian through early Ordovician (Rast and Stringer 1974; Rodgers 1970). The stratigraphic sequence thus suggests a marine trangression during the Cambrian and early Ordovician (Hayes and Howell 1937; Ruitenberg et al. 1977). The Cambrian rocks in southern New Brunswick are generally only weakly deformed, except in Saint John, where they are intensely deformed (Ruitenberg et al. 1973).

MIDDLE ORDOVICIAN - SILURIAN STRATA. - A lack of rocks of medial Ordovician through early Silurian age in southern New Brunswick suggest nondeposition during that interval (Poole 1976), possibly as a result of local uplift (Garnett and Brown 1973). Hayes and Howell (1937) reported a limestone boulder near Saint John containing fossils of possible Ordovician or Silurian age, suggesting that there may have been deposition in the Ordovician or Silurian but that all of those sediments have been eroded away. The Middle and Upper Silurian are also poorly represented in southern New Brunswick. A small zone of gray conglomerate unconformably overlying Cambrian sediments in a sliver along the Bellisle fault was assigned to the Middle Silurian by McCutcheon (1980). Around Saint John, Ruitenberg et al. (1973) described isolated sandstones with well developed graded bedding and slump structures that they correlated (on the basis of lithologic similarity) with Silurian metasediments in the Fredericton Trough to the north.

DEVONIAN PLUTONIC ROCKS. - Devonian strata are not known in the Avalon Terrane of southern New Bruswick (see Potter et al. 1979). The only rocks of that age south of the Belleisle fault occupy small plutons. A few kilometers to the north and northwest of the fault, large granitic plutonic bodies (e.g. the St. George Batholith) are Devonian in age (Potter et al. 1979; Fyffe et al. 1981). Recent work by McCutcheon (1980) has shown that the Devonian unit D-2 on the Geologic Map of New Brunswick (Potter et al. 1979) is actually a deformed mix of Devonian plutonics and Precambrian rocks.

Regional relationships suggest that during Devonian times, the Avalon Terrane was accreted to North America. The Fredericton Trough, located between the Avalon and the Miramichi Terranes, is probably to be the site of an ocean, the closure of which led to this collision (e.g. McKerrow and Ziegler 1971; Bradley 1983). Penetrative deformation, faulting, and plutonism in Avalonian rocks can all be attributed to the collision (McCutcheon 1980).

CARBONIFEROUS STRATA. - Two belts of Carboniferous rocks are present in southern New Brunswick. The first, called the Moncton Basin, extends to the northeast for about 150 km from Kennebecasis Bay to just north of Moncton (see Potter et al. 1979). The margins of the basin are straight in map view, and the bounding faults are thought to be dextral strike slip faults (Webb 1969; D. Bradley 1984). The rocks in the southwestern end of the basin near St. John (referred to below as the Kennebecasis subbasin) are red, purple, and gray nonmarine clastics (Hayes and Howell 1937; Alcock 1938). Along the basin margin is a thick unit of conglomerate known as the Kennebecasis Formation. Fossil plants identified by Bell (1927) indicate a Hortonian (Tournaisian: lower Carboniferous) age. The other major unit in the Kennebecasis subbasin is the Albert Shale (Hayes and Howell 1937; Alcock 1938). These black and gray, locally petroliferous shales are considered to be lacustrine facies equivalents of the more proximal Kennebecasis conglomerates (Greiner 1962).

Although younger Carboniferous strata are not preserved in the Kennebecasis subbasin, sedimentation continued along strike (in the main part of the Moncton Basin) well into

Pennsylvanian times. Fossil evidence summarized by Howie and Barss (1975) indicates that the first phase of post-Acadian deposition in the Moncton basin probably began around latest Devonian and continued into early Visean. These nonmarine sediments were then capped by marine sediments assigned to the later Visean Windsor Group (Bell 1927). Marine strata were in turn followed by nonmarine conglomerates that are lithologically very similar to the Kennebecasis conglomerates lower in the succession (Alcock 1938). The rocks of the Kennebecasis subbasin are generally only mildly deformed, and contrast markedly with the strongly deformed, correlative strata of the Mispec Basin that were the focus of this study. They are folded into broad open folds, with only isolated areas of more intense deformation associated with faults (Alcock 1938).

The other belt of Carboniferous rocks in southern New Brunswick extends along the coast of the Bay of Fundy approximately 40 km to the east and west of Saint John (see Potter et al. 1979). These rocks comprise what is referred to here as the Mispec Basin (see fig. 1.1). The sediments within this basin are mainly purple coarse clastics, slates, and phyllites. The succession also includes both silicic and mafic volcanics and igneous intrusive rocks (Alcock 1938). The extrusive rocks are metamorphosed to prehnite - pumpellyite facies (Strong et al. 1979).
The margin of the basin is very irregular and the relationship between the faults and the margins of the basin

is not as clear as it is in the Moncton Basin. The subsidence of most other Carboniferous basins in Maritime Canada has been related to strike slip faulting (D. Bradley 1982); if the Mispec Basin had a similar origin, it was probably formed as a result of displacement along the Harvey - Hopewell strike slip fault (fig. 1.1). The highly deformed basin fill is known as the Mispec Group, and is of probable early to medial Carboniferous age (Hayes and Howell 1937; Alcock 1938; Howie and Barss 1975). These rocks are overlain by another, presumably younger, less deformed group of sediments known as the Lancaster Formation (Hayes and Howell 1937; Alcock 1938). The Lancaster formation contains numerous plant fossils that establish its age as Westphalian B (mid-Pennsylvanian; Stopes 1914). The age relation between these two units is not well established because they are everywhere in fault contact and the age of the Mispec Group is not well constrained.

TRIASSIC STRATA. - Triassic sedimentary rocks, assigned to the Lepreau Formation, outcrop in a few small areas in southern New Brunswick (see Potter et al. 1979). These rocks are dominantly coarse clastic rift facies associated with normal faulting and extension that occurred throughout the Northern Appalachians during early Mesozoic (Stringer 1978). The Triassic rocks, which are locally several kilometers thick, lie unconformably on both Carboniferous and older rocks (Hayes and Howell 1937), thus providing an upper limit on the age of deformation of the Carboniferous beds. They are locally deformed, and in some areas a weak cleavage is associated with open folds (Stringer 1978; Stringer and Lajtai 1979). These are the youngest rocks mapped in southern New Brunswick (see Potter et al. 1979).

LATE PALEOZOIC STRIKE SLIP FAULTS. - Northeast- and east-trending strike slip faults dominate the late Paleozoic structure of the Canadian Appalachians (Webb 1969). Three of these, shown in figure 1.1, are relevant to the present study.

The Belleisle Fault is a major, northeast-trending strike slip fault that cuts through southern New Brunswick, forming the northwestern boundary of the Moncton Basin. Wilson (1962) considered it to be a strand of the Cabot fault system. Although it is long and straight, the correlation of rock units on either side of the fault suggests that the strike slip displacement along it is limited (Rast and Currie 1976; Ruitenberg and McCutcheon 1982), and is probably an order of magnitude less than the 65 km dextral displacement suggested by Webb in 1969 (Bradley, 1984).

The Harvey - Hopewell fault, which forms the southeastern boundary of the Caledonia Massif east of St. John, is important because it lies nearly along strike from the Mispec Basin. It trends northeast-southwest, and may have been the strike slip fault that originally bounded the Mispec Basin during its subsidence in early and medial

Carboniferous, before the basin fill was thrust to the northwest. The Harvey - Hopewell fault has a more complex history than other faults of similar age and orientation in the region. The earliest movement, during late Visean and Namurian, resulted in unknown (possibly 10's of kilometers) dextral displacement (Webb 1969; D. Bradley, 1984). Webb (1963) determined 16 km of later sinistral displacement during Westphalian B, i.e. during deposition of the Lancaster Formation in the Mispec Basin.

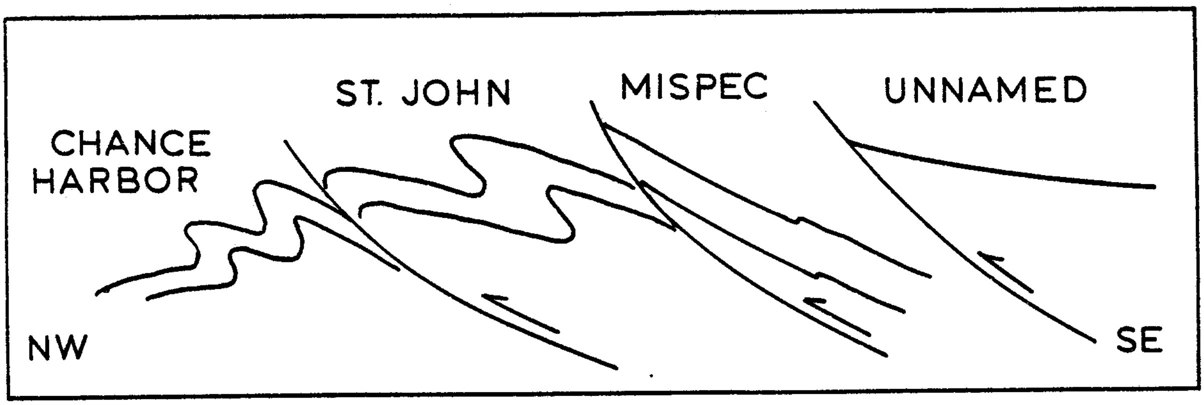
The Cobequid - Chedabucto fault system is a major east-trending strike slip fault across which the Avalon and Meguma Terranes are juxtaposed. It was first active during Devonian times (Keppie, 1982) and then again during medial Carboniferous time. This fault crosses mainland Nova Scotia, and disappears beneath the Bay of Fundy, trending toward southern New Brunswick near St. John (fig. 1.1). Post-Acadian strike slip occurred from Namurian to Westphalian C; estimates of dextral displacement during this interval range from 60 km (Bradley 1984) to 225 km (Webb 1969). While the displacement is debatable, the dextral sense of shear is well established by fabric studies within the fault zone (Eisbacher 1969, 1970; White 1983). The Cobequid - Chedabucto fault system does not emerge onshore in southern New Brunswick, but the area where it would be expected is a zone of Carboniferous thrusts (fig. 1.1). This suggests, as Arthaud and Matte (1977), Bradley (1982), and Rast (1982) have pointed out, that late Paleozoic deformation along the southern coast of New Brunswick probably occurred at a compressional bend in this fault system.

LATE PALEOZOIC PENETRATIVE DEFORMATION. - Both the Carboniferous strata of the Mispec Basin and the Precambrian rocks surrounding the basin are deformed during extensive overthrusting to the northwest (Rast and Grant 1973b). All of the intense late Paleozoic deformation in southernmost New Brunswick appears to have been associated with this thrusting event. The major structures in the area are a series of thrust nappes involving Precambrian, Mississippian, (e.g. Mispec Group), and in some areas Pennsylvanian (e.g. Lancaster Formation) rocks. West of St. John, the inland limit of this deformation can be described generally as the Belleisle Fault; however, mapping done by Rast has more clearly defined the late Paleozoic deformation front in the region (Rast and Grant 1973a; Rast et al. 1977; Rast and Currie 1976). The deformation front is marked by a zone of cataclasis that in many places follows the trend of an older ductile shear zone (Pocologan Mylonite Zone) cutting Precambrian rocks (Rast and Currie 1976; Rast, 1982). The actual location of the late Paleozoic deformation front has been established by mapping small slivers of Greenhead and Coldbrook Group in contact with Carboniferous strata in the fault zone (Rast and Currie 1976).

The deformation event that caused the extensive thrusting in the region resulted in several stacked nappes consisting of Precambrian and Carboniferous rocks (Rast and Currie 1976)(fig. 2.1). Frequent juxtaposition of rock types suggests that there are many subsidiary slices within each major nappe (Rast and Grant 1973b). To the west and the southwest of Saint John, the Chance Harbor and Saint John Nappes have been identified (Rast and Currie 1976). A higher nappe, the Mispec, has been delineated to the east and southeast of Saint John, but its boundaries have not yet been precisely defined (Rast et al. 1978). It has been thrust to the northwest over the Saint John Nappe.

There is strong deformation developed along basal and internal thrust faults associated with nappe emplacement (Rast et al. 1978). Both mylonites and breccias occur within the fault zones; near the thrusts, there is commonly a strong schistosity (Rast et al. 1978). The Chance Harbor and Saint John Nappes are large recumbent folds (F_1) with faulted lower limbs (Rast 1982), which have a well developed axial planar cleavage (S_1) (Rast and Stringer 1974). The S_1 cleavage is everywhere developed in the nappe complex; the recumbent folds (F_1), while commonly present, are not everywhere developed (Ruitenberg et al. 1977). Several generations of folds and cleavages are locally developed, (Rast and Stringer 1974; Ruitenberg et al. 1977), and the entire zone is effected by some degree of regional metamorphism (Rast 1982). The grade of metamorphism ranges from

Figure 2.1. - Schematic section from the Bay of Fundy (on right) inland (on left) through the southern New Brunswick nappe sequence, adapted from Rast et al. (1978). From structurally lowest to highest, the sequence includes the Chance Harbor Nappe, the St. John Nappe, the Mispec Nappe, and an unnamed nappe containing Lancaster Formation.



(most commonly) lower greenschist facies to (in some areas) low garnet grade (Rast et al. 1978; Rast 1982). Carboniferous granites in the region are affected by what is probably the latest phase of deformation associated with this event (Rast 1982).

The age of the deformation in the zone is difficult to constrain. It is certainly pre-Triassic, since Triassic rocks on the coast of the Bay of Fundy are not intensely deformed (Rast et al. 1978). Both the pre-Carboniferous rocks and the probable Lower to Middle Carboniferous Mispec Group rocks are strongly deformed (Rast and Grant 1973a; Rast and Grant 1973b). The Pennsylvanian rocks of the Lancaster Formation are locally cleaved, and in some areas folded, but are generally not as deformed as what are regarded as the older Mispec Group rocks (Rast and Grant 1973a; Rast and Stringer 1974; Rast et al. 1978). In the Chance Harbor Nappe, Pennsylvanian (Westphalian B) Lancaster formation rocks are thrust over rocks of probable Mississippian age from southeast to northwest (Rast and Grant 1973a). Elsewhere, slivers of Pennsylvanian rocks are faulted within Mississippian rocks, also indicating that Pennsylvanian rocks are involved in the deformation (Rast and Currie 1976). Rast (1982) speculated that the deformation could have begun as early as late Devonian, but the major part of the deformation probably occurred during late Carboniferous times, after the deposition of the Lancaster Formation. According to Rast's present interpretation, rocks of the Lancaster Formation are less deformed than the rocks of the Mispec Group because the Lancaster rocks were higher in the nappe pile (Rast, pers. comm., 1981).

The presence of this anomalous zone of severe late Paleozoic deformation has been interpreted in terms of regional tectonics in several ways. Ruitenberg et al. (1973, 1977) labelled this area the "Fundy Cataclastic Zone", and suggested that the deformation was not related to thrusting at all, but rather was the result of large scale upwarping and subsequent faulting at the edges of the upwarped belt. They attributed the upwarping to "post-Acadian readjustment".

Arthaud and Matte (1977) suggested by means of a figure (their figure 5), that the deformation in southern New Brunswick was associated with a bend in a major right lateral strike slip fault system. On an Atlantic reconstruction, they extended this strike slip system from the Southern Appalachians through the Northern Appalachians and Hercynian Europe, to the Urals.

Rast and Grant (1973a and b) correlated this deformation with the Variscan (Hercynian) Orogeny in western Europe. By comparison with the pre-drift configuration of the Northern Appalachians and the British Isles, Rast and Stringer (1974) suggested that it represented a full-scale orogeny associated with plate convergence. Rast (1982) has since adopted Arthaud and Matte's (1977) hypothesis that the Variscan deformation in the Maritimes resulted from strike slip. Rast (1982) has suggested that the Chedabucto Fault in Nova Scotia, which is generally interpreted as a right lateral strike slip fault (Eisbacher 1969; Webb 1969; Bradley 1982; Keppie 1982), was probably linked with the Variscan overthrust belt in the British Isles.

GEOLOGY OF THE MISPEC NAPPE

The study area is located entirely within the Mispec Nappe. Most of the recent geological mapping within the nappe has been done at a scale of 1:15,840 by A. A. Ruitenberg of the New Brunswick Bureau of Mines and Natural Resources (Ruitenberg et al. 1973). Prior to that, Hayes and Howell (1937) and Alcock (1938) had mapped the St. John area at scales of 1:63,360 and 1:152,046. Mapping was also done in the late 1800's by Matthews and others. While the earliest works provided a useful description of the lithologies present, it was not until the work of Stopes (1914) and Hayes and Howell (1937) that fossils were found that could help to establish the age relations of the various rock units. To this day, most of the geologic mapping in the area has been done on a reconnaissance scale. Ruitenberg et al. (1973) mapped the major structural features in the area, but because of the size of the deformed zone, did not concentrate on the finer scale structural features. This lack of detailed structural mapping in a complex area has made geologic interpretation difficult.

The Mispec Nappe is made up primarily of two rock units of probable Carboniferous age that together comprise the Mispec Group. The older unit is the Balls Lake Formation, which consists mainly of clastic sedimentary rocks. The younger unit is the West Beach Formation, which is dominantly igneous rocks with a few interbedded sedimentary layers.

BALLS LAKE FORMATION. - The Balls Lake Formation consists of fine to coarse clastic sediments, probably all nonmarine. These include conglomerates, arkoses and shales, some of which are at least in part tuffaceous (Alcock 1938). The lowest beds in the unit are schistose, with layers of strongly foliated quartz-pebble conglomerate between them (Hayes and Howell 1937). Overlying these beds are interbedded sandstones and shales. In places, the latter are metamorphosed to phyllites. Most of rocks assigned to the Balls Lake Formation are bright hematite-red or purple in color (Hayes and Howell 1937; Alcock 1938). Some fossils have been recovered from the Balls Lake Formation, but generally these are deformed plant fragments (of probable Carboniferous age), on which species determination is difficult.

WEST BEACH FORMATION. - The West Beach Formation (Alcock

1938) consists mainly of igneous rocks with some interbedded metasediments (Hayes and Howell 1937; Alcock 1938). It includes the Cranberry Point, the Courtney Bay, and the Partridge Island Formations of Hayes and Howell (1937). The West Beach Formation structurally overlies the Balls Lake Formation, and is thought to conformably overlie it (Hayes and Howell 1937). About 750 m of strata area exposed at Black River (Hayes and Howell 1937), but the total thickness of the unit is unknown.

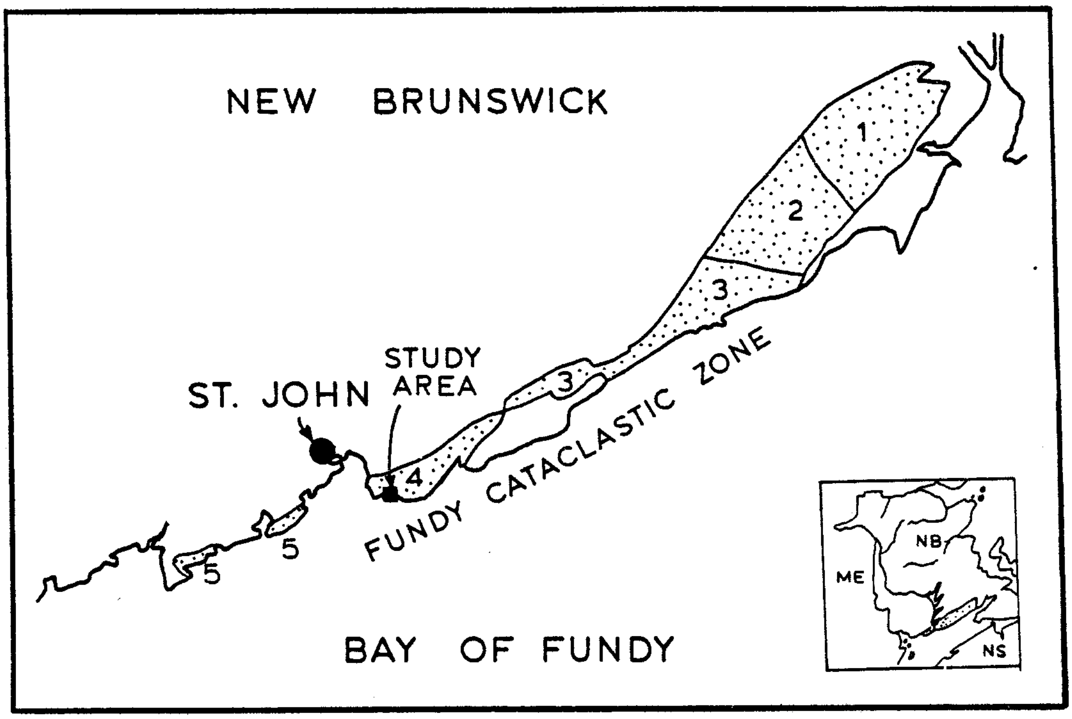
The igneous rocks consist of both intrusive and extrusive types, and they range in composition from silicic to mafic. Andesitic and basaltic flows in the unit are interbedded with sedimentary rocks (Hayes and Howell 1937). Some of the flows are amygdaloidal, while others are pillowed, indicating both subaerial and submarine volcanism (Alcock 1938). The intrusive rocks consist of dikes, sills, and some plutonic bodies (Hayes and Howell 1937; Alcock 1938). Various chemical classifications consistently show the lavas to have primitive calc-alkaline affinities (Strong et al. 1978). However, the chondrite-normalized REE plot for West Beach volcanics is parallel with that of oceanic basalts (Dostal and Strong 1983). Strong et al. (1978) reported that the mafic volcanics of the West Beach Formation have been metamorphosed under prehnite-pumpellyite facies conditions, probably between 315-370/C and 1-2 kbars.

The sedimentary rocks present in the unit resemble those that make up the Balls Lake Formation, and include conglomerates, sandstones, and shales, with minor limestone and tuff (Hayes and Howell 1937; Alcock 1938). These sediments are mainly purple in color with some gray layers (Hayes and Howell 1937; Alcock 1938). The conglomerates are mostly purple pebble conglomerates in which the pebbles are commonly flattened or rotated parallel to cleavage, and in places are strongly sheared. The finer grained rocks (shales and shaly sandstones, now phyllites and schists) are intensely cleaved, and in places the cleavage so obscures the bedding that it can only be seen at major lithologic changes (Hayes and Howell 1937). Limestone is not common in this unit, but one bed 15 meters thick (possibly a Windsor equivalent) was reported by Alcock (1938). Several fragmental volcanic and tuffaceous beds were reported by Hayes and Howell (1937).

STRUCTURAL GEOLOGY OF THE MISPEC NAPPE. - The Mispec Nappe is contained entirely within the "Fundy Cataclastic Zone" as defined by Ruitenberg et al. (1973), who divided southern New Brunswick into 5 subzones based on structural style (fig. 2.2). Although Ruitenberg et al. (1973) did not recognize the existence of the Carboniferous nappe complex, the Mispec Nappe can be loosely correlated with their Subzone 4.

The general deformation sequence as described by Ruitenberg and others (1973) is very complex, with at least three generations of structures present. The most

<u>Figure 2.2</u>. - Area of late Paleozoic penetrative deformation in southern New Brunswick, subdivided into structural zones according to style and degree of deformation, after Ruitenberg et al. (1973). The Mispec Nappe, as recognized in reconnaissance by Rast, roughly corresponds to Ruitenberg's Subzone 4.

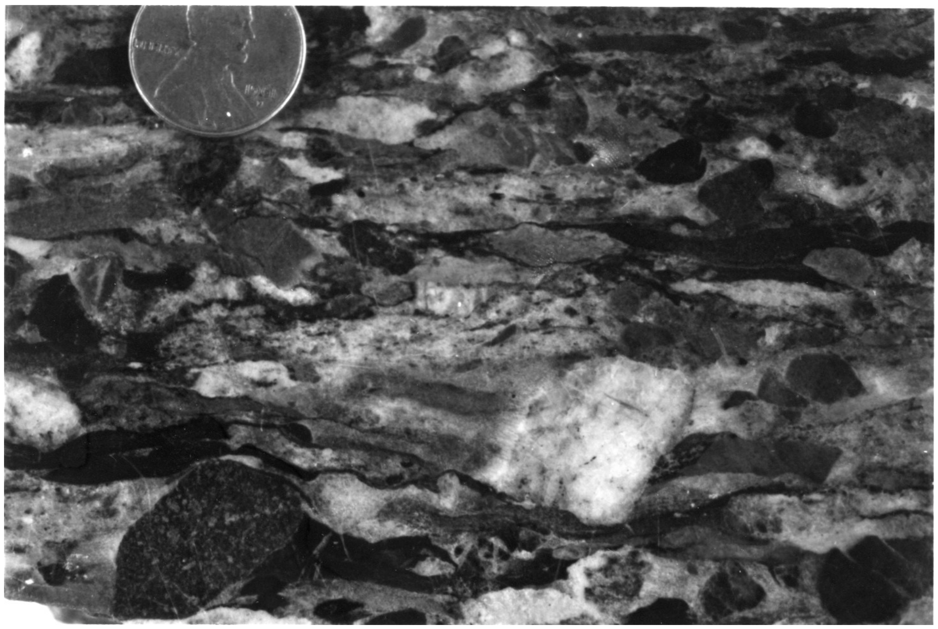


prominent structure is a penetrative slaty cleavage (S_1), which appears to be at least locally associated with a folding event (F_1) (fig. 2.3). The F_1 folds are not as extensively developed as the S_1 cleavage in the nappe. The slaty cleavage was folded during a second phase of deformation. These F_2 folds are open to close and gently plunging. They vary in size from microscopic crenulations of S_1 to folds with amplitudes of hundreds of meters. An axial surface-parallel cleavage (S_2) associated with the F_2 folds is locally recognized. The dip of this cleavage varies from gentle to moderate. The F_2 folds are commonly asymmetric, and are overturned both to the northwest and southeast. A third generation of folding (F_3), characterized by open to close chevron-type folds, is developed in some areas of the "Fundy Cataclastic Zone". These folds are not common in the Mispec Nappe.

Much of the deformation within rocks in the Mispec Nappe is internal within beds. Within fine grained beds, cleavage is extensively developed, while in coarser units, grains and pebbles are crushed and rotated (fig. 2.4). Most of the rocks within the nappe are cut by quartz, chlorite, and (locally) carbonate veins. In many areas these veins have also been deformed. Some of the veins are folded and some have cleavages developed within them. All these earlier structures have been deformed by later high angle reverse and normal faults and kink bands (Ruitenberg et al. 1973).

Figure 2.3. - F1 folds in the Mispec Nappe. This photograph was taken on the Black River Road 5 km east of the town of Mispec, and about 5 km north of Mispec Beach. F1 folds are only locally developed within the Mispec Nappe, but are thought to correlate with large recumbent folds recognized by Rast in lower nappes exposed southwest of St. John (see fig. 2.1). The lack of large recumbent folds in the Mispec Nappe suggests that it, unlike the other nappes, was transported as a single, flat sheet

Figure 2.4. - Polished surface of deformed conglomerate, from a quarry along the coastal highway, 2 km west of Mispec Beach.



CHAPTER 3

GEOLOGIC OVERVIEW OF THE MISPEC BEACH MAP-AREA

GENERAL

LOCATION AND DESCRIPTION. - The field area is a coastal exposure at the northwestern end of Mispec Beach, in the town of Mispec, New Brunswick. Mispec is about 8 km southeast of the city of Saint John on the coastal highway. Mispec Beach is a public park maintained by the town. Access to the outcrop is easy, with parking available at the park. The outcrop is reached in five minutes by walking northwest along the beach. It extends along the coast from the beach to the mouth of the Mispec river, about 200 meters. Tide fluctuates nearly 10 meters; part of the area shown in plate 1 is submerged daily.

The map area is a single large outcrop, approximately 5900 sq. meters in area. It is bounded on the south by Saint John Harbor, on the north by glacial sediment, on the west by the Mispec River, and on the east by the beach. The outcrop surface is generally flat, sloping toward the sea (fig. 3.1). The maximum elevation change from one side of the outcrop to the other is only about 3 meters. There is only minor topographic relief along the outcrop surface where blocks of rock have been pulled away along joints and faults and have since been removed.

METHOD OF MAPPING. - The outcrop was mapped at a scale of

Figure 3.1. - Photographs of the study area showing beds dipping gently toward St. John Harbor. (A) Low tide; (B) High tide. Diurnal tidal range is about 10 meters.

1:50. To ensure accurate mapping, a grid was drawn with chalk on the outcrop. The grid lines were drawn trending north-south and east-west. These directions were established by sighting with a compass on a tripod, and a ten meter distance was marked off along that line using a tape. Thus, each grid square was a 10 meter square with north-south and east-west borders. Once the pattern was established, each square was mapped onto a 20 cm square on Bristol board that was divided into centimeter squares. Physical features such as veins, faults, joints, and kinks, as well as lithology, were drawn directly on the map. The orientation of each of these structures was recorded on the map next to the structure. The field maps were reduced by a factor of three (to a scale of 1:150) to make the size of the final map more managable. The final map (plate 1) was traced directly from the reduced field maps. The original grid pattern is drawn on plate 1, and locations mentioned in the text (e.g. E,3,a) are found by referring to the numbers and upper case letters on the abscissa and ordinate of Plate 1; the lower case letter then identifies the structure within a particular grid square.

LITHOLOGY

The rocks that outcrop at Mispec Beach vary in lithology from mudstones to conglomeratic sandstones, all at low metamorphic grade. They were assigned by Hayes and Howell (1938) to the Balls Lake Formation. The rocks vary from a deep red in mudstones to a tannish red in sandstones, the color being a hematite staining.

The mudstones have been metamorphosed to phyllites. In thin section, these red rocks are dominantly quartz and muscovite, with less abundant dolomite, chlorite, hematite, pyrite, magnetite, and lithic fragments. The mudstones are strongly cleaved; small lumps on cleavage faces (approximately 1 mm in diameter) are seen under magnification to be spherical clots (framboids?) of opaque minerals, including hematite and magnetite.

Other, more silty beds, also have a phyllitic sheen. They are very similar in mineralogy to the mudstones, varying only in the size of quartz grains. These siltstones consist of mainly angular quartz grains surrounded by silt-sized grains of opaque minerals (probably hematite, magnetite and possibly some pyrite), mica flakes, strings of carbonate, and a few grains where dolomite has totally replaced plagioclase (?). One grain of detrital plagioclase was observed. The quartz grains in the siltstones vary in degree of plastic deformation, with some of the grains exhibiting undulose extinction, and other grains having uniform extinction.

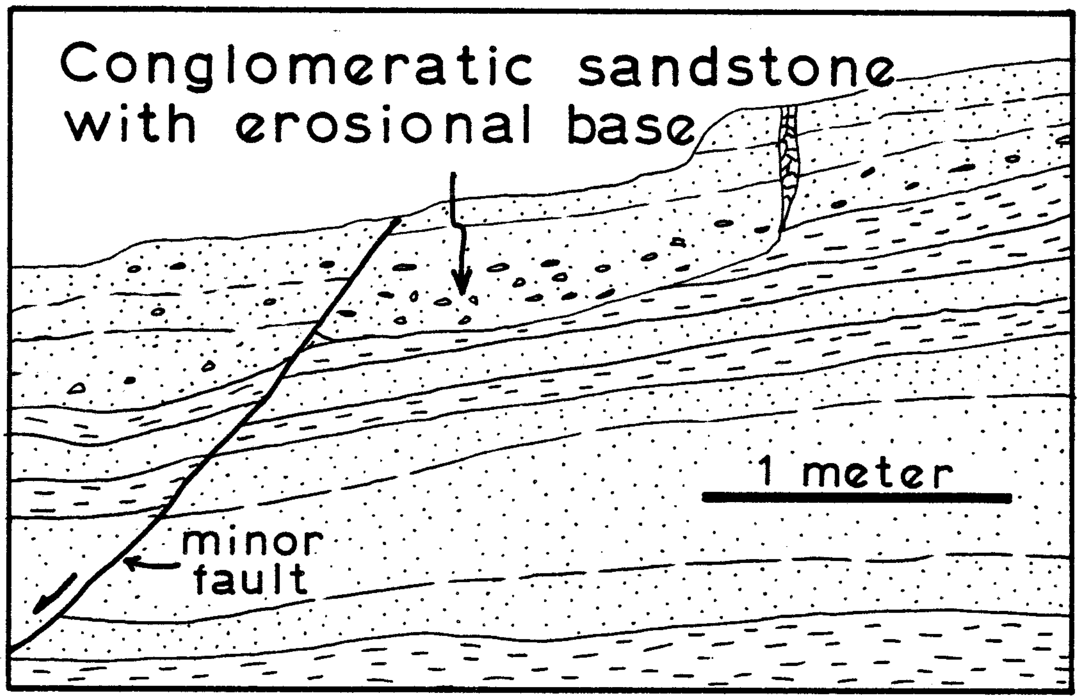
Interbedded with these phyllitic layers are beds of sandstone 0.5 to 2 meters thick. The sandstones are coarse and many include scattered pebbles. The sand-sized grains are mainly quartz and metamorphic lithic fragments; silt-sized grains form a matrix that consists of quartz, colorless mica, dolomite, chlorite, and opaque minerals. Accessories include zircon, tourmaline, and metamorphic epidote. In places, the dolomite is concentrated along cleavage planes, suggesting that it may be a later replacement mineral filling in cleavage cracks. The quartz grains in the sandstones vary in degree of deformation, some showing undulose extinction and some not.

The pebbly sandstones and conglomerates are lithologically similar to the sandstones described above, with the addition of pebbles mainly consisting of rounded vein quartz and long, cigar-shaped clasts of red mudstone. Also present, but less abundant, are lithic fragments including quartzite, phyllite, and other metamorphic lithologies. The pebbles range in size up to about 5 cm. in long dimension. Most are ellipsoidal in shape, and their long axes show a preferred orientation.

BRIEF DESCRIPTIONS OF THE MAIN STRUCTURAL FEATURES PRESENT AT MISPEC BEACH

BEDDING. - Bedding at Mispec Beach is nearly horizontal, with dips of 10 to 15 degrees to the southeast (fig. 3.1). At loc. (L,19,a), the channeled base of a conglomeratic sandstone horizon cuts into underlying siltstone, indicating that the section is upright (fig. 3.2). Local, irregular bedding surfaces between layers of sandstone and siltstone are probably bottom structures, and if so, also indicate upright beds. Structures such as cleavages and quartz

Figure 3.2. - Cross section showing interbedded sandstones (stipple), siltstones (dashes), and conglomeratic sandstones (stipple with pebbles), the channelled base of the conglomeratic unit demonstrates that beds are upright. Sketch from a photograph.



veins, without significant folding of bedding, indicate that most of the deformation in the rocks is internal within beds.

While original bedding is quite intact at the outcrop scale, in thin sections of finer-grained lithologies, one can see evidence of a disrupted compositional layering. These relict layers are almost totally destroyed, leaving no evidence of how the zones were once connected on the finer scale (fig. 3.3). The destruction of the original layering may have begun during initial compaction of the sediments, or during later deformation.

VEINS. - A brief description of veins, distribution and geometry will be given in this section, with a more extensive description to follow in Chapter 5. The distribution of veins is controlled by lithology. They are extensivly developed in the mudstones and in the sandstones, but not well developed in the siltstones; in some beds there are no quartz veins at all. Veins that cut vertically through sandstone beds abruptly end in clots of quartz, and do not penetrate adjacent siltstone beds (fig. 3.4).

Veins are most extensively developed within sandstone beds. These veins vary in length from a few centimeters to 10 or more meters, and in width from a few millimeters to about one meter. On the map (plate 1), many sets of sigmoidal, *en echelon* veins are visible (fig. 3.5). These are commonly cut by throughgoing veins that penetrate along

Figure 3.3. - Sketch from a thin section of disrupted compositional layering in siltstone. Light areas are coarser grained and quartz-rich. Darker areas are finer grained and dominated by platy minerals. Black circles are probable framboids. Slaty cleavage trends from lower left to upper right; crenulation cleavage trends vertically. 40x.

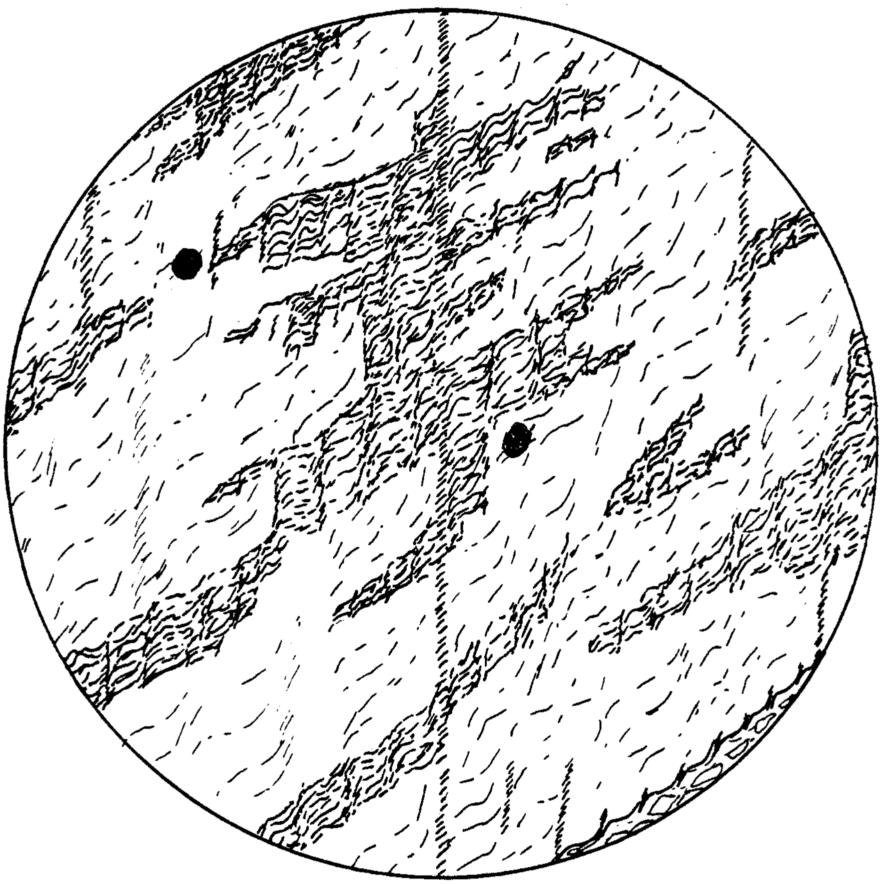


Figure 3.4. - Steeply dipping quartz veins that cut sandstone, and end abruptly rather than continuing through an underlying siltstone bed.

Figure 3.5. - A large, *en echelon* set of sigmoidal quartz veins, looking southeast in Quad. F,4.

the general trend of the _en echelon_ sets (as at location M,12,a). At the northwest end of the outcrop is a large, conjugate shear set of sigmoidal _en echelon_ veins (plate 1, location F,4,a; Shainin 1950). The veins associated with this structure dip steeply. A large proportion of the _en echelon_ vein sets in the outcrop are parallel to one set or the other of the large conjugate pair, suggesting that those veins all formed during the same deformation event. Several other patterns of quartz veins also occur repeatedly. Those patterns, the mineralogy of the veins, and a description of fibers within the veins will be presented in Chapter 5.

Quartz veins are also very well developed in the mudstones, but are quite different in size, shape, and orientation from those in the sandstones. In the mudstones they cut the tops of the beds in a series of parallel lines (plate 1, location E,3,a; fig. 3.6). These veins curl sharply just below the tops of beds (fig. 3.7). Close examination reveals that the veins are actually composites of several narrow, parallel veins separated by thin films of mudstone (fig. 3.8). A more detailed description of these veins is given in Chapter 5.

PEBBLES. - Within pebbly sandstone and conglomerate beds, pebbles are typically aligned with a trend of 050 degrees and a plunge of 10 degrees to the southwest. The alignment of these pebbles is probably the result of a combination of causes: while it may be in part a sedimentary feature, there

Figure 3.6. - Photograph of quartz veins in mudstone, taken looking down onto a bedding surface. The veins intersect the top of the mudstone bed in a series of parallel lines.

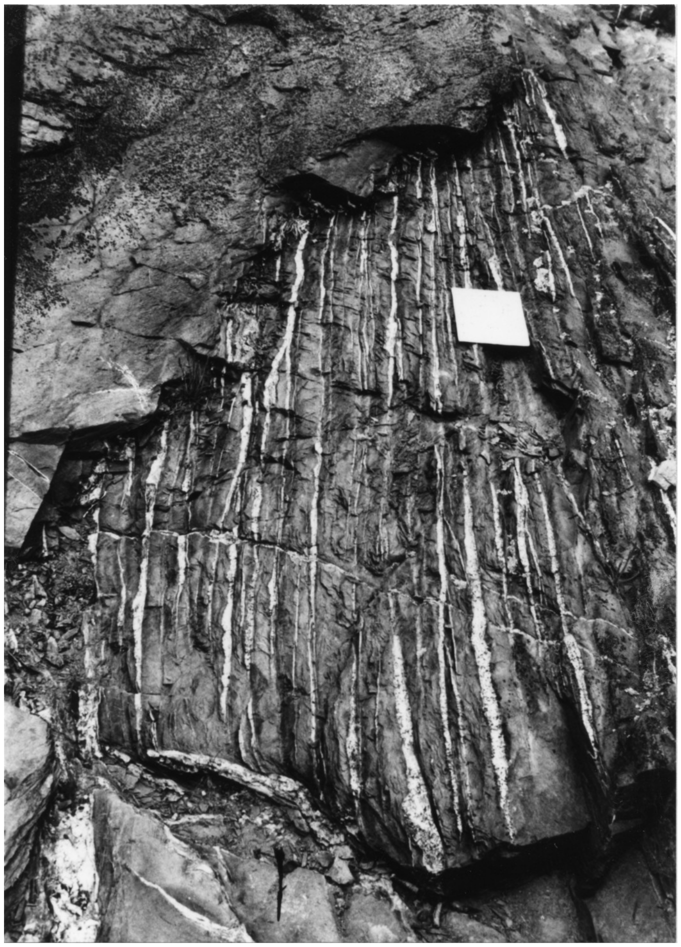


Figure 3.7. - Polished surface showing a quartz vein in mudstone dipping steeply at a bedding surface (top of photograph), and bending into near parallelism with S1 (subhorizontal) within the bed.

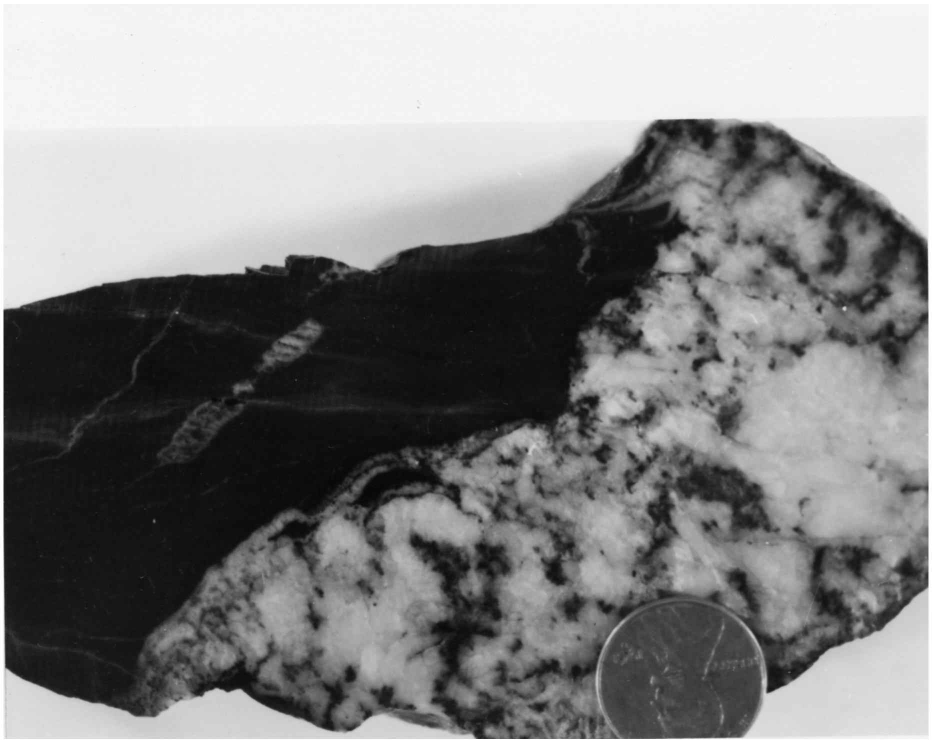
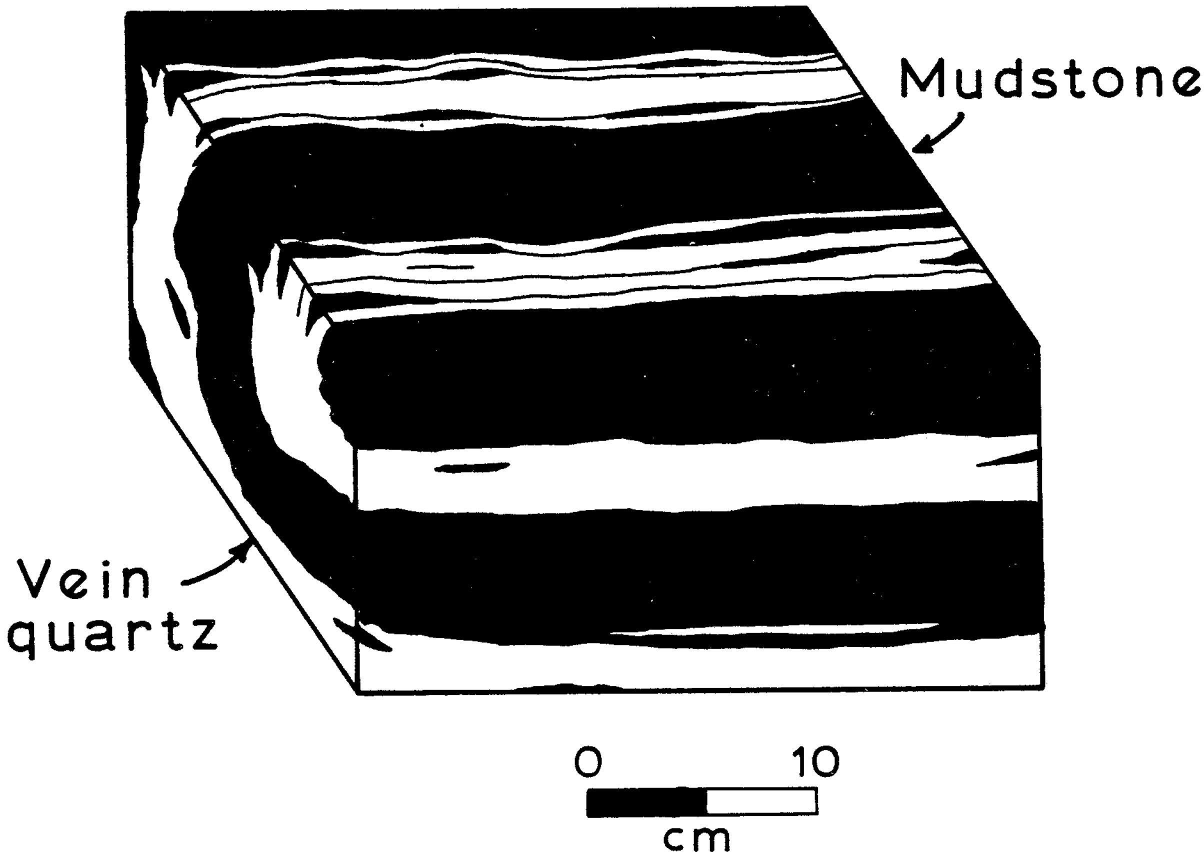


Figure 3.8. - Sketch of quartz veins in mudstone, showing that each thick vein is actually a composite of many narrower veins.



is evidence suggesting that it is also in part a tectonic feature. Quartz beards on the ends of the pebbles indicate that, at least to some extent, the alignment has resulted from the deformation and reorientation of them in the sandstone matrix (fig. 3.9). In some areas, these beards are clearly asymmetric and curved. Other pebbles are cut by quartz veins that appear to have been rotated out of alignment with the vein in the matrix around the pebble (fig. 3.10). Most of the pebbles are cigar shaped, implying constriction rather than plane strain deformation (Flinn, 1962). This is especially true for the mudstone pebbles, which seem to be stretched into an elongate shape but do not appear to be flattened along any particular plane. The metamorphic and vein quartz clasts are generally less elongate, but equally well rounded. I think that the roundness of the pebbles, as well as their orientation, may be partially a product of their sedimentary history, with some element of readjusting and reshaping resulting from constriction.

<u>CLEAVAGES</u>. - A brief, general description of the slaty and crenulation cleavages will be given here, with a more detailed discussion taken up in Chapter 4. Thin sections of the sandstones and silty sandstones show a slight slaty cleavage oblique to bedding. This is manifested by mineral elongation and concentration of micaceous minerals along parallel planes. In the finer grained siltstones and mudstones, the slaty cleavage is well developed and the

Figure 3.9. - Pebbles with asymmetric quartz beards, possibly indicating rotation of the pebbles within their sandstone matrix.

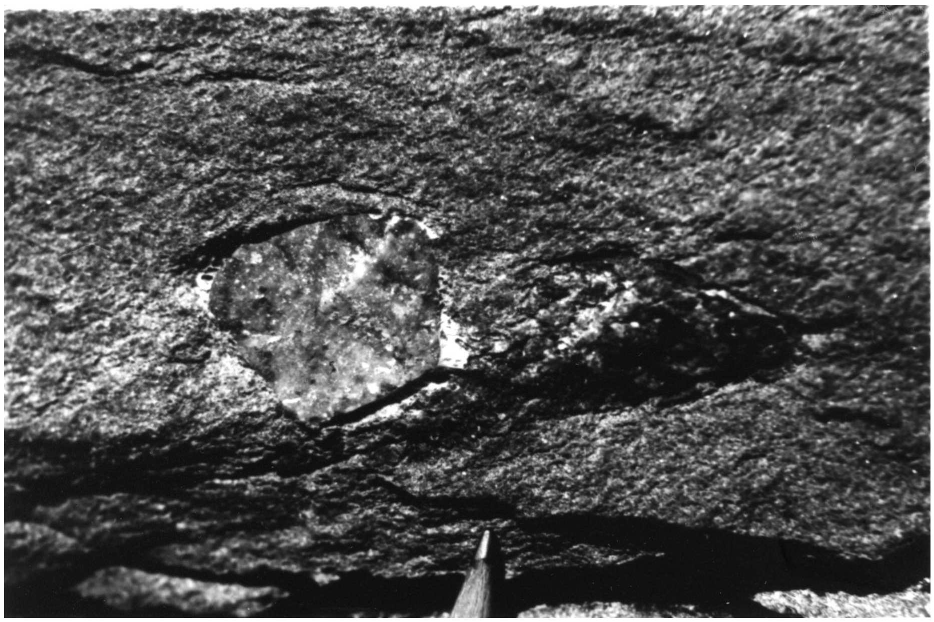
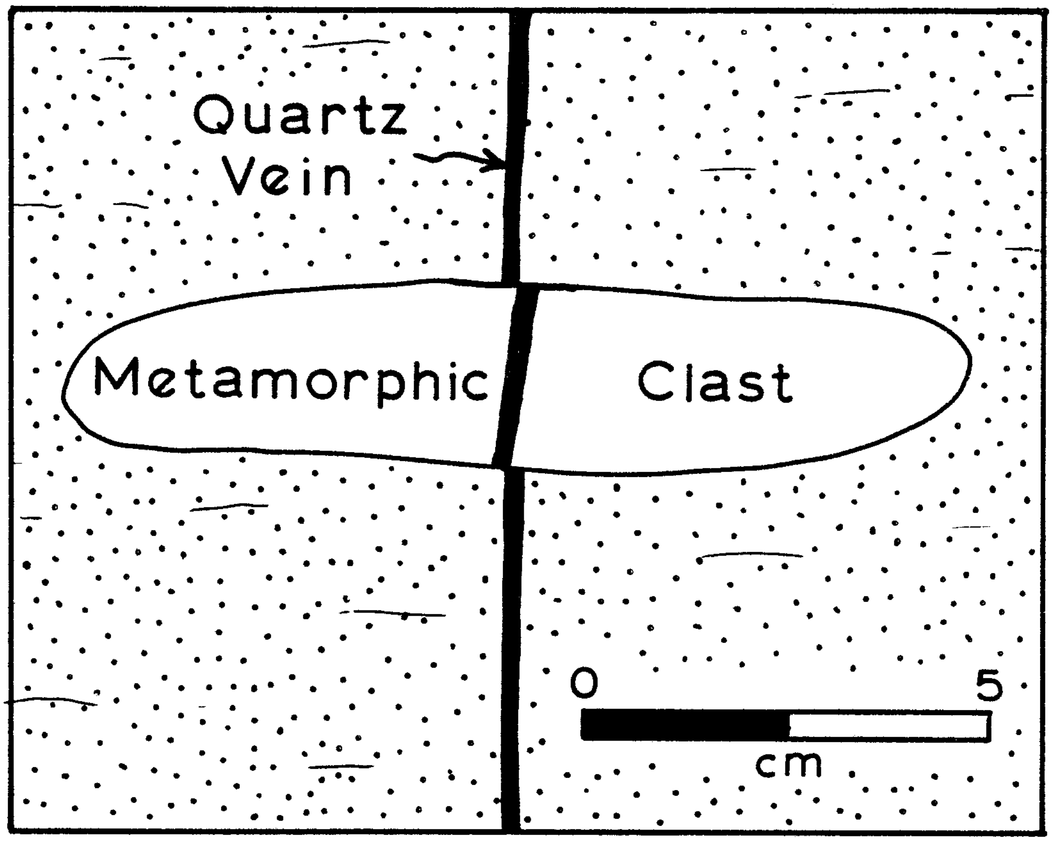


Figure 3.10. - Sketch of an elongate pebble of metamorphic rock that was cut by a quartz vein. The offset of the vein from the sandstone matrix into the pebble may have resulted from rotation of the pebble after formation of the vein.

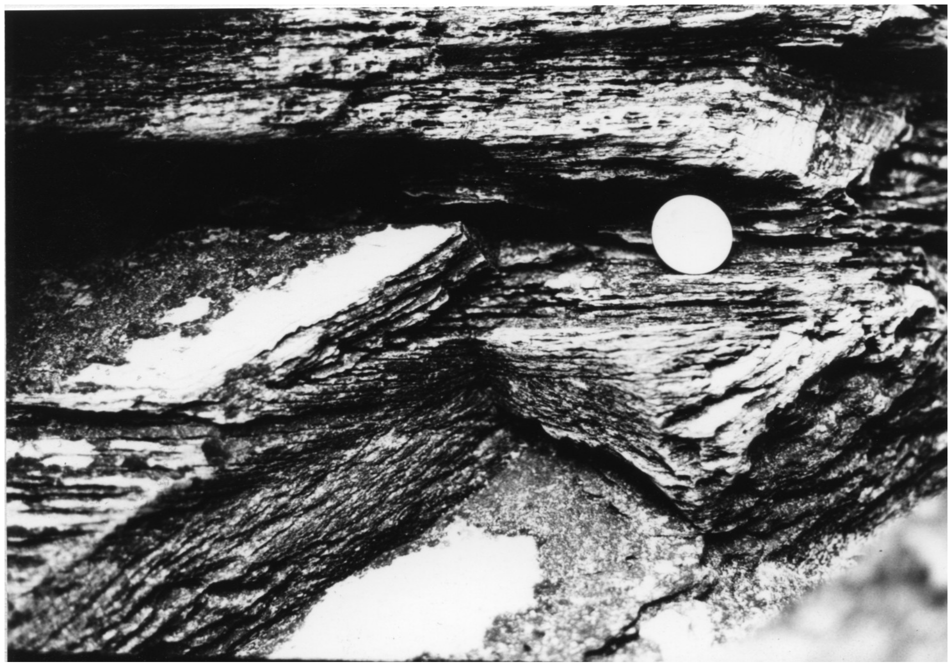


cleavage faces have a phyllitic sheen. In some areas, a mineral lineation is present on the cleavage surfaces.

In many places, the slaty cleavage appears to be folded into chevron folds (fig. 3.11) and in a few localities, what appears to be an associated axial surface cleavage is recognized; significantly, it is parallel with the enclosing bedding surfaces. However, this folding and axial surface cleavage is only an illusion resulting from the nature of cleavage development in the rocks. These fold forms or psudo folds are the result of the presence of the slaty cleavage and a later crenulation cleavage. The strenght of the development of these two cleavages is strongly controlled by any variation in grain size within individual beds. In places where the grain size is coarser, the slaty cleavage dominates and forms a parting surface. In finer grained zones the crenulation cleavage forms the parting surface. This zonal development of the two cleavages gives the appearence of tight chevron folds in the outcrop. The slaty cleavage and psudo folds are discussed in more detail in Chapter 4.

The crenulation cleavage that overprints the slaty cleavage is well developed in finer grained units, but is less prevalent and even disappears in coarser grained horizons. Small crenulation folds can be seen on some of the slaty cleavage surfaces. In thin section, the crenulation cleavage planes are dark, hematite-filled, anastomosing bands. More detailed descriptions and a dis-

<u>Figure 3.11</u>. - Slaty cleavage in siltstone. This cleavage appears in outcrop and hand sample to be folded into tight, chevron folds.



cussion of the evolution of this crenulation cleavage are presented in Chapter 4.

<u>KINK</u> <u>BANDS</u>. - Kink bands in bedding and cleavage surfaces are common in the outcrop. They vary from small bands 1 cm. wide with only a few millimeters of offset to large bands about 40 cm. in width with offsets of 15 - 20 cm. The larger kink bands can be followed across the outcrop for several tens of meters. There are two common orientations: 180 with east-side-up displacement, and 250 degrees with south-side-up displacement; kink band orientations are shown in figure 3.12 and are plotted on the map (plate 1). Some kink bands have cracked along their margins and have been filled with quartz (fig. 3.13). This may indicate that the kink band formed prior to the last phase of vein formation in the outcrop. This is only a guess since the kink bands are not common in the sandstones where the quartz veins are extensivly developed. I have seen no examples of quartz veins deformed by kink bands.

<u>NORMAL</u> <u>FAULTS</u>. - The prominent, northeast-trending faults on the map are normal faults that produce southeast-side-down offsets of a few tens of centimeters in bedding. The displacement on some of the faults is difficult to determine, but none appear to be major. Most fault surfaces are marked with mineralized, down-dip slickenside striations, which support the suggested normal displacement on the faults.

Figure 3.12. - Rose diagram showing trends of kink bands. Two sets are present, with azimuths of about 250 (south-side-up), and 175 (east-side-up).

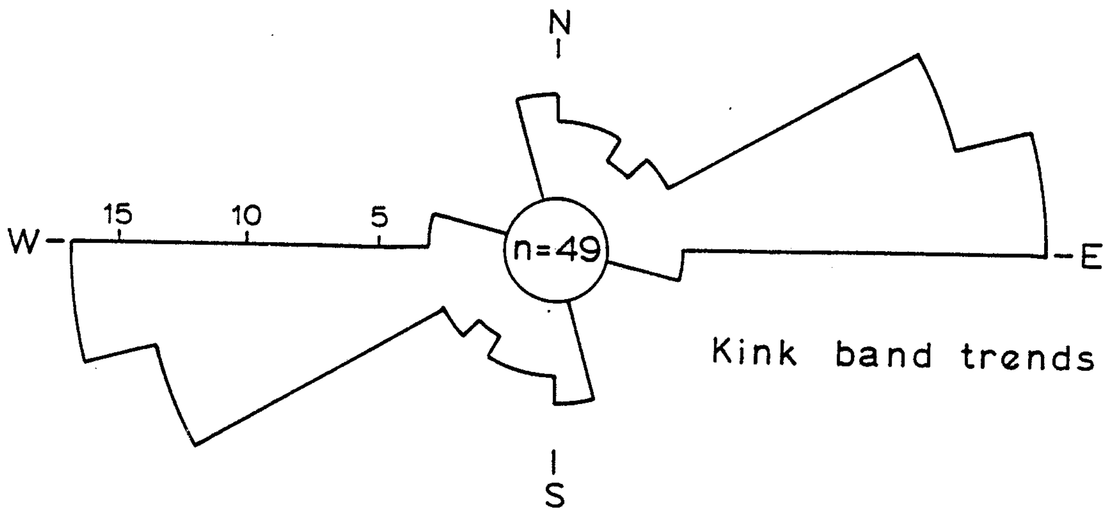


Figure 3.13. - Quartz filling cracks along the margins of a kink band.

CHAPTER 4

CLEAVAGES

As was discussed in the previous chapter, the outcrop is affected by two generations of cleavage. The first, a slaty cleavage, is developed to some extent in all beds, but especially in the siltstones and mudstones. The second, a crenulation cleavage, is not as widely developed, and is only persistent in the finer grained lithologies.

SLATY CLEAVAGE IN SILTSTONES AND MUDSTONES

MACROSTRUCTURE. - The slaty cleavage is strongly developed in siltstones and mudstones. These lithologies generally part along slaty cleavage planes. The cleavage planes are defined by a preferred orientation of micas that give the rocks a phyllitic sheen. These planes are finely crenulated, producing a lineation on the cleavage surface. The cleavage surfaces are also marked with lineations formed from the alignment of small, dark mineral clots (framboids ?). Slaty cleavage strikes about 065, dips southeast at about 20 degrees, and intersects bedding at a low angle (15 to 30 degrees).

MICROSTRUCTURE. - In thin section, the slaty cleavage is a continuous cleavage (terminology of Borradaile et al. 1982. The cleavage planes are defined by a strong mineral preferred orientation, and by mineral elongation. The degree of cleavage development is very closely controlled by the lithologies present, and can vary greatly on the scale of a single thin section. This variation in degree of development of slaty cleavage is widespread, since there is a remnant compositional layering which has been pulled apart and disrupted in the siltstones and mudstones (see fig. 3.3). The geometric relationship between the layering and the slaty cleavage suggests that the layering may have been at least partially deformed prior to the development of the cleavage. The layering is disrupted and pulled apart in a direction oblique to the slaty cleavage. Simply stretching the layers parallel with the cleavage planes would not produce this structure since the direction in which the layers are pulled apart is not parallel to the slaty cleavage planes.

SLATY CLEAVAGE IN SANDSTONES

MACROSTRUCTURE. - In the coarsest of the sandstone horizons, cleavage is not obvious. The rocks are massive and have no preferred orientation of parting when broken. The broken surfaces are rough and sandy. Locally, cleavage is developed in the vicinity of pebbles, where a concentration of dark minerals forms lines bowing around them. In sandstones with a slightly higher concentration of silt-sized matrix, the rocks do have a preferred orientation of fracture. Because it is so weakly developed, the orientation of the slaty cleavage in sandstones cannot be accurately determined in the field.

MICROSTRUCTURE. - The slaty cleavage is more obvious in thin section, and is defined by spaced, narrow bands of mica with preferred orientation. The cleavage planes do not appear to cut through quartz grains, and their closest spacing is the diameter of an average quartz grain. The quartz grains are truncated by the mica-rich bands. The truncation of the grains could be explained by partial dissolution of the individual quartz grains along the cleavage planes. This truncation could also be associated with the shearing-off of parts of quartz grains, but I have not been able to document this by matching any grains across the foliation. Some quartz grains have solution pits where adjacent grains have indented them (fig. 4.1). The quartz grains are flanked by symmetric pressure shadows of colorless mica (fig. 4.2). These pressure shadows are not developed on all sides of the grains, but rather are present in only one orientation; thin sections cut perpendicular to that orientation show no shadows. The pressure shadows and grain truncations probably resulted from shortening perpendicular to the cleavage and some preferred orientation of elongation within the cleavage planes.

CRENULATION CLEAVAGE IN SILTSTONES AND MUDSTONES

MACROSTRUCTURE. - In the field, the crenulation cleavage is the dominant cleavage in some domains. The crenulation cleavage planes are distinguished from slaty cleavage planes

Figure 4.1. - Schematic sketch of a solution pit in a quartz grain, as seen in thin sections of sandstone. Platy minerals define S_1.

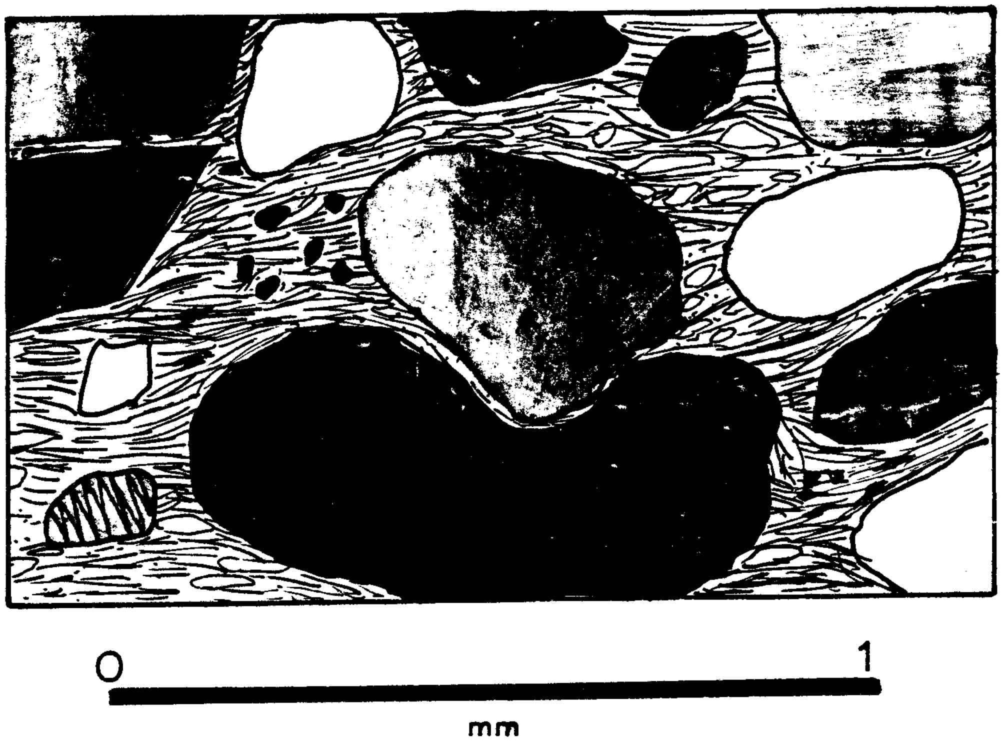
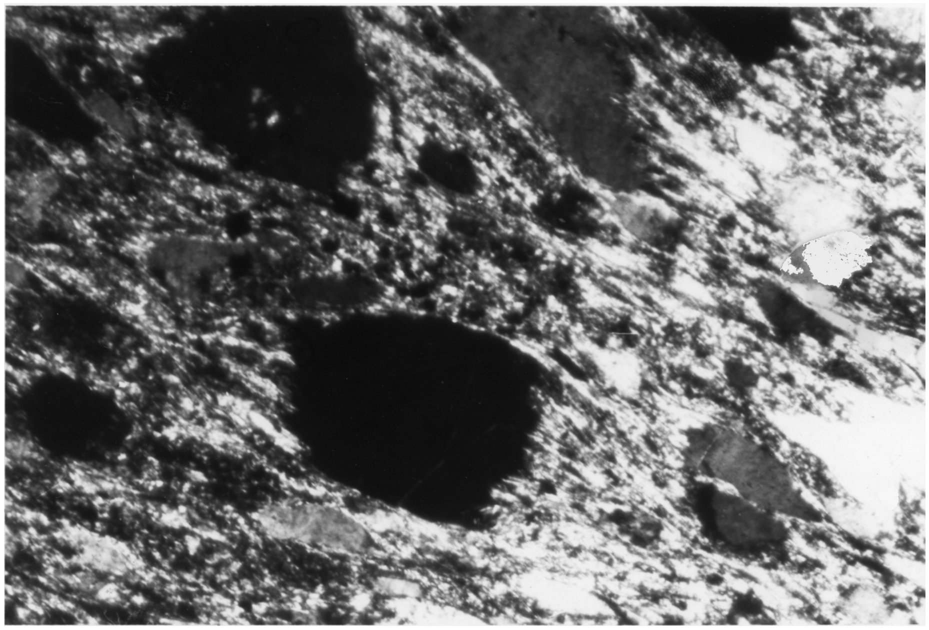


Figure 4.2. - Quartz grains flanked by symmetric pressure shadows of colorless mica. Thin section of sandstone cut normal to slaty cleavage.



by texture of the surface, and by orientation. Unlike the slaty cleavage planes, these surfaces are not deformed by crenulation folds. They have only a weakly developed mineral lineation, but like the slaty cleavage surfaces, they have a phyllitic sheen. The crenulation cleavage strikes about 045 and dips 15 to 30 degrees to the northwest (fig. 4.3).

MICROSTRUCTURE. - In thin section, crenulation cleavage is pervasive wherever the slaty cleavage is continuous. It is more sporadically developed in areas where the grain size, and thus the spacing of the slaty cleavage, is greater. The cleavage planes in the fine siltstones and mudstones are dark zones filled mainly with hematite and other opaques. In the coarser grained siltstones, the crenulation planes are commonly zones of mica concentration, but are sometimes cracked and filled with quartz and carbonate minerals. Anomalously large quartz grains are truncated or folded into parallelism along the cleavage surfaces. The cleavage planes are generally made up of a series of well developed, anastomosing surfaces (fig. 4.4). Yet in some zones, the planes are wispy and discontinuous and do not cut through certain compositional bands (fig. 4.5). The crenulation cleavage is generally associated with a regular, asymmetric folding of the slaty cleavage, but in a few places, the folding of S_1 is irregular, with inconsistent directions of asymmetry. The cleavage bows out around opaque mineral clots (fig. 4.6). A concentration of crenulation cleavage

Figure 4.3. - Lower hemisphere equal area projection of 49 poles to crenulation cleavage. Estimated average strike and dip of crenulation cleavage is 044, 24 NW.

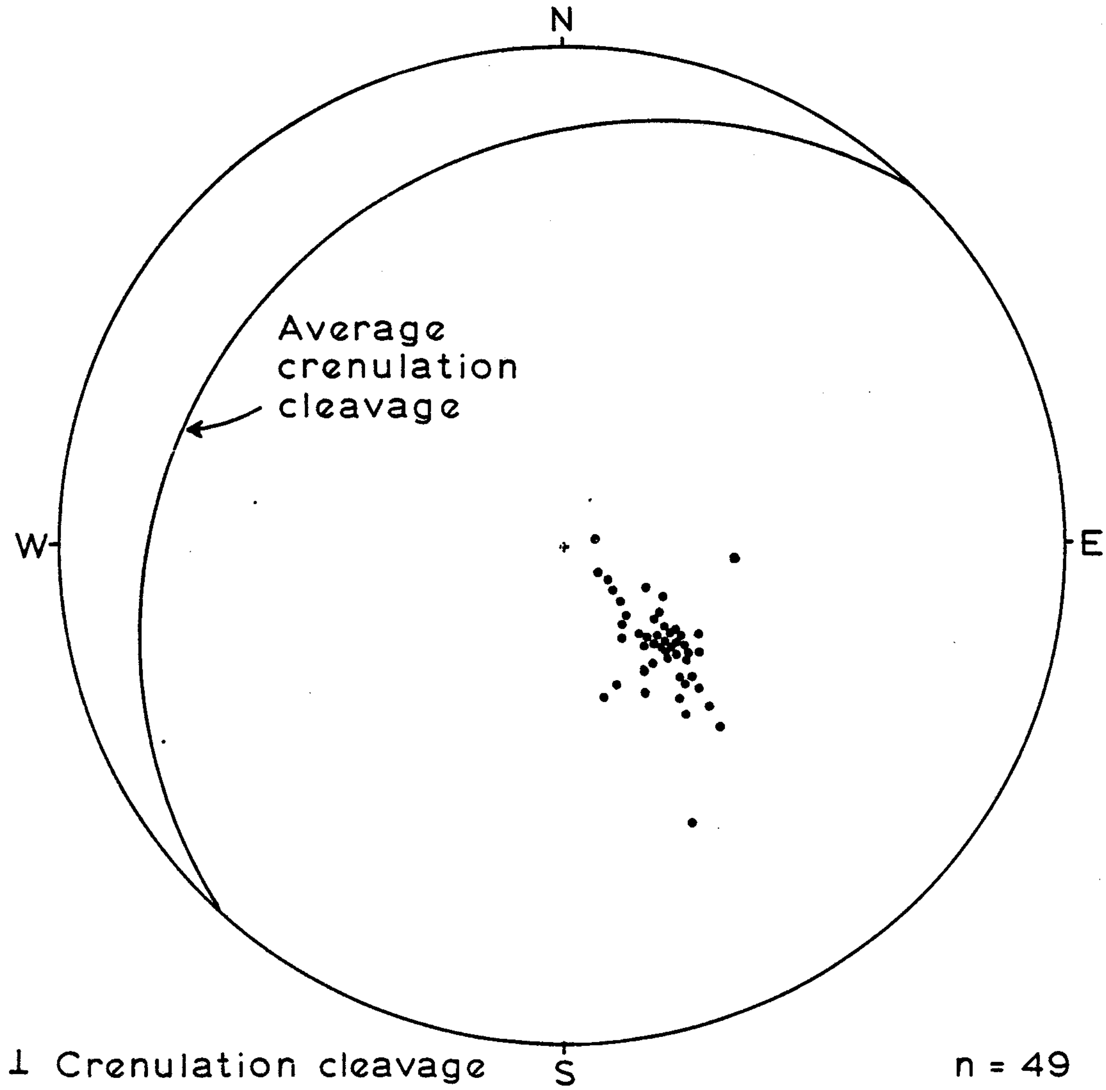


Figure 4.4. - Anastomosing surfaces forming crenulation cleavage planes in coarse siltstone.

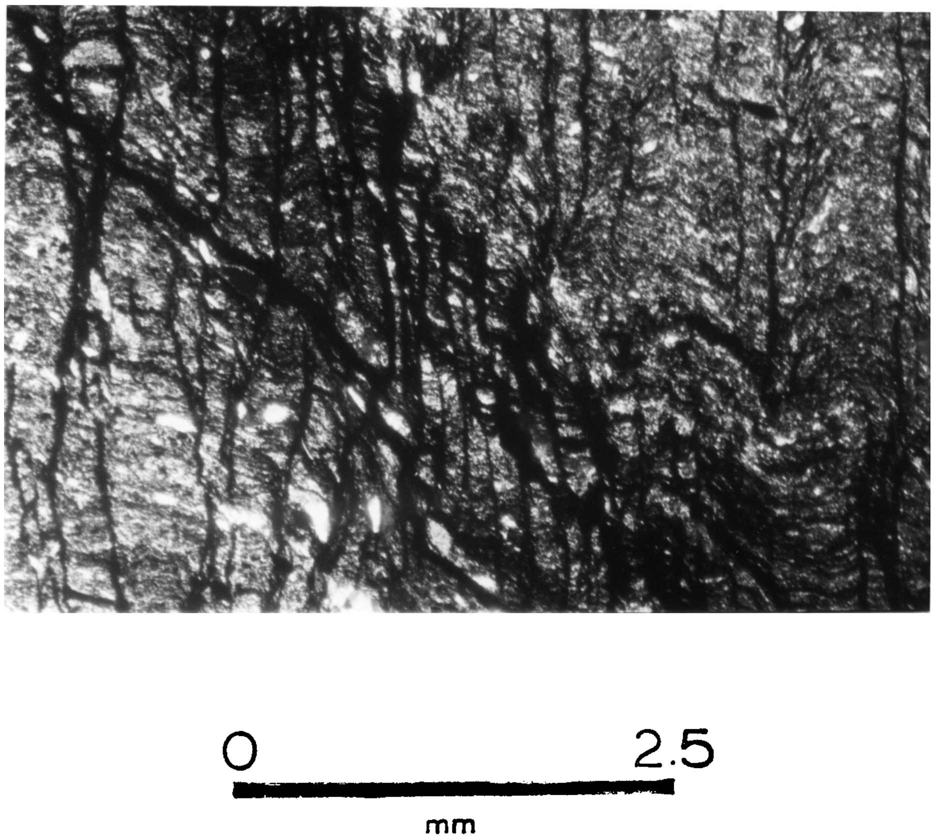
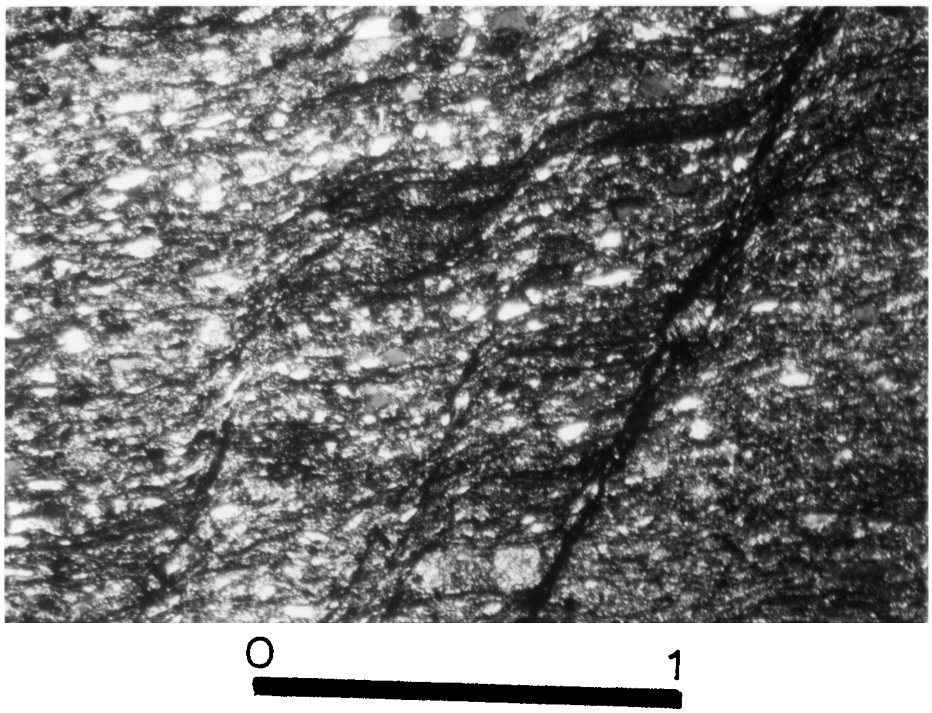
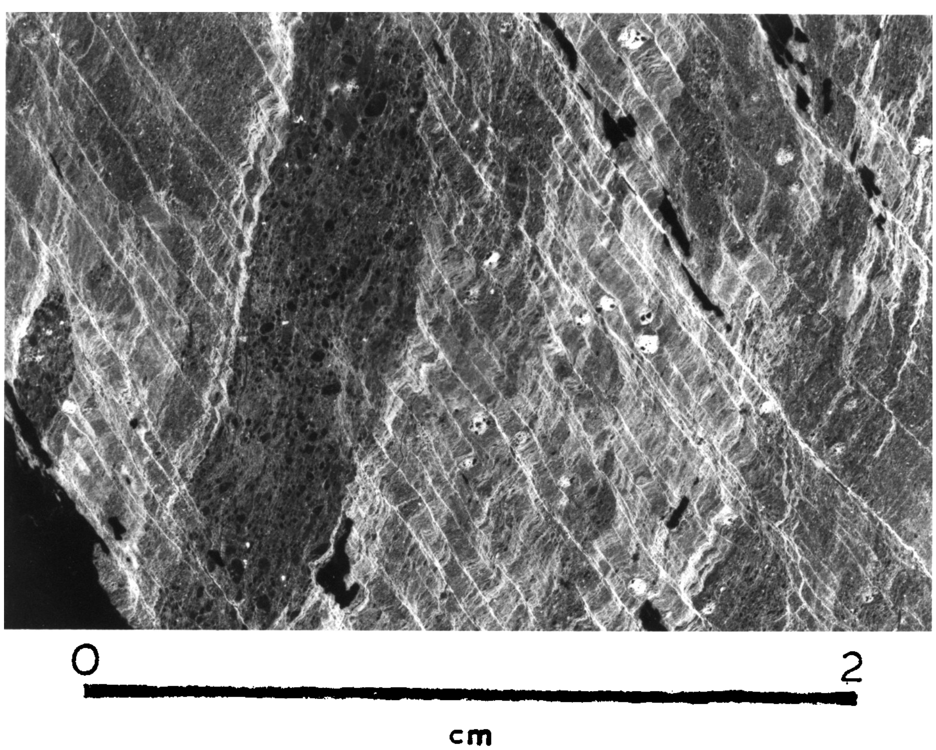


Figure 4.5. - (A) Wispy, discontinuous crenulation cleavage planes in siltstone, with consistent sense of asymmetry. S1 is subhorizontal. (B) Abrupt change in density of crenulation cleavage at a lithologic boundary. Crenulation cleavage is virtually absent in sandstone bed (light zone extending from upper left to lower right), but is the dominant foliation in siltstone beds on either side.





planes at the top and base of the clot in figure 4.6 converge to form wider cleavage bands in the more highly shortened zones above and below the clot.

CRENULATION CLEAVAGE IN SANDSTONES

MACROSTRUCTURE. - No crenulation cleavage is apparent in the sandstones in hand sample or outcrop. In the few places where a slaty cleavage is evident, it appears undeformed. There is no parting surface in the sandstones parallel with the crenulation cleavage orientation.

MICROSTRUCTURE. - There is evidence, in a few thin sections of sandstone, of crenulation folding of the slaty cleavage. Where developed, the crenulation cleavage is gradational in style (see Borradaile et al. 1982, pp. 5-6) (fig. 4.7). The cleavage planes are wispy and discontinuous, and the crenulation folds have a consistent sense of asymmetry.

DEFORMATION HISTORY DURING CLEAVAGE FORMATION

The slaty cleavage intersects relatively undeformed bedding at an acute angle. There is a mineral lineation on the cleavage surfaces, defined by small lumps (framboids ?). The average orientation of the lineation within the cleavage plane trends 045 and plunges 4 SW (fig. 4.8). On the microscopic scale, quartz grains are truncated along the cleavage planes. Some grains are pitted by other grains adjacent to them. Symmetric pressure shadows of fibrous mica

Figure 4.6. - Crenulation cleavage planes bowing out around opaque mineral clots (framboids?) in siltstone. S1 is subhorizontal.

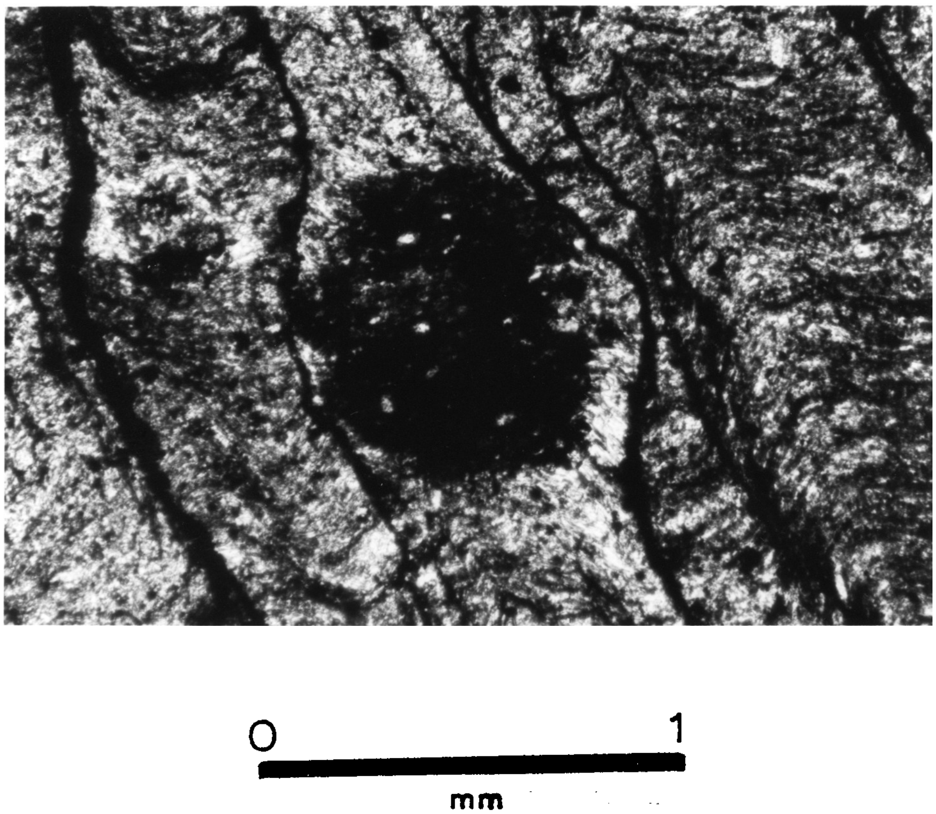


Figure 4.7. - Gradational crenulation cleavage (Borradaile et al. 1982) in coarse siltstone. This type of cleavage is also characteristic of sandstones.

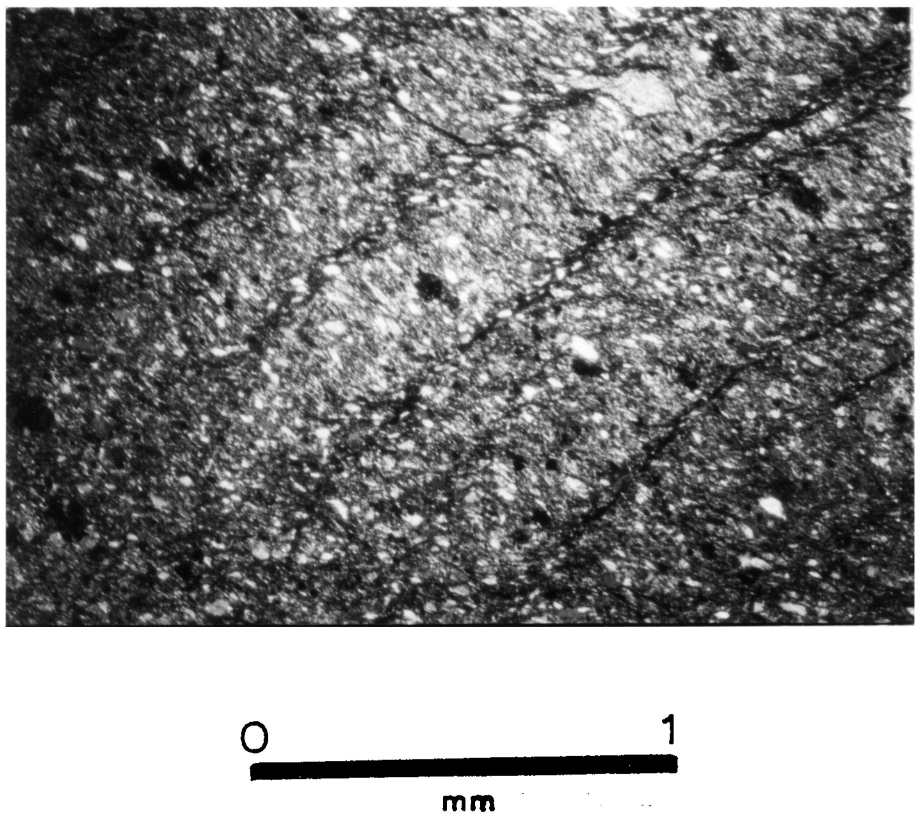
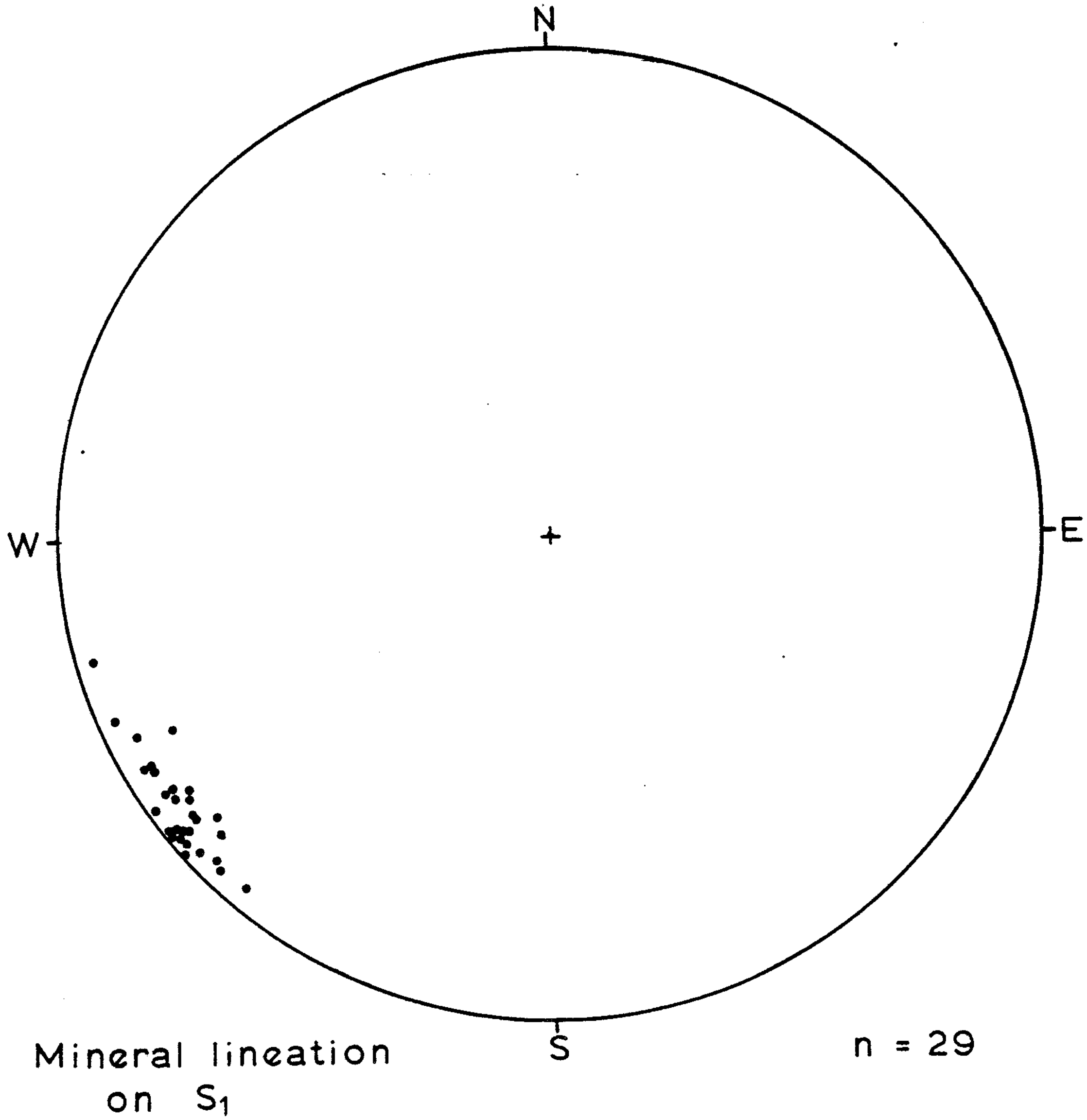


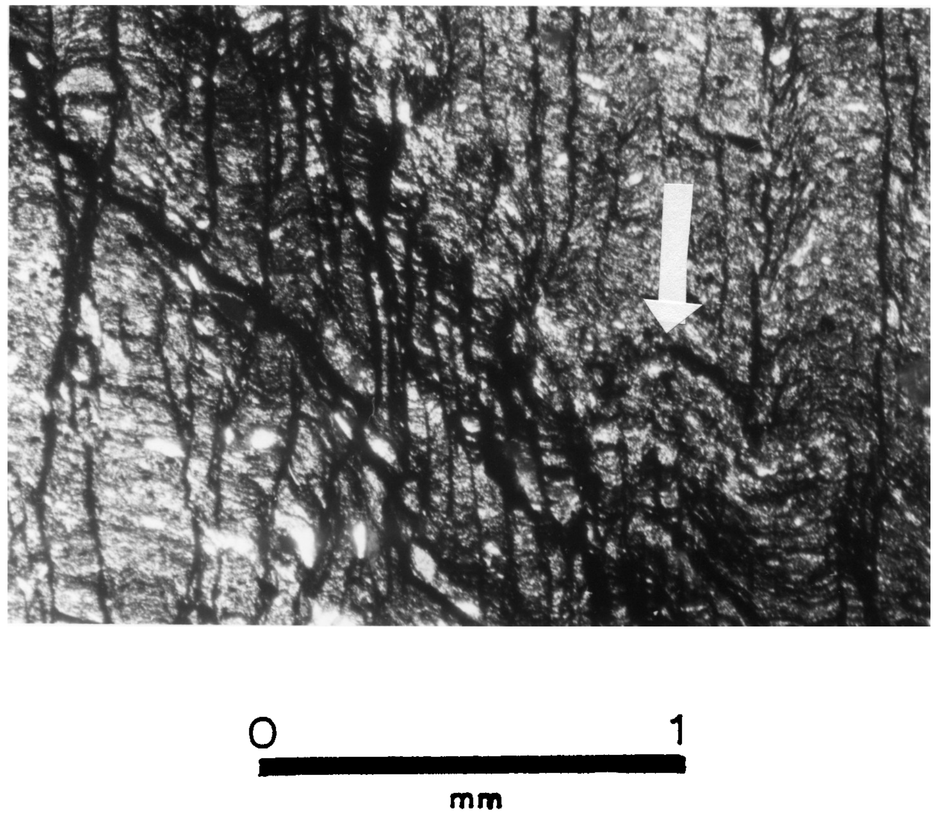
Figure 4.8. - Lower hemisphere equal area projection of mineral lineations on slaty cleavage planes. N = 29.



are developed around some quartz grains. These pressure shadows are parallel with the S_1 foliation in thin section, and have a preferred orientation. These structures indicate a shortening perpendicular to slaty cleavage, with some degree of preferred orientation of elongation within this plane.

The crenulation cleavage appears to be the result of pressure solution, as well as some slip along the cleavage planes. Evidence of pressure solution within the rock includes buckle folding of the slaty cleavage, with the truncated limbs and hinges of these between crenulation cleavage planes (fig. 4.9). These buckle folds, with their inconsistent asymmetry, are indicative of shortening perpendicular to the axial surfaces of the folds, which may have resulted in the dissolution of quartz along the cleavage planes. Further evidence for pressure solution is the concentration of dark minerals along the cleavage planes. It is possible that some proportion of the dark material along the planes is fine grained material resulting from cataclasis of material during slip along them, but observation under reflected light demonstrates that most of the dark material is composed of opaque minerals. In some areas the slaty cleavage is folded into regular, assymetric crenulation folds. The consistency of the asymmetry of these folds suggests some component of slip along the cleavage planes. However, many of the cleavage planes are discontinuous, even on the scale of one thin section, and

Figure 4.9. - Buckle folding (arrow) of S1 in siltstone.



thus could not have accomodated much shear.

In figure 4.10, a minimum shortening of 30% is required to accomodate the difference in width of the band from plane A to plane B in the rock above the mineral clot and its width across the clot. This shortening is probably largely the result of pressure solution along the crenulation cleavage planes and therefore the 30% shortening calculated may represent a minimum shortening for the rock. The 30% represents a minimum shortening since the calculation determines the difference between the width between cleavage planes straddling the mineral clot and between those same planes below the clot. This calculation does not consider any shortening prior to the point where the crenulation cleavage planes come into contact with the mineral clot.

PSEUDO FOLDS

In some areas, the rock parts along both cleavage surfaces (fig. 4.11). In most areas, however, one cleavage or the other dominates, and the rocks preferentially part along either the slaty cleavage or the crenulation cleavage surface. The surface separating domains within which one cleavage or the other dominates is parallel with bedding in outcrop. Because the two cleavages dip in opposite directions, and are so domainal in nature, the phyllites appear to be folded. Even a close examination of hand samples gives the impression that the two cleavage

Figure 4.10. - Quartz-rich mineral clot with crenulation cleavage bowing around it, suggesting shortening of 30% normal to S_2. See text for discussion.

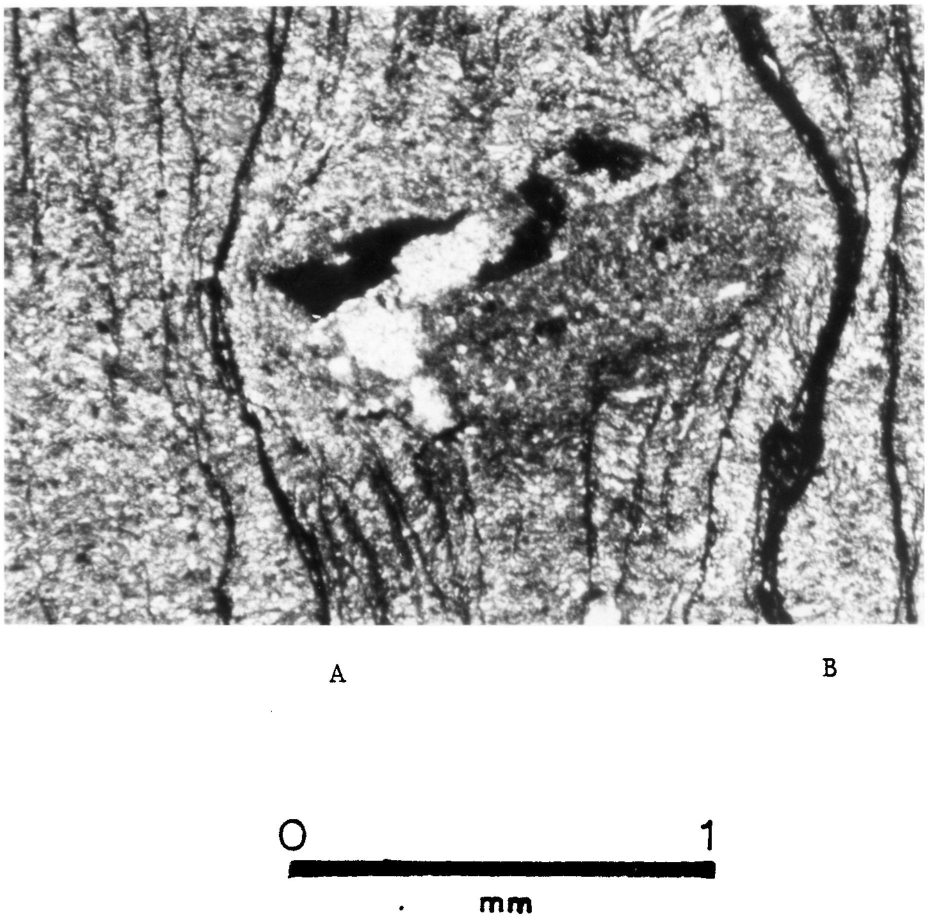
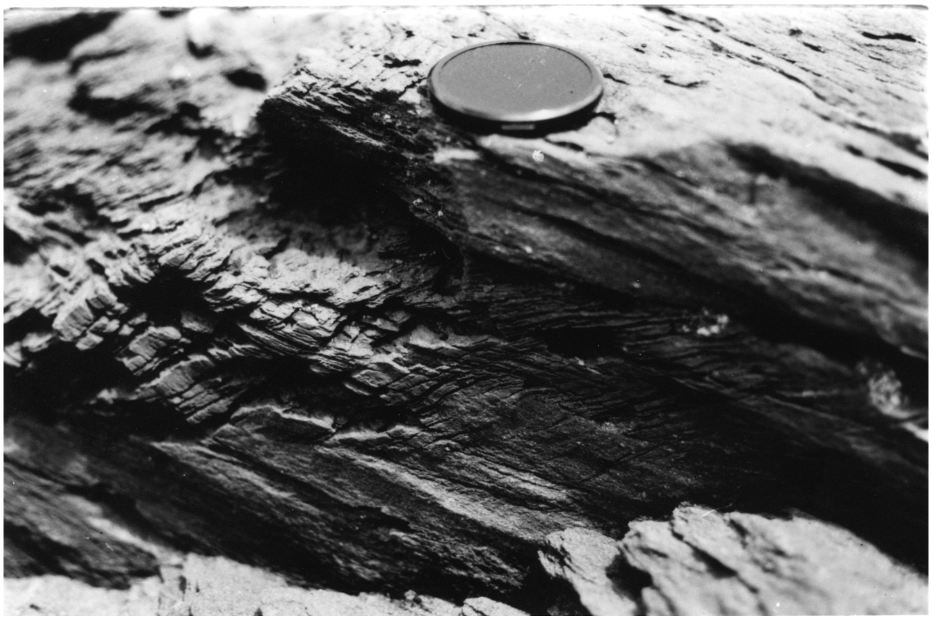


Figure 4.11. - Cross-cutting cleavages in outcrop. Slaty cleavage dips to the right (southeast) and crenulation cleavage dips to the left (northwest).



directions are actually one folded cleavage (fig. 4.12).

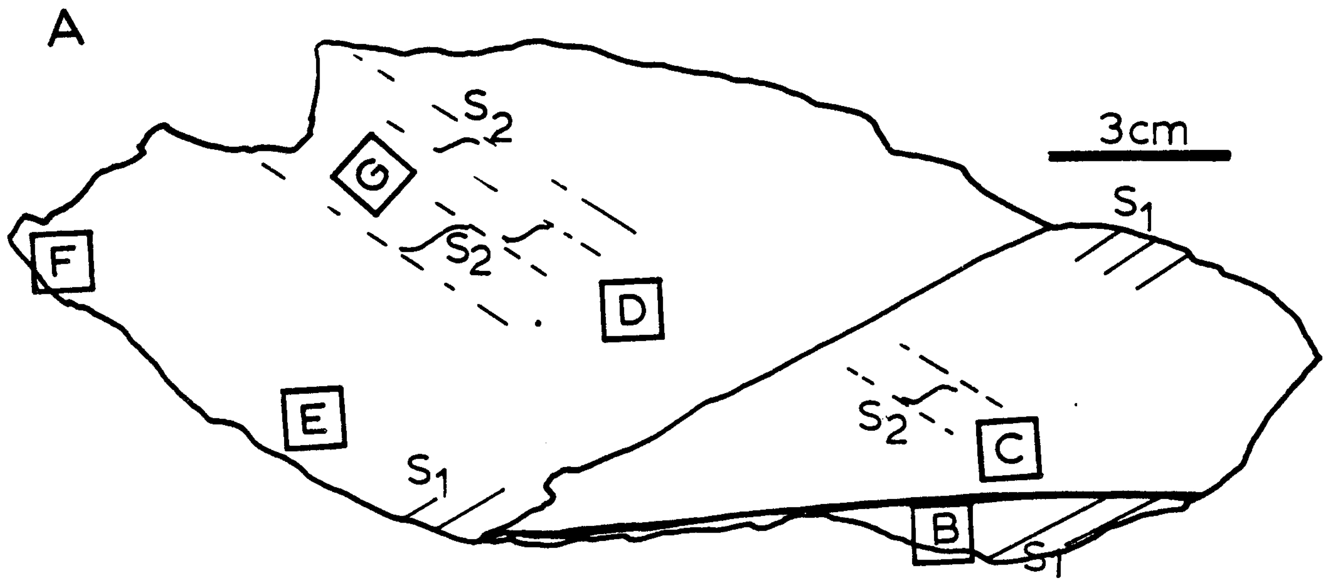
These pseudo folds are the result of subtle lithologic layering within what appear to be individual beds in the outcrop. The variation within the beds is just a minor change in the grain size that strongly controls the slaty cleavage development. In the finest grained layers, the slaty cleavage is well developed, and in turn is strongly deformed by the later crenulation cleavage. In the zones of slightly larger grain size the slaty cleavage is not as well developed and hence not very crenulated. It is in these regions that the slaty cleavage dominates, while in the finer grained regions the crenulation cleavage dominates. The lithologic layering within the individual beds is parallel to the overall bedding in the field area. The domainal nature of the two cleavages results in the presence of pseudo folds. These apppear to be tight chevron folds with their axial surface parallel with bedding. These pseudo folds might be expected in other areas where a preexisting subtle lithologic variation is affected by a slaty cleavage and a later crenulation cleavage. A simple field test to determine if a cleavage is folded or if the structure present is these pseudo folds is to inspect both limbs of the folds looking fore crenulation lineations on the cleavage surfaces. A folded cleavage will have crenulation lineations on both limbs while pseudo folds will only have them on one "limb" since the other one is actually the crenulation cleavage plane. An example of a

Figure 4.12. - Hand samples of apparent chevron-folded cleavage. Note weakly developed 'axial planar' parting.

A

B

Figure 4.13. - (A) Sketch map of the surface of a polished and etched slab that was cut from the sample in figure 4.12B, with locations and orientations of detailed photographs of the slab under reflected light (B through G). The photographs show no evidence a folded slaty cleavage, but rather show one orientation of S1 (left-dipping) deformed by a single later crenulation cleavage (S2; right-dipping). Photographs from various parts of the slab demonstrate the domainal nature of the two cleavages.



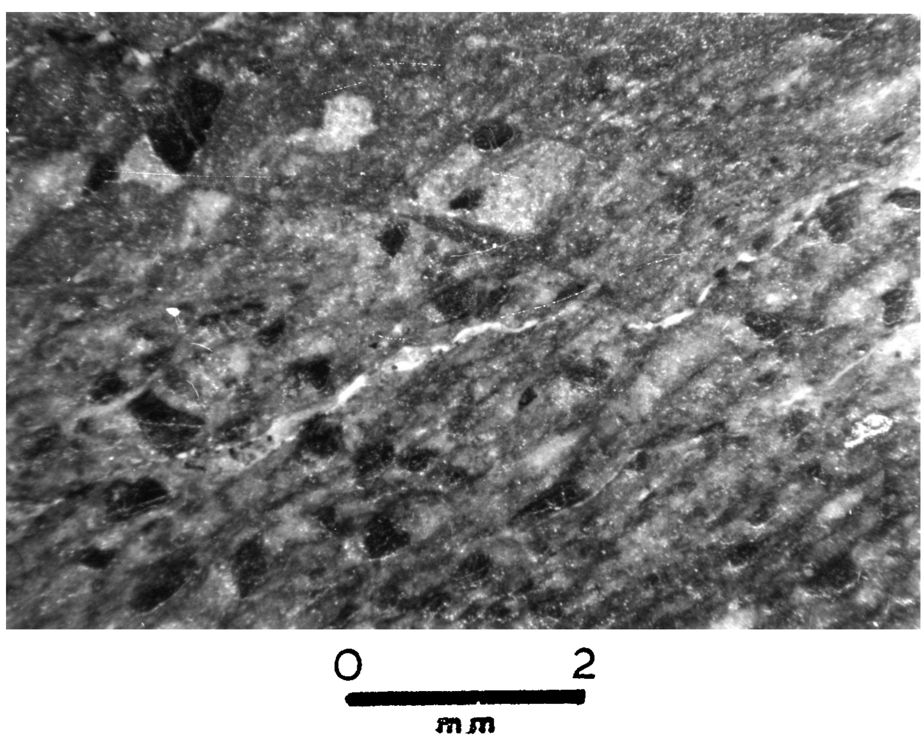
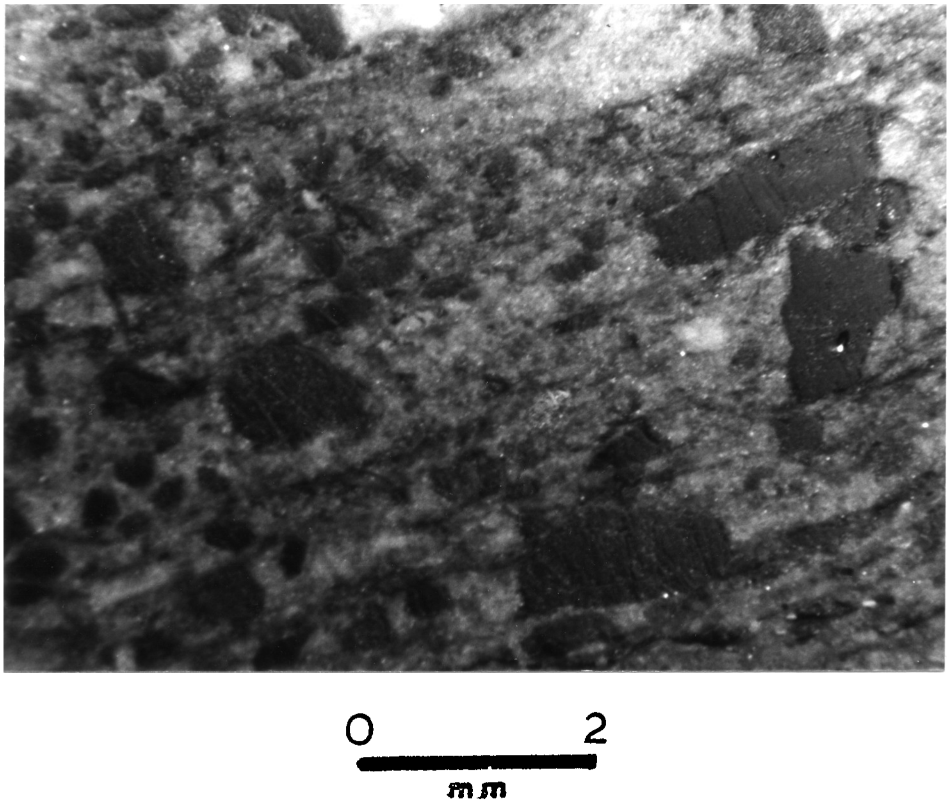


Figure 4.14 E and F



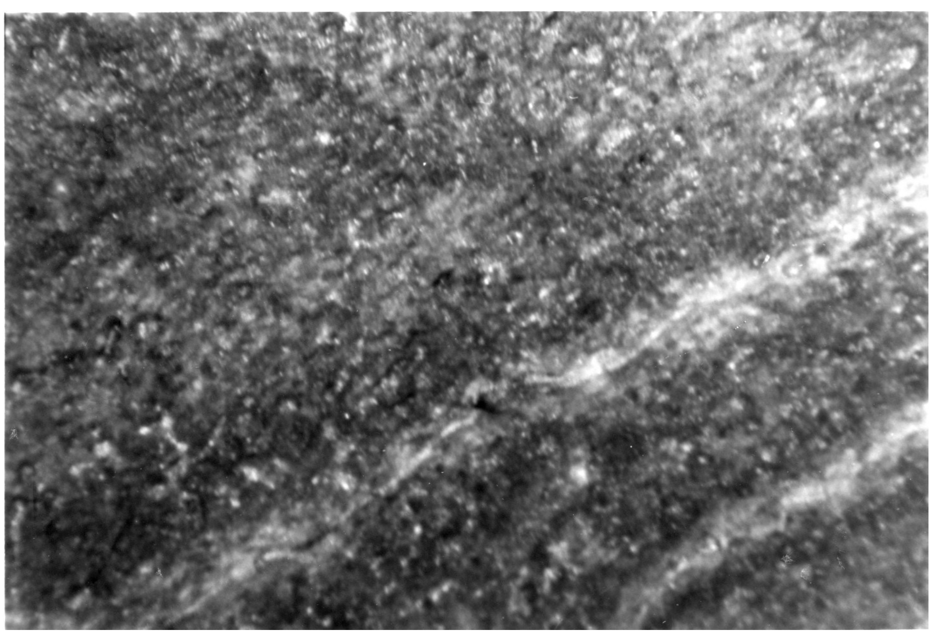


Figure 4.14 C and D

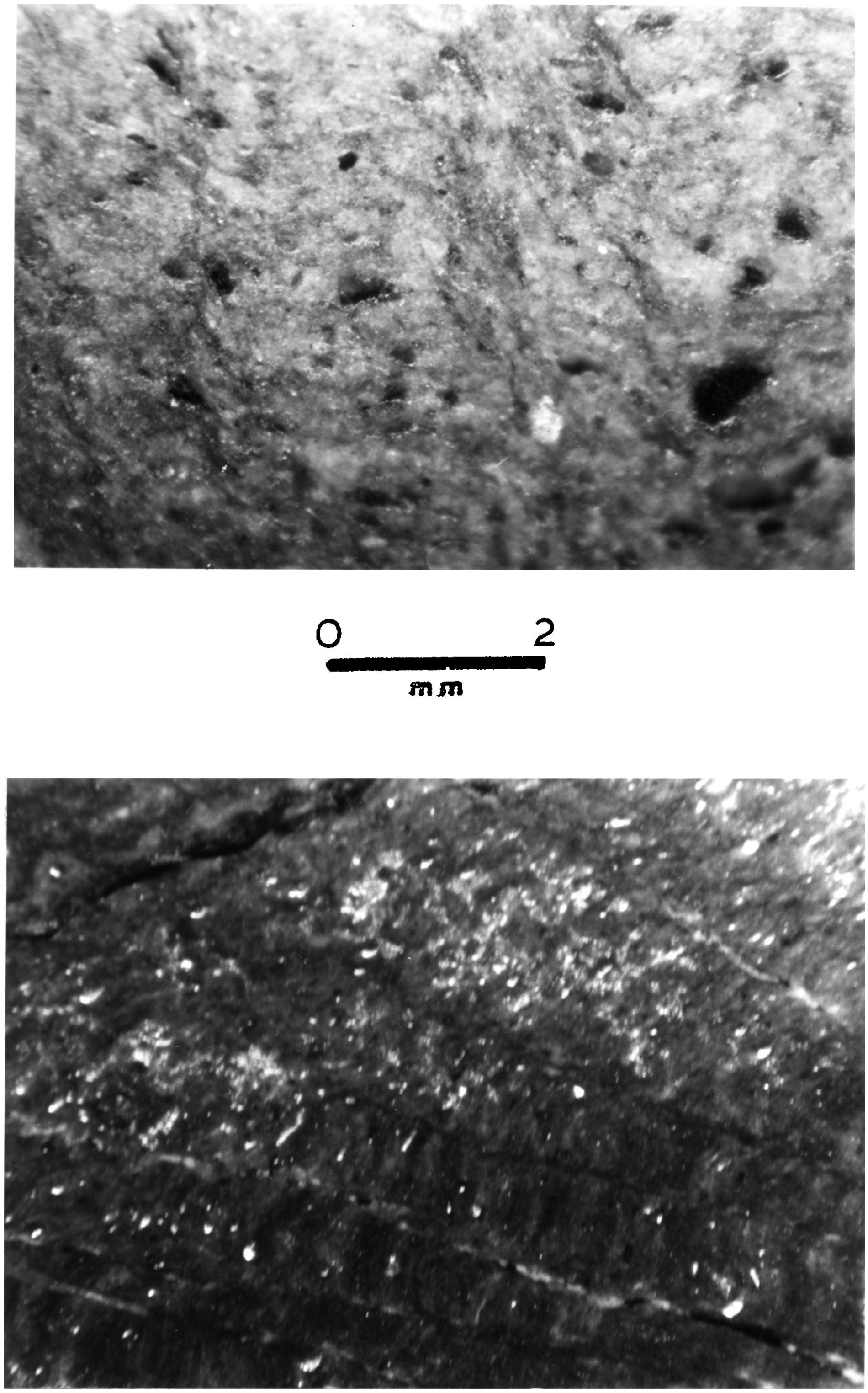
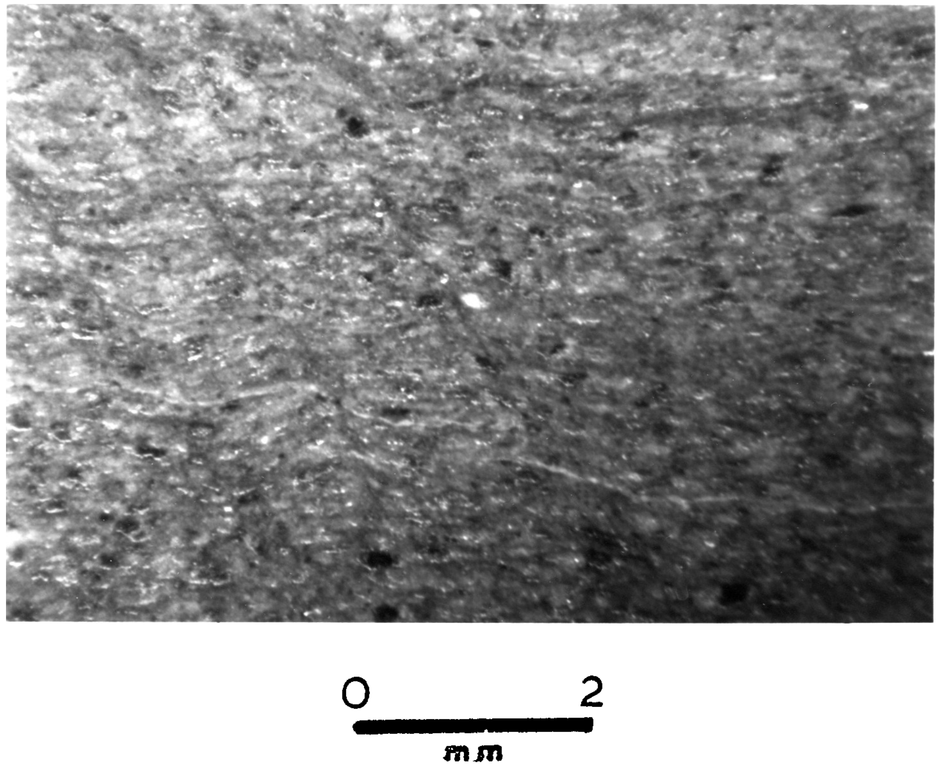


Figure 4.14 G



pseudo fold from the Mispec Beach area is shown in figures 4.14 and 4.15.

CHAPTER 5

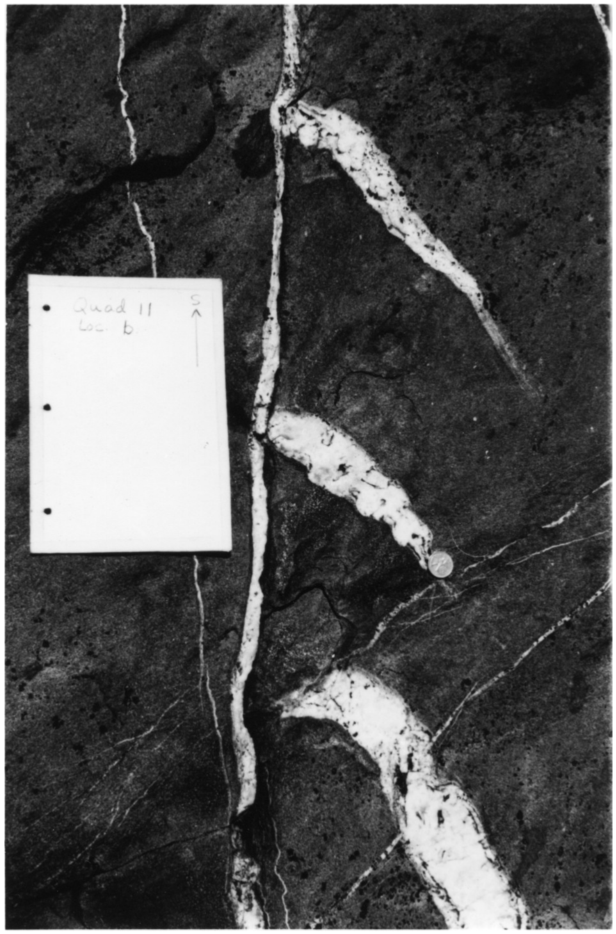
VEINS

As seen in the outcrop map (plate 1), the Mispec Beach area is extensively cut by quartz veins. These veins are developed in the two coarse lithologies, the sandstones and pebbly sandstones, and in the fine mudstones. There are very few veins in the siltstones. A brief description of these veins is given in Chapter 3, a more complete description follows.

MACROSCOPIC DESCRIPTION OF VEINS IN SANDSTONES

GEOMETRY. - As seen on the outcrop map (plate 1), most of the quartz veins are developed in the sandstones. The most striking feature of plate 1 is a large, conjugate shear set of *en echelon* veins (loc. H,4,a). Many of the other veins in the sandstones are *en echelon* sets, with the general trend of the sets being 180 and 125 degrees (parallel with one or the other of the *en echelon* array in the conjugate pair). Some of the larger sigmoidal sets have been cut through by very thick quartz veins (plate 1, locs. 10-M and 11-M). At location (E,2,a), individual sigmoidal veins widen in the center such that they begin to coalesce with adjacent sigmoids. All of the veins in the outcrop dip either vertically, or steeply (60 to 90 degrees) to the south. Although the most common style of vein in the sandstones is the sigmoidal *en echelon* set, there are also

Figure 5.1. - Ladder structure resulting from a straight vein connecting a series of sigmoidal veins. (15cm. wide board for scale)



many narrow veins which are straight and vertically dipping, with trends parallel with one or the other of the shear zones in the conjugate shear sets (loc. J,8,a). These straight veins commonly occur at the margins of en echelon sets, forming ladder-like structures (fig. 5.1), described as 'feather structures' by Roering (1968). Another common configuration is a sigmoidal vein shape bisected by a straight vein (e.g. loc. J,9,a). All of these vein geometries are well developed in the outcrop, and most are consistent in orientation with the conjugate shear set. This consistency suggests that they are all the result of the same deformation event.

The terminations of the veins vary greatly. Some trail off in a single thin fracture, while others splay off in complex, feathering veinlets (fig. 5.2). Sigmoidal en echelon sets are commonly truncated on one side by a straight vein, so that they terminate abruptly at one end and taper off into a curved vein at the other end (fig. 5.3). In other arrays, a few veins terminate at vein intersections, while some in the same set are offset by the crosscutting veins (fig. 5.4).

Many of the veins are fibrous. The fibers at the ends of the veins are commonly intact, and their orientation can be measured in the field (fig. 5.5). Toward the center of the veins, the quartz is commonly massive. Most of the fibers seen in the field are perpendicular to the walls of the veins.

Figure 5.2. - Feathering termination of a quartz vein. (15 cm. wide board for scale)

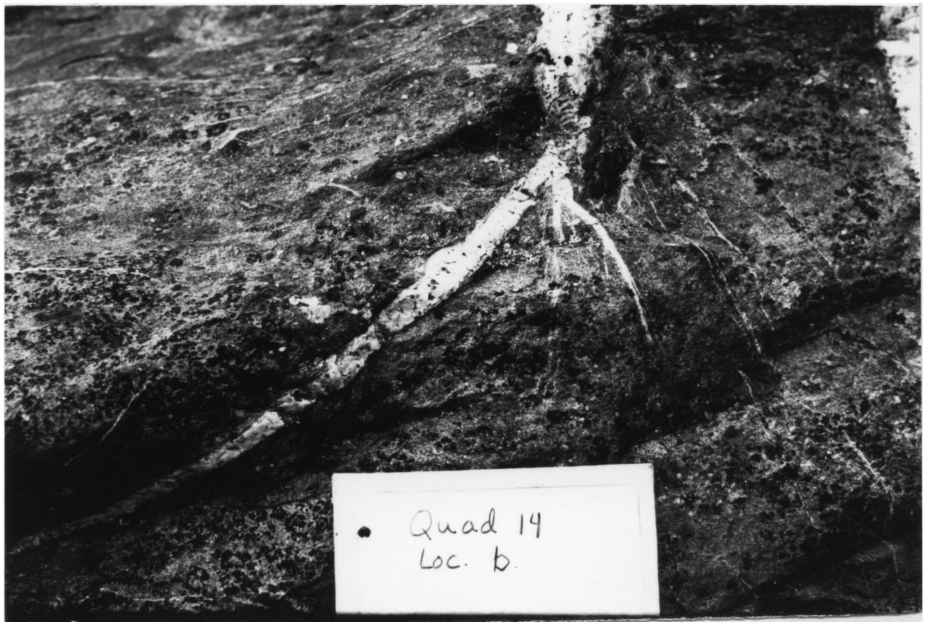
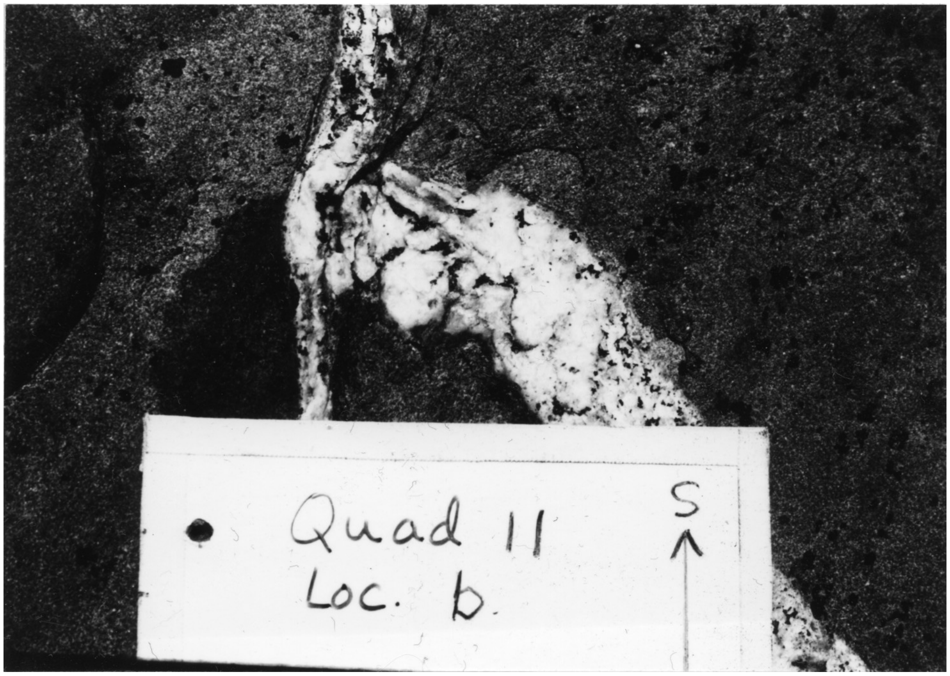


Figure 5.3. - Set of curved quartz veins truncated on one side by a straight vein. (15cm. wide board for scale)



<u>Figure 5.4</u>. - An <u>en echelon</u> vein set in which some sigmoids are truncated by later veins, and others are nearly cross cut. (15cm. wide board for scale)

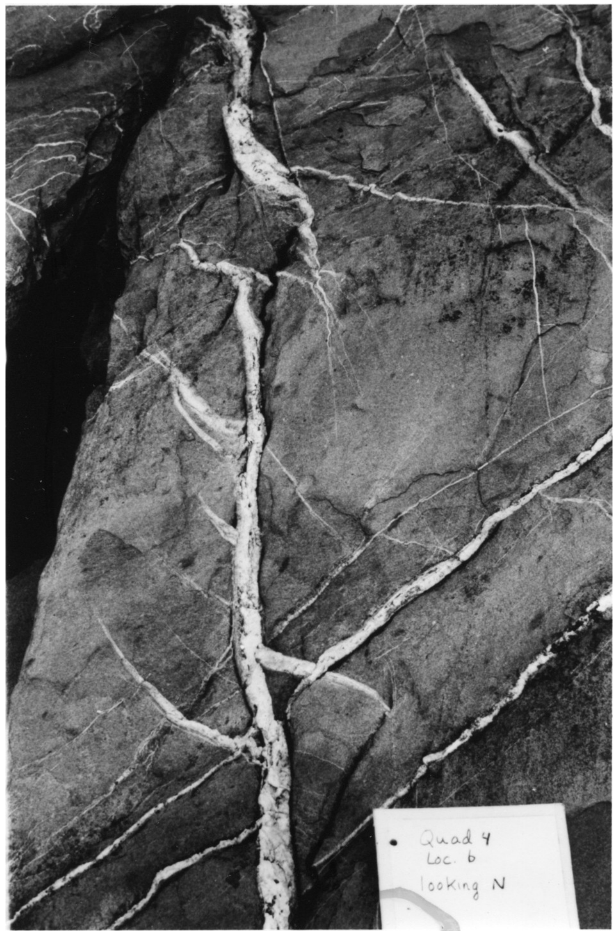
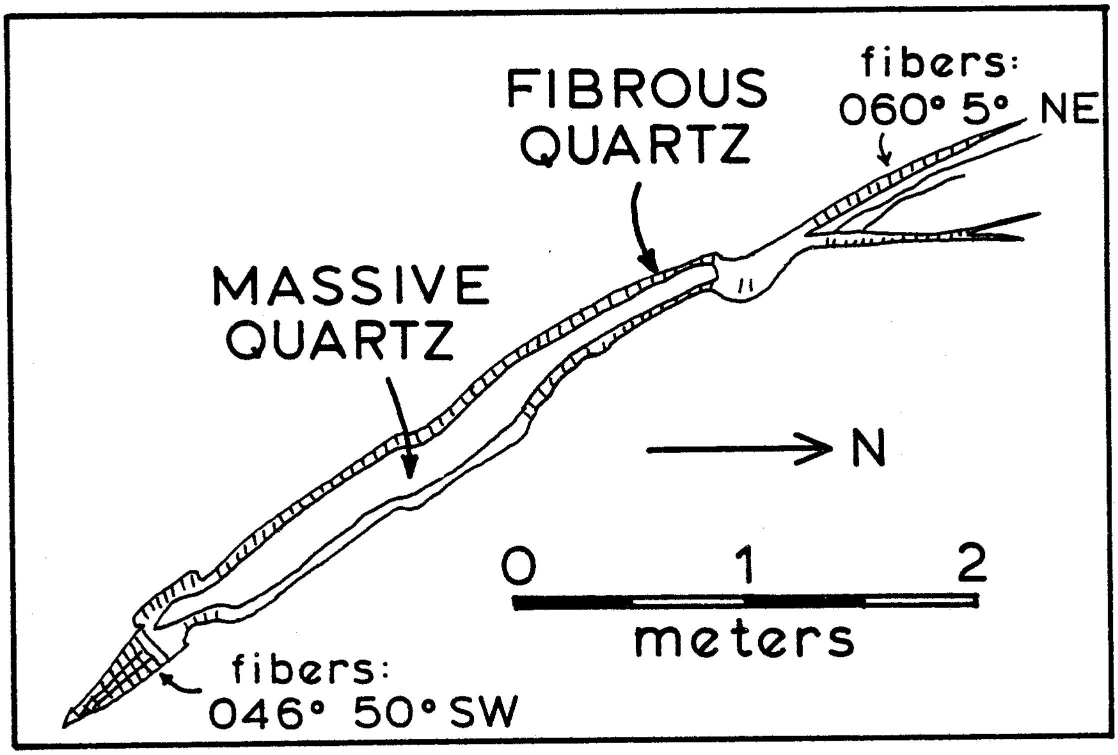


Figure 5.5. - Field sketch of fibers at the ends of a vein.



MICROSTRUCTURE OF QUARTZ VEINS IN SANDSTONES

Thin veins, 10 cm wide and smaller, were sampled. The feathering ends of a few small quartz veins were also collected. The very wide quartz veins (up to 1.5 meters) in the sandstones were not sampled.

MINERALOGY. - The mineralogy within the veins in the sandstones is very consistent. The major mineral is quartz which is present in large cracked and strained grains, and smaller recrystallized, strain-free grains. Some dolomite is also present (0-5%), commonly filling cracks between quartz grains, and as isolated rhombs surrounded by quartz (fig. 5.6). Many rhombs have been altered or deformed, but they are still readily recognized. Chlorite and a few opaque grains (hematite and magnetite) are also present.

GEOMETRY. - In the sandstones, vein walls are very irregular. The centers of the veins are typically fibrous, while the margins are occasionally lined with many small quartz grains (fig. 5.7). This concentration of tiny grains, together with a high concentration of quartz within the wall rock, obscures the contact between vein and wall rock. The contact can be best observed under plane light.

These veins tend to feather out at their ends, and disappear into sandstones. In the feathering zones, the change from vein to wall rock is transitional (fig. 5.8).

Figure 5.6. - Thin section photograph of dolomite rhombs in quartz veins.

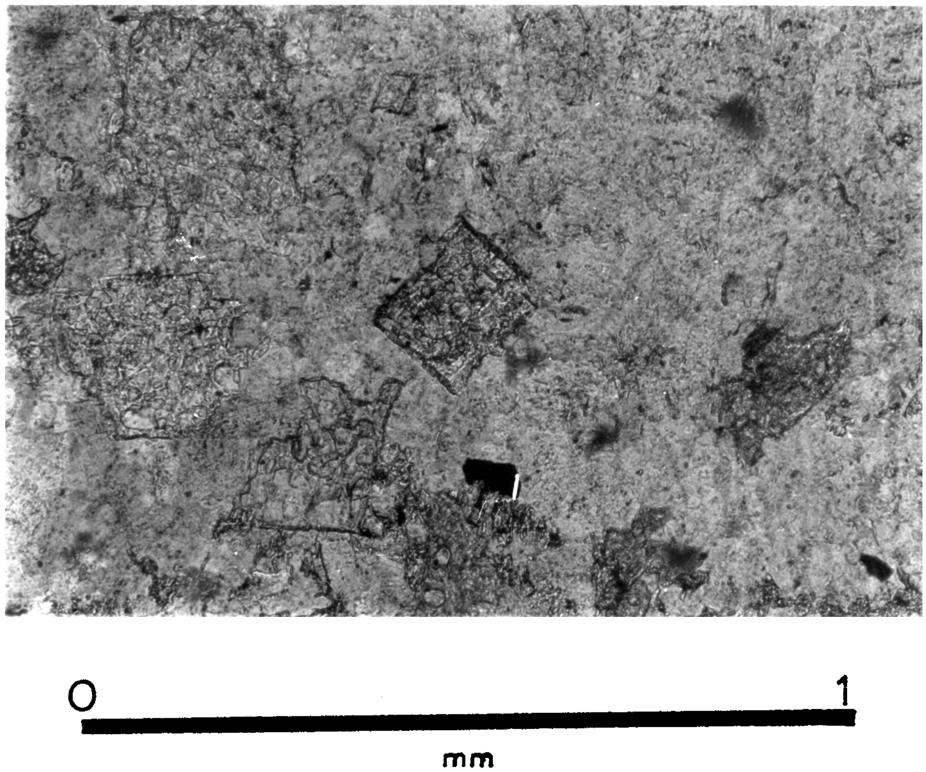


Figure 5.7. - Thin section photograph illustrating change in grain size from the wall-vein contact towards the center of the vein.

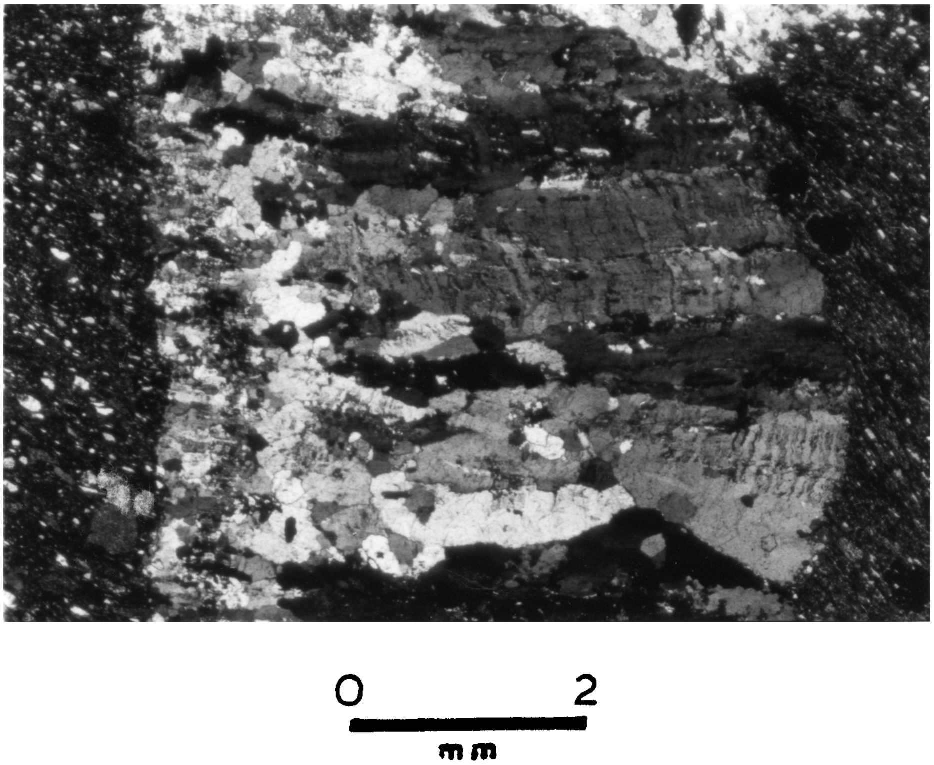
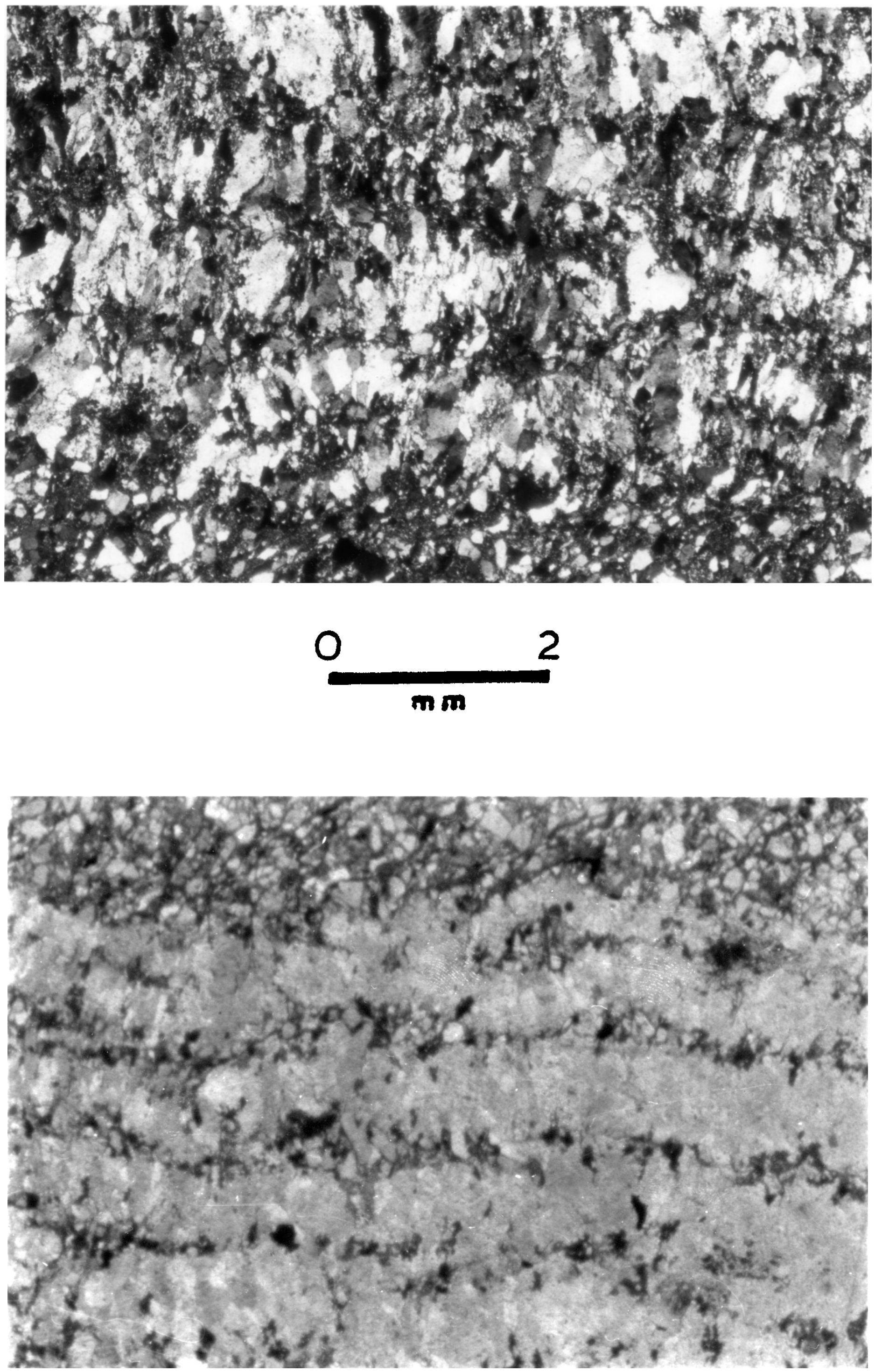


Figure 5.8. - Thin section photographs of the transition from vein to wall rock at a feathering vein termination in sandstone. (A) Plane light; (B) Crossed nicols.



Fibers within the veins are typically perpendicular to the walls. In some veins, the fibers are curved perhaps indicating a change in principal strain orientation during opening (Durney and Ramsay 1973).

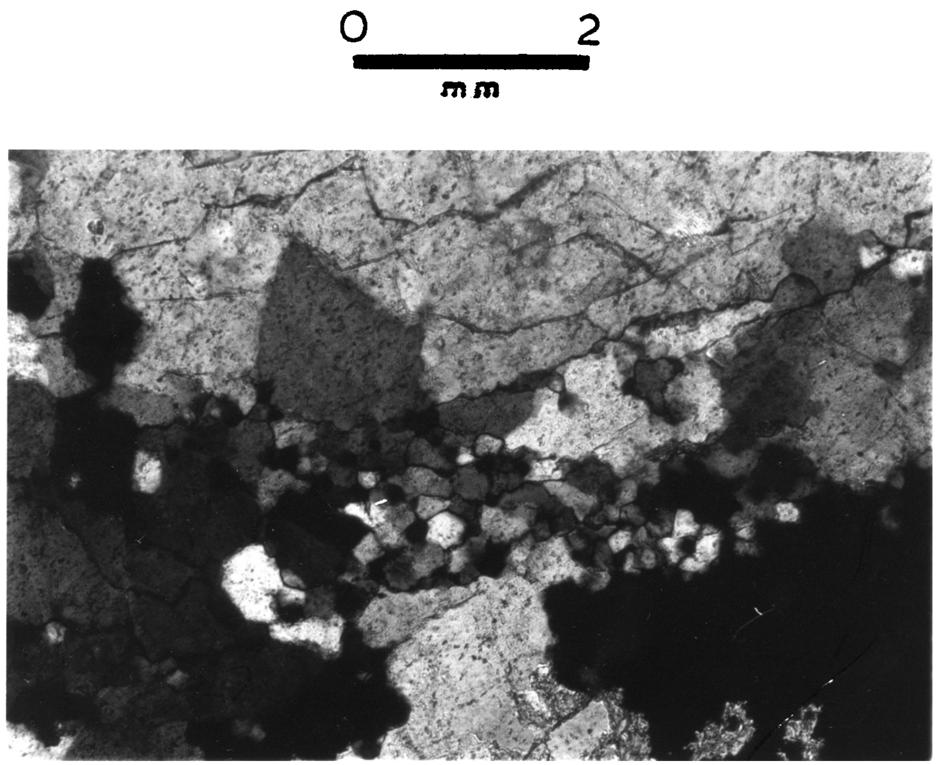
RELATIVE TIMING OF VEIN FORMATION IN THE SANDSTONES AND DEFORMATION. -From field observations, it is difficult to determine the sequence of vein development in the outcrop. Figure 5.9 gives evidence supporting synchronous growth of propagating veins in several orientations during one event. In this photo, veins in three orientations crosscut one another in such a way that no vein can be shown to be older or younger than the others. These complex crosscutting relationships are seen across the whole outcrop.

The fact that the slaty cleavage is crosscut and deformed by the veins (fig. 3.4) shows that the veins are younger. Yet, in a few areas the veins seem to be deformed along the cleavage planes, suggesting that the veins have formed before or during a late phase of slaty cleavage development. The relationship between the veins and the crenulation cleavage cannot be determined in the field since the crenulation cleavage is very poorly developed in the rocks having a high concentration of quartz veins.

Microscopic observations of the veins show that they are somewhat deformed. The boundaries and subboundaries of quartz grains within the veins have concentrations of small, recrystallized grains along them (fig. 5.10). Most of the

Figure 5.9. - Propagating network of veins that crosscut one another such that no vein can be established as younger or older than the rest. (15cm. wide board for scale)

Figure 5.10. - Thin section photograph of recrystallized quartz grains along the boundaries of fibrous grains.



elongate quartz fibers are strained and cracked. In a few veins, the fibers are somewhat bent, resulting in undulatory extinction (fig. 5.11). Some of the large quartz grains have deformation bands. Calcareous minerals within the veins are twinned.

The early slaty cleavage is truncated by the veins, indicating that they are synchronous with or younger than the cleavage. In some areas, the veins are deformed along cleavage planes, suggesting a later phase of cleavage development or some later phase of shear along the S_1 planes. The small quartz grains around the walls of the veins may suggest two phases of opening, (or some later deformation), resulting in recrystallization of quartz along the walls of the veins. Many of the veins are folded into small folds (fig. 5.12), that are responsible for changes in the orientation of the quartz fibers. The very ends of some of the veins are accentuated by a concentration of cleavage surfaces wrapping over them. These zones, as well as some zones of offset within the veins, support the idea that there was further deformation on the slaty cleavage planes after the veins were formed.

POSSIBLE MODE OF ORIGIN. - Within the sandstone beds, there is no clear evidence for crack-seal opening of the vein. The walls of the vein are very irregular, probably as a result of vein material growing off quartz grains along the wall of the veins (Ramsay, 1980). There is some microstructural evidence of a replacement origin for the

Figure 5.11. - Thin section photograph of bent quartz fibers in a vein in sandstone.

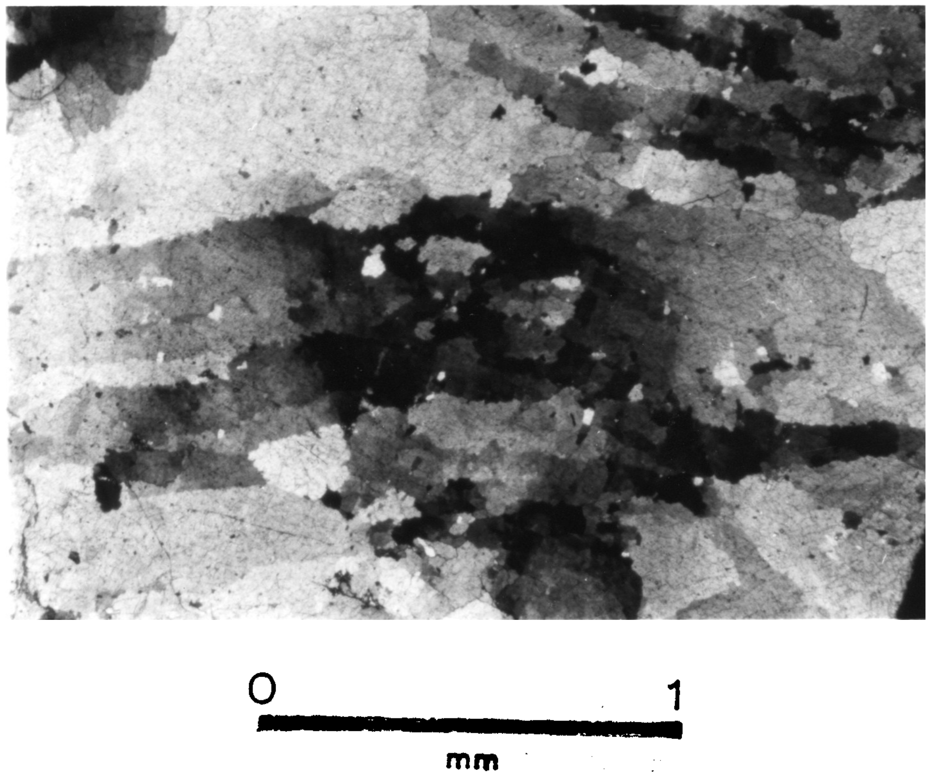
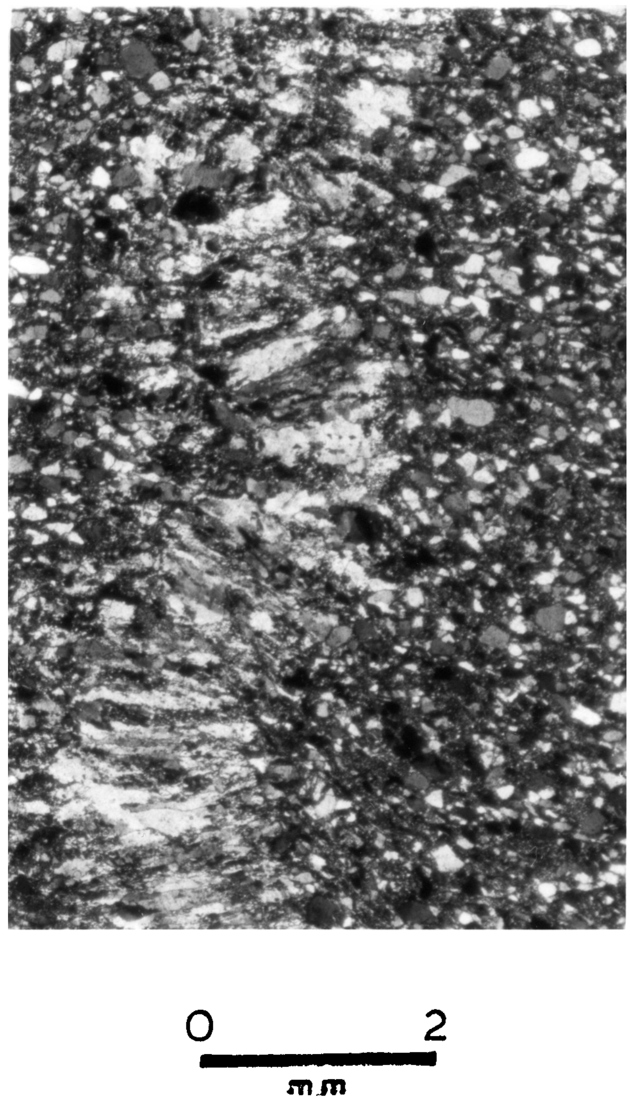


Figure 5.12. - Thin section photograph of a gently folded vein.



veins where zones of wall rock appear to continue through the veins (Hobbs et al. 1976, p. 293); and in outcrop many of the veins in the siltstone beds appear to be only concentrations of quartz along a line since all the wall rock componants can be traced through the veins. However, there is no conclusive evidence to support replacement as a mode of formation of these veins either. The veins may have developed with components of both replacement and crack-seal; more data are required to determine in detail the mode of formation of the veins.

MACROSCOPIC DESCRIPTION OF VEINS IN MUDSTONES

GEOMETRY. - In their geometry, the quartz veins in mudstones are very different from those in sandstones. They are straight in map view rather than sigmoidal, and parallel trending about 030. They appear to be about 5 cm wide, but on closer examination, each band appears to be an amalgamation of several zones, each from 0.5 to 3 cm thick (see fig. 3.7). These veins outcrop at the surface of the mudstone beds as steeply dipping, closely spaced, parallel structures (loc.E,3,c). Within 2-3 cm of the upper surface of the beds, they bend abruptly, and then continue dipping shallowly to the southeast (see fig. 3.7).

Some of the veins appear to be fibrous, with the fibers oriented perpendicular to their walls. Because there is limited exposure of mudstone in the outcrop (plate 1) no conclusive data on the orientations of the fibers in the mudstones was collected.

The three dimensional shape of these veins is very complex. Many of them fork, with the branches of the fork diverging at angles of up to 90 degrees.

MICROSCOPIC DESCRIPTION OF VEINS IN MUDSTONE

MINERALOGY. - The veins in the mudstone are quartz veins containing some dolomite, chlorite, and opaques.

GEOMETRY. - The geometry of the veins is very complex. Along the margins of the veins splays and branches are common. These branches commonly run parallel with the main vein for some distance, and then diverge at some acute angle (fig. 5.13). In zones where a branch and a main vein are parallel, the branch is typically separated from the main vein by either a thin film of wall rock, or it is marked by an abrupt change in quartz grain size (fig. 5.14). The contact between wall rock and vein is generally very sharp. Commonly, the contact is somewhat sharper on one side of the vein than the other.

While some quartz within these veins is fibrous, many of the grains, especially in the center of the veins, are equigranular with no preferred orientation. In some branching veins, one splay is fibrous while the other is not. Some fibers within these veins are indicative of slightly oblique opening, as they are not exactly

Figure 5.13. - Thin section photograph of a quartz vein (top center) that diverges from the main body of the vein (bottom right to top left).

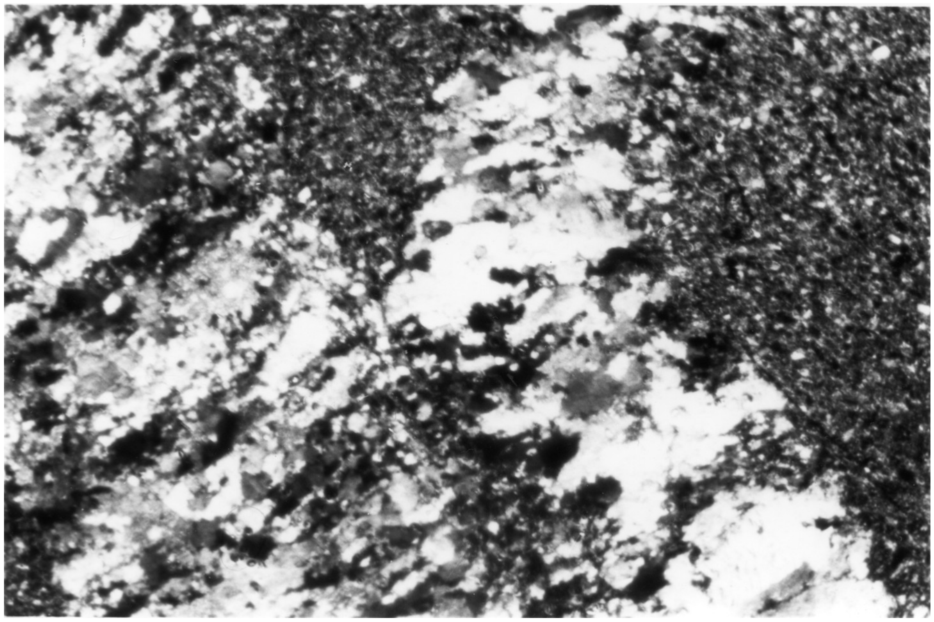
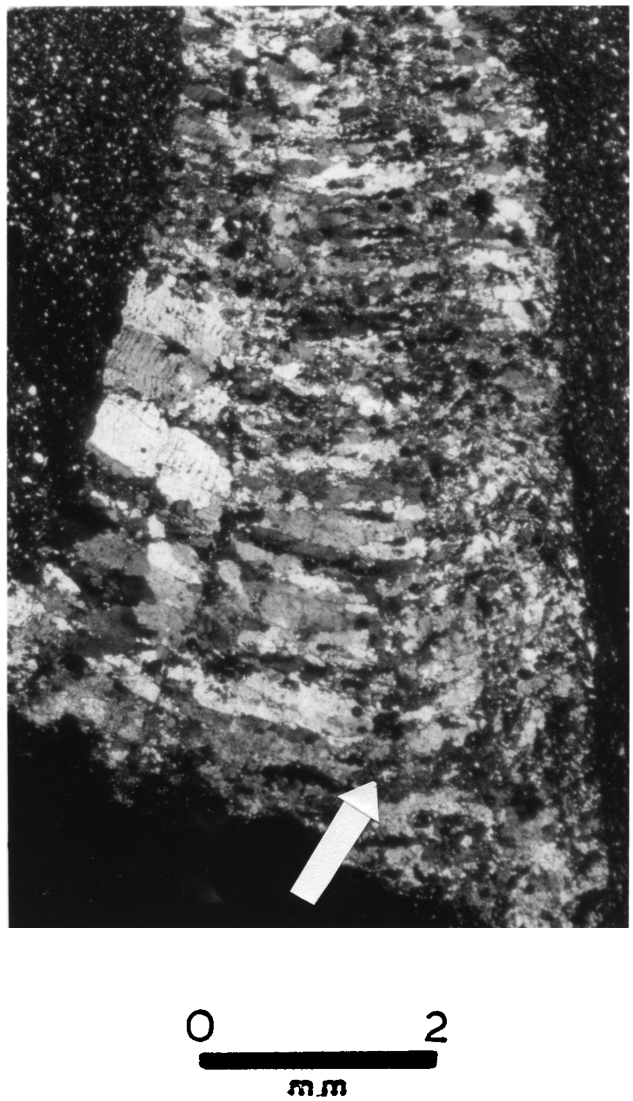


Figure 5.14. - Thin section photograph illustrating a change in grain size between two parallel veins.



perpendicular to the walls of the veins.

RELATIVE TIMING OF VEINS FORMATION AND DEFORMATION.

Throughout the outcrop the slaty cleavage appears to curve into quartz veins, suggesting some deformation of the cleavage during the formation of the veins (fig.5.15). The relative timing of crenulation cleavage and quartz vein development cannot be determined from outcrop evidence.

The red mudstones are generally overlain by siltstones. At a well exposed contact surface between the two lithologies at locality (E,3,b), veins in mudstone widen into clots, and penetrate only a few centimeters into the adjacent siltstone (fig. 5.16). The zigzag appearance of the quartz veins within the siltstone bed suggests that both the slaty cleavage and the crenulation cleavage were present when the veins formed. Since the quartz veins in the siltstone are parallel with the slaty and crenulation cleavage, I think that quartz may have filled in along these surfaces. It is possible that the quartz in these cracks is associated with a late stage of vein filling. An alternative explanation for the folded appearence of the quartz veins is that a vein which formed in the rock was deformed by later slip along crenulation cleavage planes to produce fold forms.

The slaty cleavage in the mudstones around the veins is strongly disrupted (fig 5.17). Near the veins, the cleavage is commonly folded, and the dark minerals defining

Figure 5.15. - Deformed cleavage along the walls of a quartz vein.

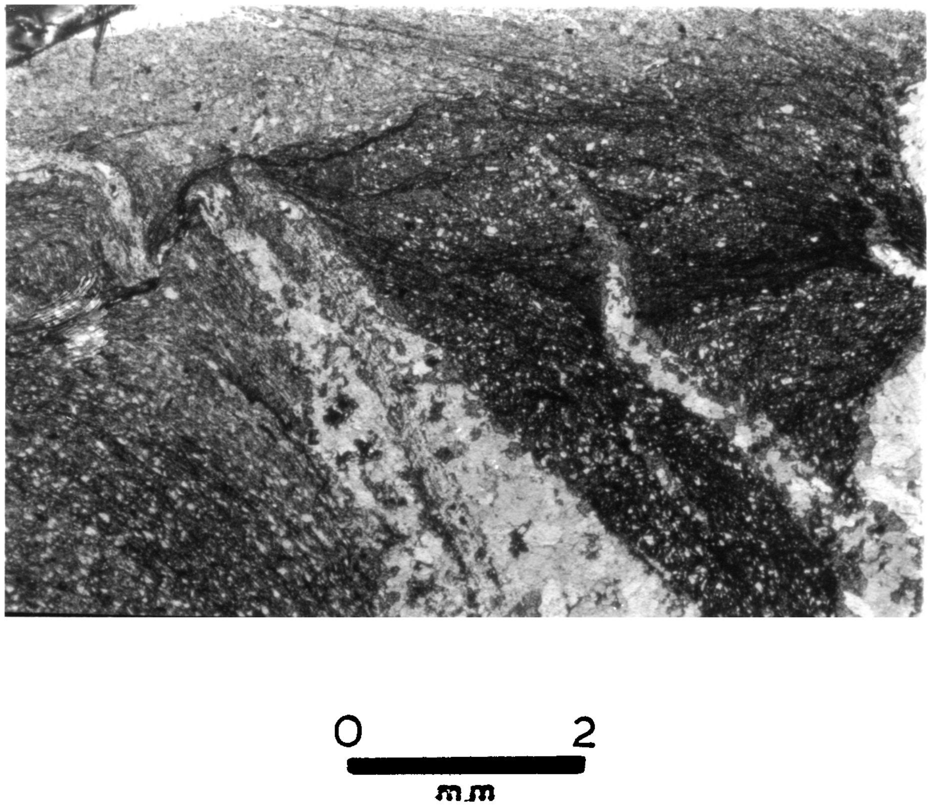


Figure 5.16. - Quartz veins in mudstone which do not penetrate far into the siltstone bed above. The quartz seems to accumulate in clots at the base of the siltstone.

Figure 5.17. - Cleavage planes changing orientation and curving into veins.

the foliation are very concentrated. The cleavage is sheared along the edges of the veins, and commonly wraps around their terminations (fig. 5.18). The quartz veins are deformed by the crenulation cleavage (fig. 5.19), however, the crenulation cleavage does not exist in islands of wall rock surrounded by vein material. This relationship supports the idea of a crenulation cleavage that is younger than the veins, and that could not form in zones where the wall rock was protected by surrounding vein material.

The quartz is present in large, cracked and strained grains, as well as in small, recrystallized strain-free grains. The recrystallized quartz is generally present along the boundaries of the older quartz veins. Dolomite fills racks between quartz grains and appears to replace some of the quartz in the grains. The chlorite is fibrous and appears to fill cracks between grains. There is a band of carbonate material with chlorite around the edges of some of the veins (fig. 5.20), whose presence at the edge of the veins is suggestive of precipitation during a separate phase of opening of the veins.

POSSIBLE MODE OF ORIGIN. - These veins show very clear evidence of a crack-seal origin (Ramsay, 1980). Many quartz grains include thin bands of wall material that are exactly parallel with the walls of the veins (fig. 5.21).

Figure 5.18. - Thin section photograph of slaty cleavage concentrated at the tip of a quartz vein.

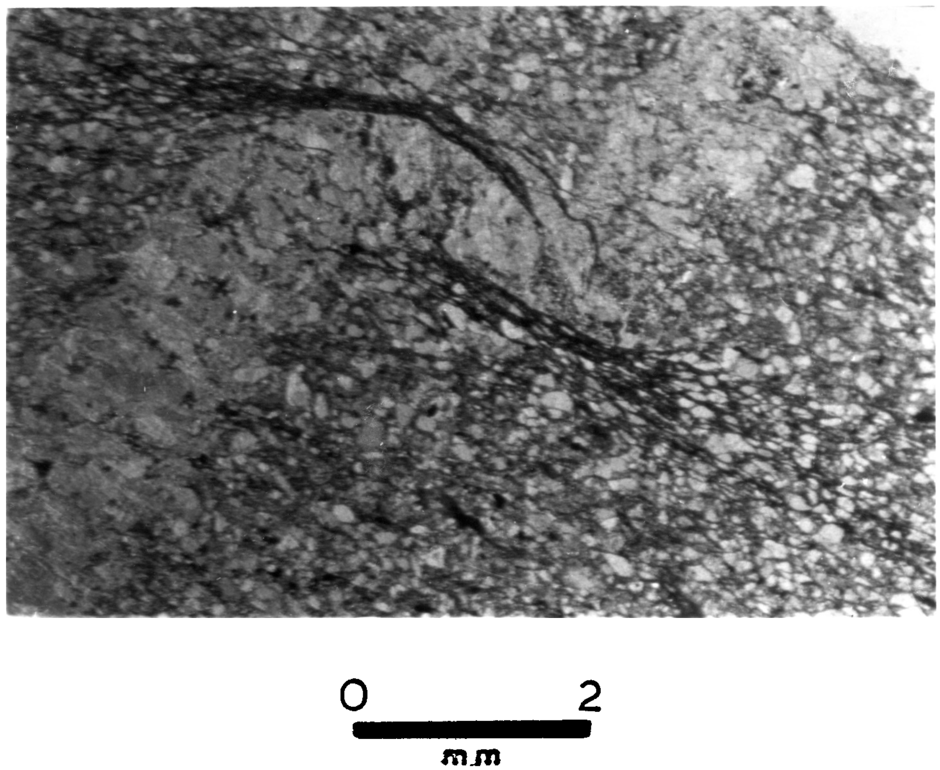


Figure 5.19. - Thin section photograph of crenulation cleavage deforming a quartz vein.

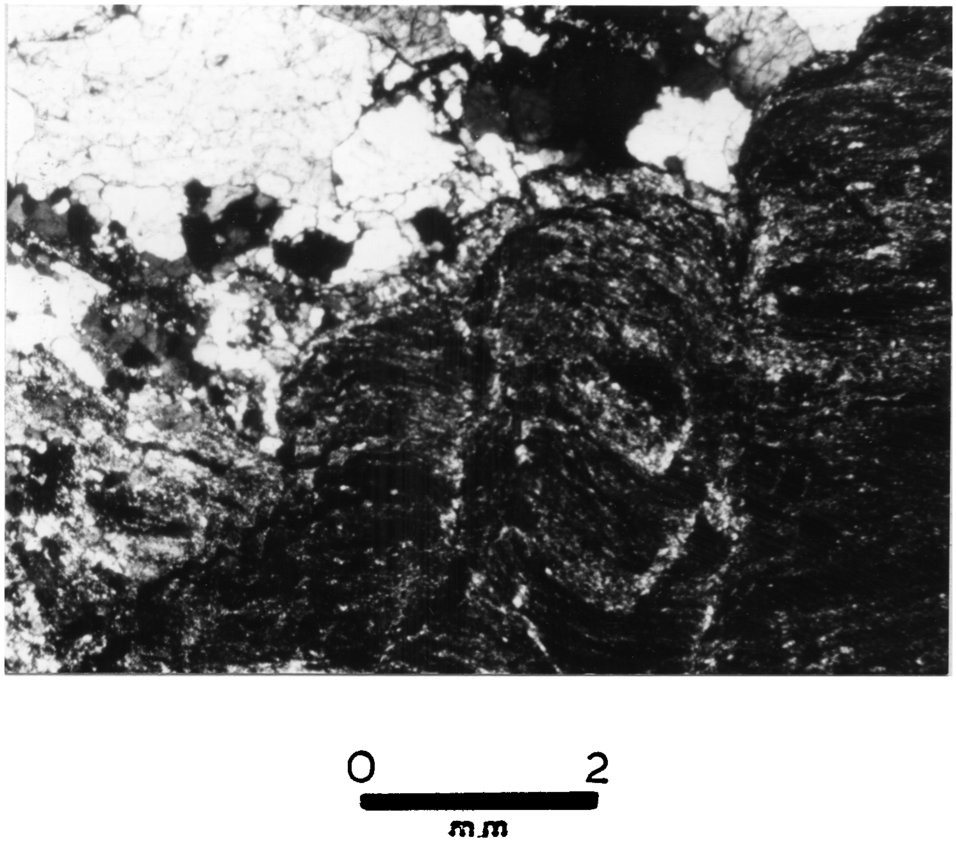


Figure 5.20. - Band of calcareous mineral at the margin of a vein.

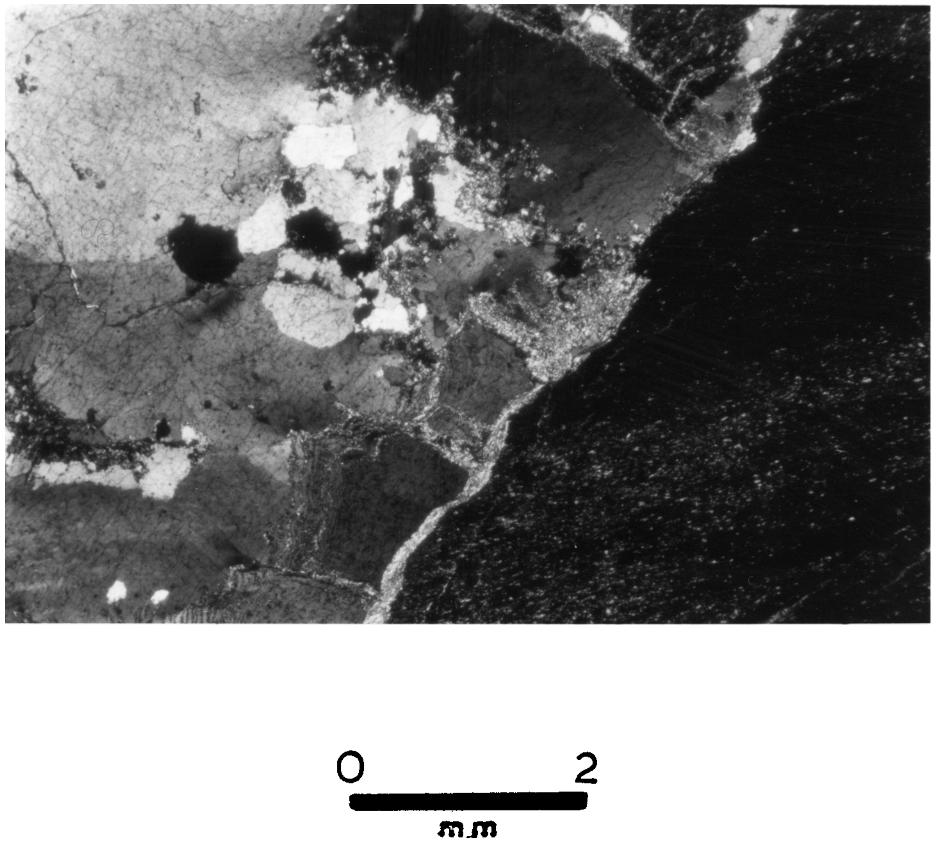
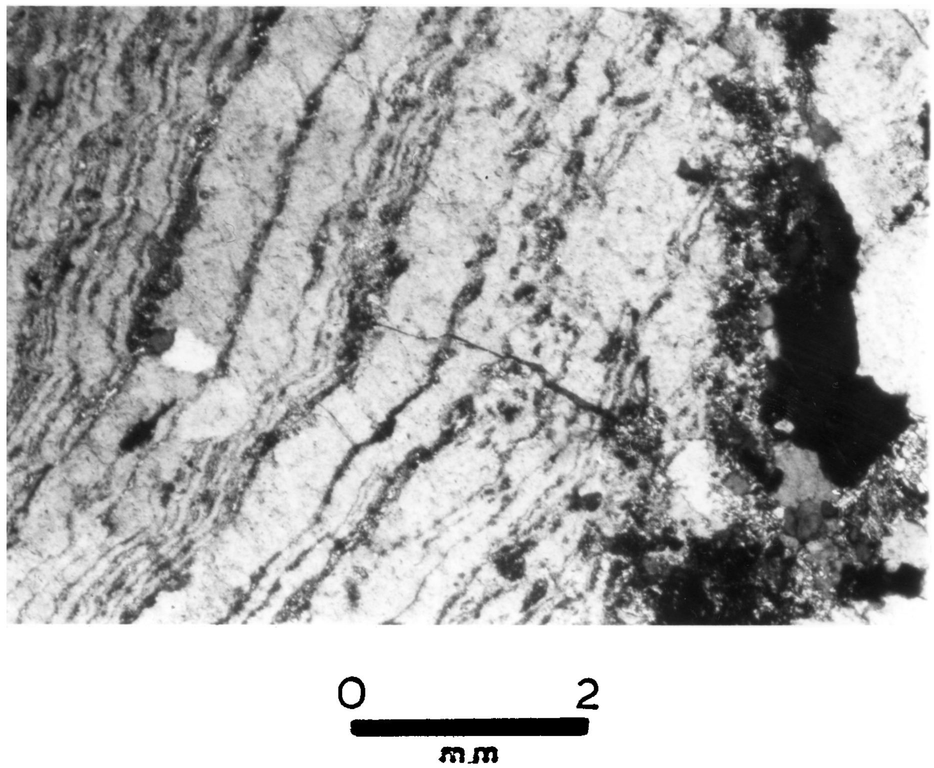


Figure 5.21. - Thin section photograph of wall rock inclusion bands in a quartz vein.



CORRELATION BETWEEN VEINS IN SANDSTONE AND MUDSTONE

The vein development in the sandstones and mudstones appears to be synchronous. Both vein sets deform the slaty cleavage and are deformed by the crenulation cleavage. They both appear to have formed during the late stages of slaty cleavage development, as is indicated by the concentration of slaty cleavage planes around the ends of the quartz veins, and the displacement of some quartz veins along S_1 planes.

CHAPTER 6

CORRELATION OF STRESS AND STRAIN RELATED STRUCTURES

BEDDING. - An orientation diagram of poles to bedding (fig. 6.1) shows that either bedding is neglibly folded within the Mispec Beach area or that it is isoclinally folded. The latter is ruled out by: (1) regional evidence cited in Chapter 2 that there are no major fold hinges known in the Mispec Nappe, and (2) local evidence cited in Chapter 3 that bedding at the Mispec Beach outcrop is upright. The structures developed within individual beds are evidence of the deformation the rocks have experienced.

PEBBLES. - The elongation and preferred orientation of pebbles and related quartz beards (which parallel the elongation direction) can be used as a crude estimate of the orientation and relative magnitudes of the principal strains. Field measurements of these structures are plotted in figure 6.2. The principal elongation direction (λ 1) is determined from an estimated average of the strong concentration of data in figure 6.2. Many of the pebbles are approximately cigar shaped (k approaching infinity, Flynn 1962), suggesting that the magnitudes of λ 2 and λ 3 are nearly equal and small with respect to λ 1 (Hobbs et al. 1976, their figure 1.13a). Because there is no obvious preferred orientation of the short axes within the pebbles, the λ 2 and λ 3 directions can only be determined to lie perpendicular to one another in a plane normal to λ 1 (fig. 6.2).

Figure 6.1. - Lower hemisphere equal area projection of 33 poles to bedding. The bedding is only slightly folded, resulting in a tight cluster of poles.

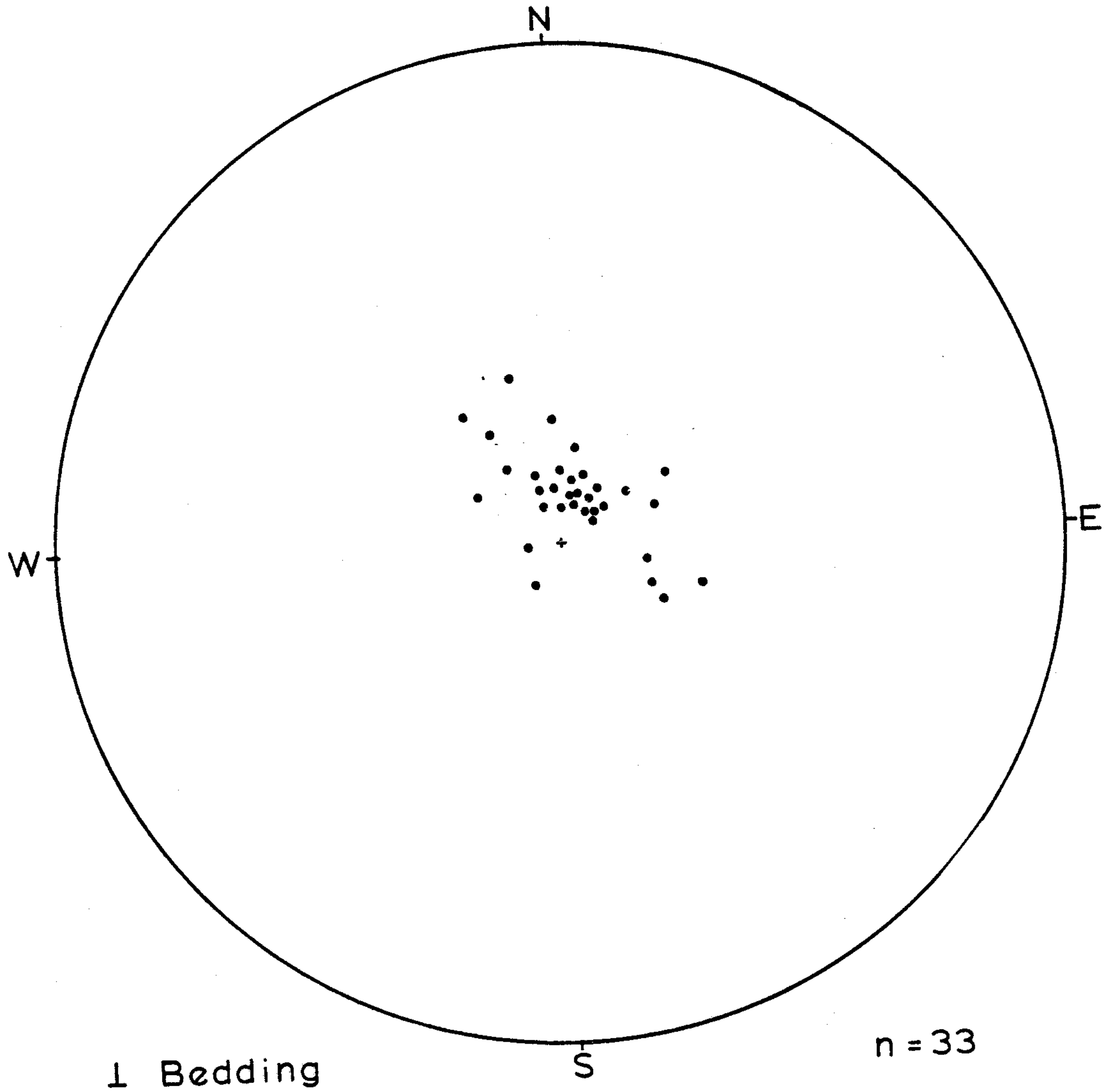
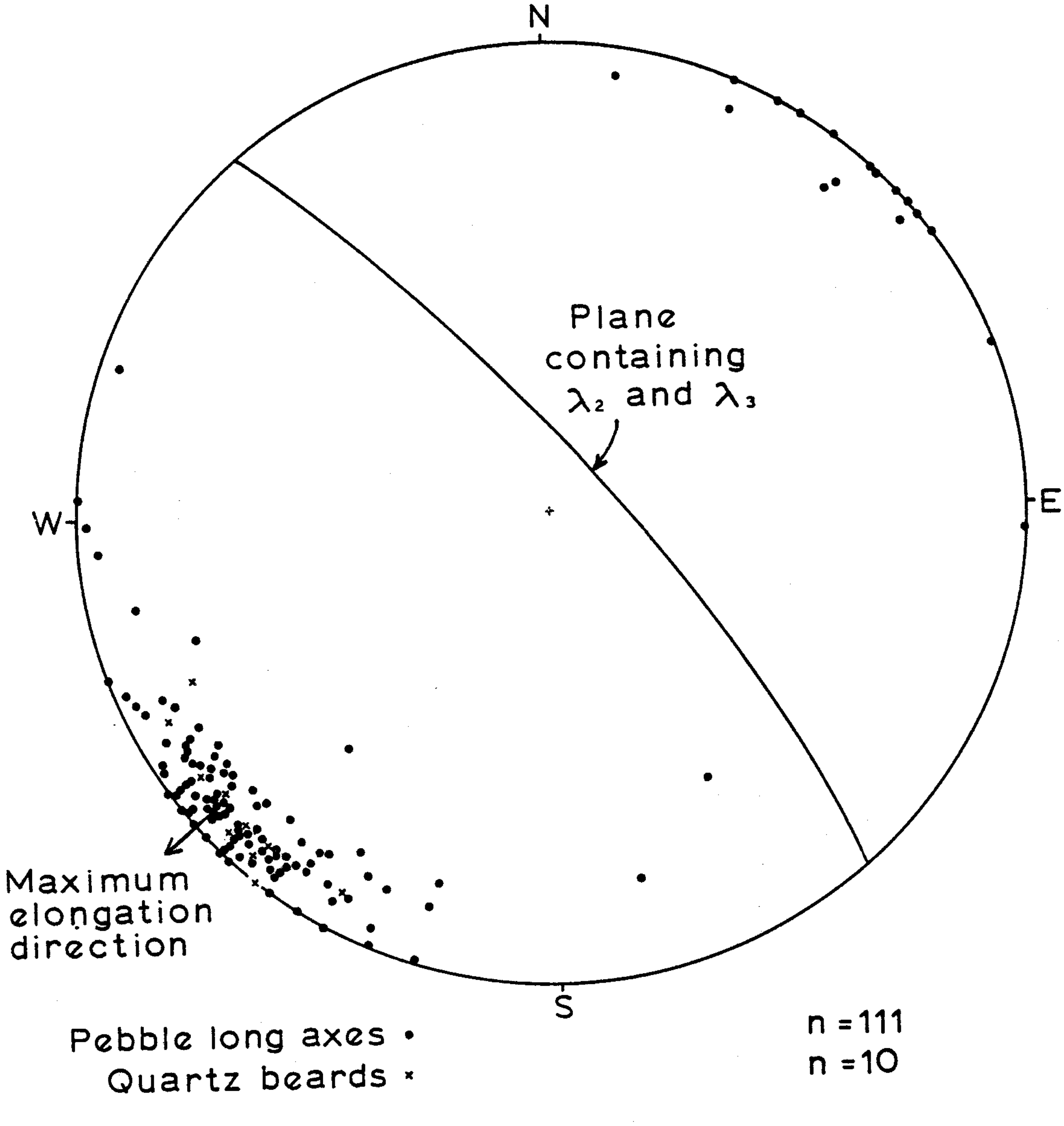


Figure 6.2. - Lower hemisphere equal area projection of 111 long axes of pebbles and 10 fiber orientations in associated quartz beards, showing the approximate λ 1 direction and a plane normal to λ 1 that should contain λ 2 and λ 3.



It should be noted that the observations made about the relative magnitude and orientation of strain from these pebbles are only estimates. No quantitative data on the shapes of the pebbles (intermediate or short axes orientations) was collected. These observations assume no volume loss during the deformation and nearly homogeneous deformation. The presence of a well developed slaty cleavage in the rocks (which appears to be at least to some extent a solution feature) and the pitting of grains between cleavage planes, suggests that the sandstone matrix around the pebbles underwent volume loss. The deformation resulting in the elongation of pebbles and the development of quartz beards at the ends of some clasts is not homogeneous. Mudstone clasts are readily deformed into long narrow ellipsoids, indicating that these clasts may have deformed in a nearly homogeneous fashion with respect to the surrounding matrix. However, vein quartz and metamorphic clasts appear to be much more competent than the surrounding matrix, and tend to be rotated into alignment, or to remain unaligned with respect to the mudstone clasts, and develop quartz pressure shadows in response to the matrix stretching around them. In the matrix surrounding these clasts, a well developed slaty cleavage curves around the clasts, and is absent in the adjacent sandstone. This suggests that these pebbles are acted as rigid objects in a ductilely deforming matrix. The orientations of the principal strains near these clasts are probably quite variable (Ramsay 1967, p.

181).

SLATY CLEAVAGE. - An orientation diagram of poles to slaty cleavage (fig. 6.3) shows a strong cluster of poles with an estimated average orientation perpendicular to a plane striking 064 and dipping 22 degrees to the southwest. Assuming that the slaty cleavage resulted primarily from pressure solution displacements perpendicular to the cleavage planes, then the pole to the slaty cleavage plane would represent the maximum shortening direction (λ 3) of the deformation that formed the cleavage (Ramsay 1967, p. 180). Support for this assumption is seen in thin sections where S_1 is defined by a strong preferred orientation of layer silicates and sand-sized quartz grains. Many of the sand-sized grains have pressure shadows around them, which are commonly symmetric and parallel with the slaty cleavage. Other evidence that the slaty cleavage formed as a result of shortening parallal to λ 3 is the apparent dissolution of grains now truncated along the cleavage planes and solution pitting of quartz grains perpendicular to the cleavage planes (described in Chapter 4). The slaty cleavage surfaces show a well developed crenulation lineation and a moderately well developed mineral lineation (see Chapter 4). Figure 6.4 is an orientation diagram of the mineral lineation in S_1. This lineation orientation is coaxial with the λ 1 orientation determined for the pebbles in the sandstone. Thin section observations of the minerals forming the lineation show that they are small

Figure 6.3. - Lower hemisphere equal area projection of 72 poles to slaty cleavage. The average trend and plunge (064, 66 NW) should represent the average orientation of λ 3. λ 1 and λ 2 directions should be contained within the average slaty cleavage plane.

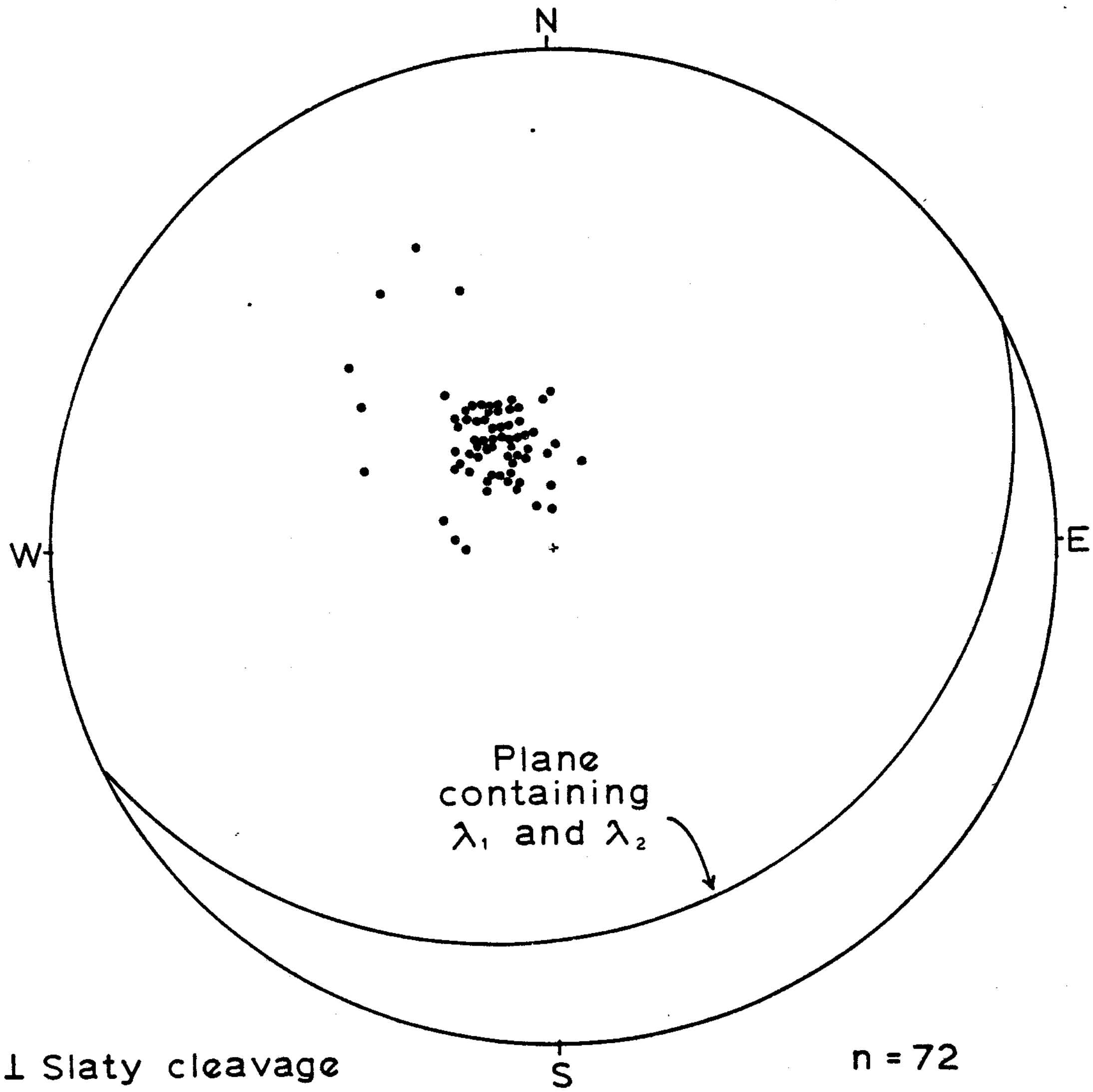
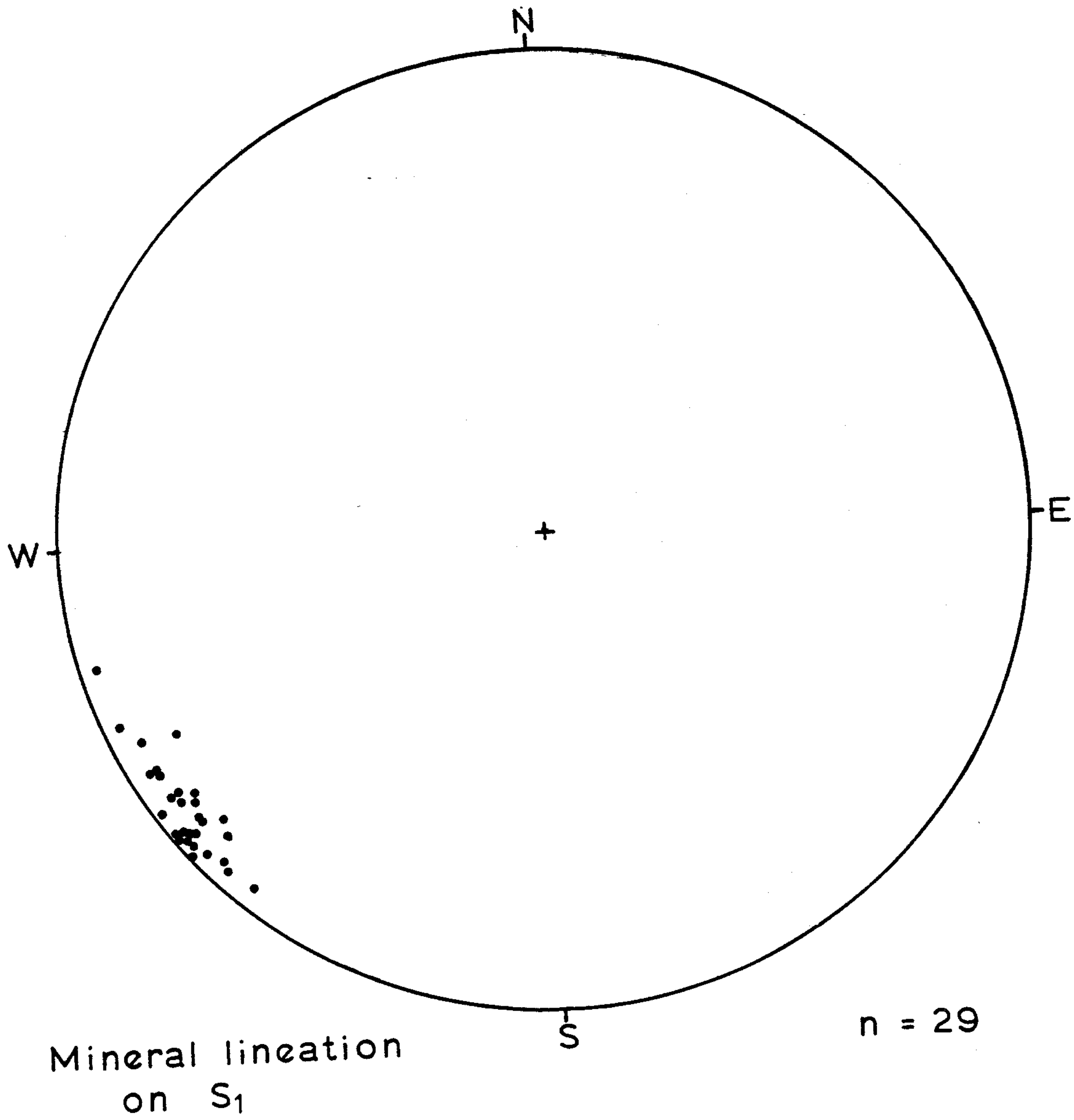


Figure 6.4. - Lower hemisphere equal area projection of 29 mineral lineations on slaty cleavage. Note that the orientation of this lineation coincided with the intersection of the slaty cleavage with the λ 1 direction determined from the pebbles (fig.6.2).



opaque clots (framboids ?) that appear to have grown throughout the slaty cleavage developement. Slaty cleavage traces can be followed through the clots yet others have pressure shadows around them which appear to be associated with the slaty cleavage (the long axis of the shadows is parallel with the S1 foliation).

The mineral lineation may suggest that there is a preferred orientation of elongation within the slaty cleavage planes. Other evidence indicating a preferred orientation of elongation within the foliation plane is a preferred orientation of the micas within those planes. In thin sections oriented perpendicular to the inferred λ 3 direction and perpendicular to the λ 1 direction suggested by the (framboid ?) mineral lineation, micas are seen as small equidiminsional grains defining poorly developed pressure shadows. In sections cut perpendicular to λ 3 but parallel to the inferred λ 1 direction there are well debeloped pressure shadows of fibrous mica (with the long axes of the mica grains defining the shadows) around the quartz grains. Also seen in this orientation are other grains with pressure shadows of fine grained quartz around them.

Thus, from the slaty cleavage we can estimate the orientation of λ 3 and from the elongation direction indicated by structures within the foliation we can determine the orientation of λ 1. The intermediate strain axis, λ 2 is oriented perpendicular to both λ 1 and λ 3.

VEINS. - The quartz veins so extensively developed in the outcrop commonly occur in one of three orientations (fig.6.5). The largest of the three maxima on this diagram is the orientation of the poles to the tips of sigmoidal quartz veins in *en echelon* arrays and to veins oriented parallel with the tips. Figure 6.6 is an orientation diagram showing the orientation of only sigmoidal vein tips. The intermediate maximum (fig.6.5) is the pole to the veins that in the outcrop are oriented just east of north-south. These veins frequently trend along the margins ot the *en echelon* arrays producing ladder structures (see fig.5.1). The third maximum is a plot of the poles to the veins in the mudstone (loc.E,3,c). A few faint replacement veins in the siltstones also parallel this trend.

The tips of sigmoidal quartz veins within the *en echelon* sets in the northwestern part of the map (loc. F,4) dip approximately 55 degrees to the southeast (fig.6.6). The extension direction (perpendicular to the tips of the sigmoidal tension gashes) trends at 070-075 degrees. If we assume that these tension gashes are the result of brittle failure, then their arrangment in conjugate shear sets allows us determine the stress orientation during deformation. Anderson (1905) suggested, and it has since been experimentally documented (Griggs and Handin 1960, p. 222), that brittlely deforming rocks fracture in such a way that the acute angle of conjugate

Figure 6.5. - Lower hemisphere equal area projection of 186 poles to quartz veins. Contour intervals (per one percent area): 0 to 0.5% = light stipple; 0.5 to 1% = coarse stipple; 1 to 2% = vertical dashes; 2 to 4% = vertical lines; 4 to 8% = vertical and horizontal lines; 8% and above = black.

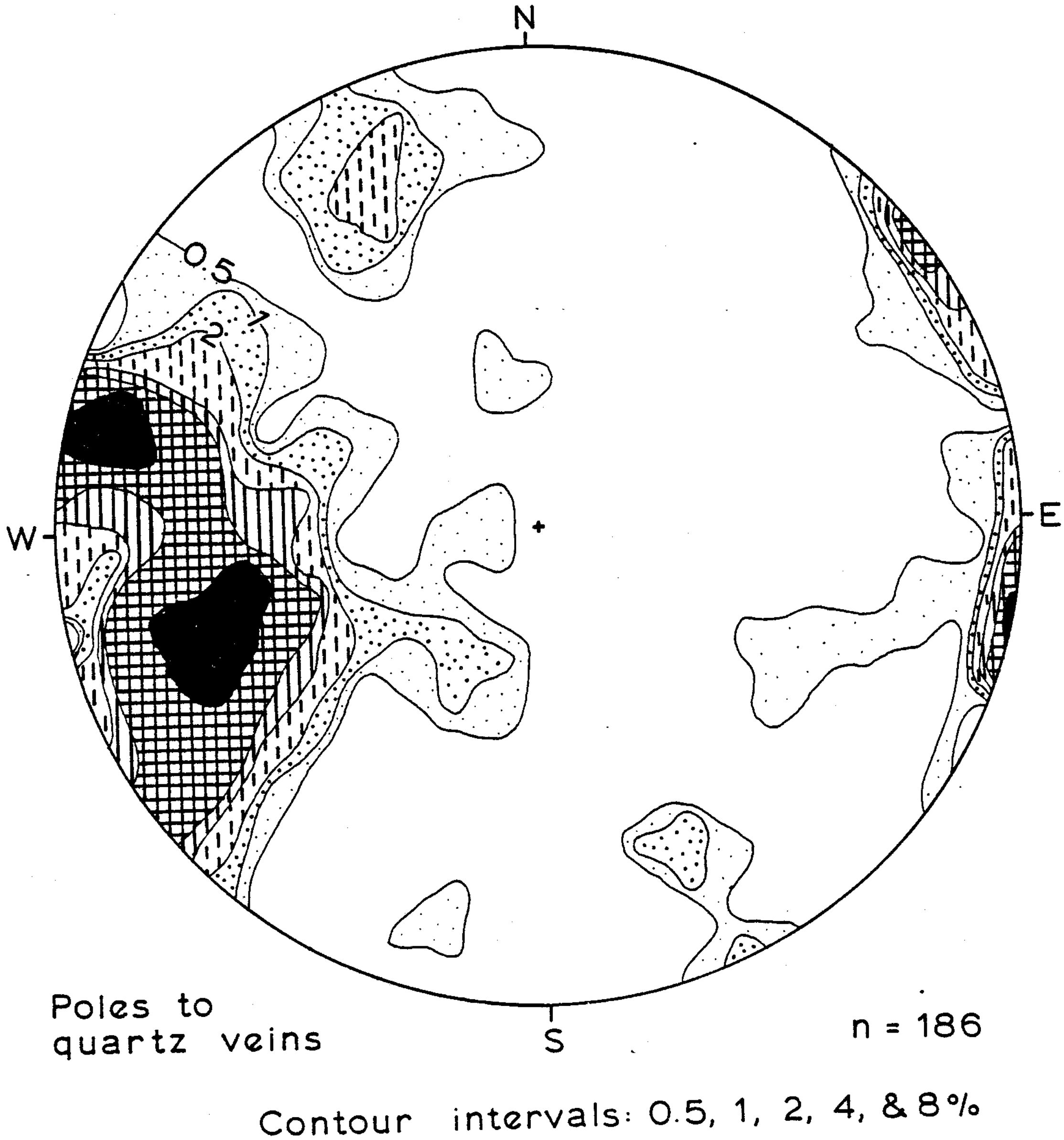
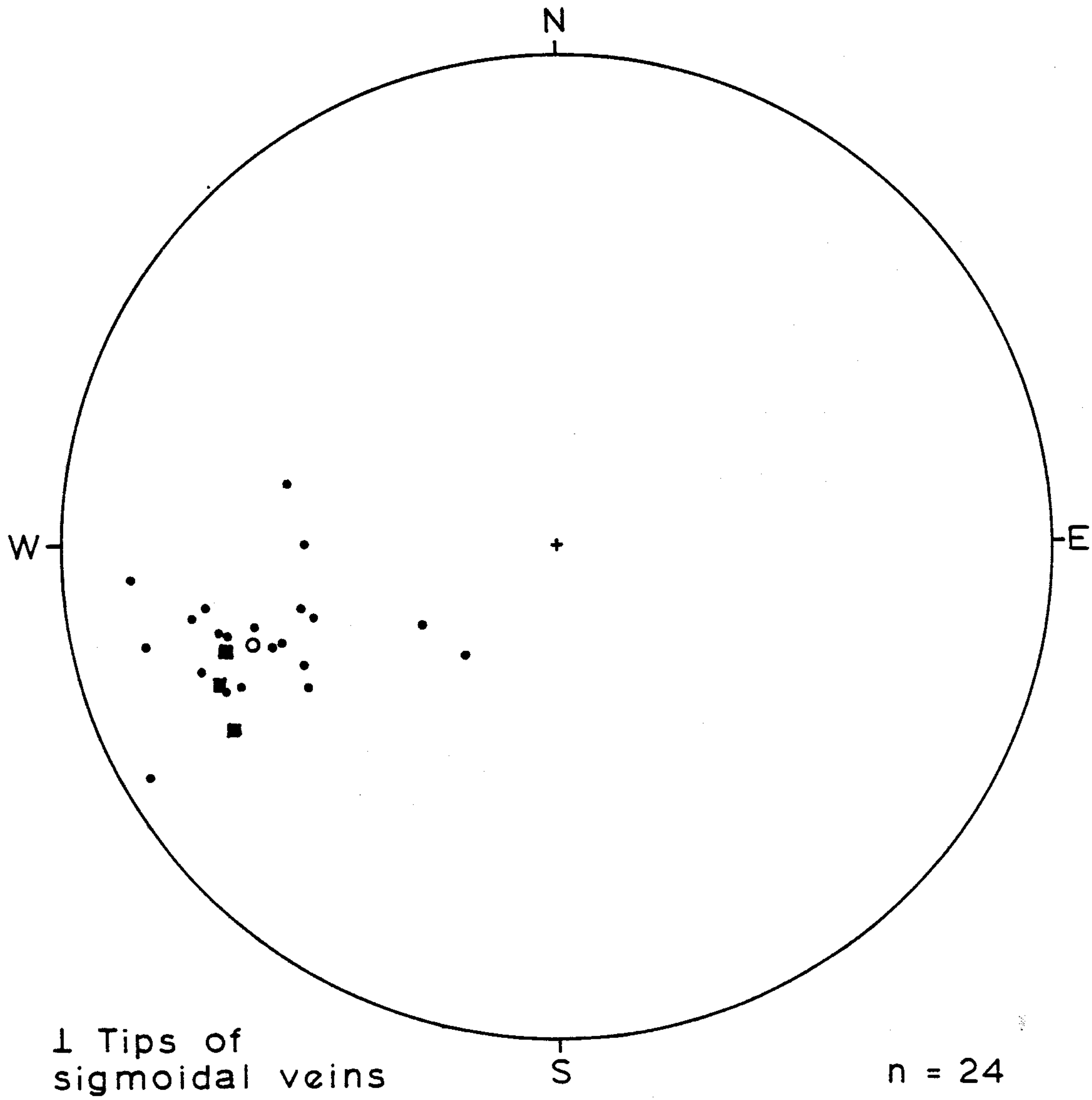


Figure 6.6. - Lower hemisphere equal area projection of 24 poles to tips of sigmoidal quartz veins. Open circle indicates estimated average orientation, used as an approximation of the σ 3 direction associated with the vein-forming event. Filled circles are vein tips from the northwest trending *en echelon* sets. Squares are vein tips from the north trending set



shear set is bisected by the maximum principal stress, σ 1. This concept was first used by Shainin (1950) to interpret conjugate sets of <u>en echelon</u> quartz veins. Since then, this interpretation has been used by many authors (Roering 1968; Beach 1975; Rickard and Rixon 1983). If the angle between the two shear zones is high (near 90 degrees) and the maximum extension direction within each individual shear zone is an ideal 45 degrees, then tension gashes in the two sets will open parallel to one and other and perpendicular to σ 1. Since the angle of intersection of the two shear zone orientations in my field area is so large (82 degrees), I consider the trend of the tips of the tension gashes to be the same as the trend of σ 1.

The three dimensional orientation of the conjugate shear set cannot be determined from its outcrop pattern. Both the σ_3 and the σ 2 directions must lie somewhere in the plane of the sigmoid tips. In order to determine the σ 2 and σ 1 directions within the plane, the hinge line orientations of the folds in the sigmoids was plotted (fig.6.7) These hinge lines should represent the σ 2 orientation (fig.6.8). A check to see if the hinge lines do give a reasonable estimate is provided if they plot on the plane perpendicular to σ 3. In this case the field measurements did plot on that plane, supporting the interpretation. The orientations of principal stresses determined for this conjugate shear set indicate that its outcrop expression is just a planar cut through a set whose two orientations are striking 185

Figure 6.7. - Lower hemisphere equal area projection of 11 hinge lines from sigmoidal quartz veins. The circle represents an estimated average orientation of the hinge lines. This average direction is used as an approximation of the σ 2 direction associated with the vein-forming event.

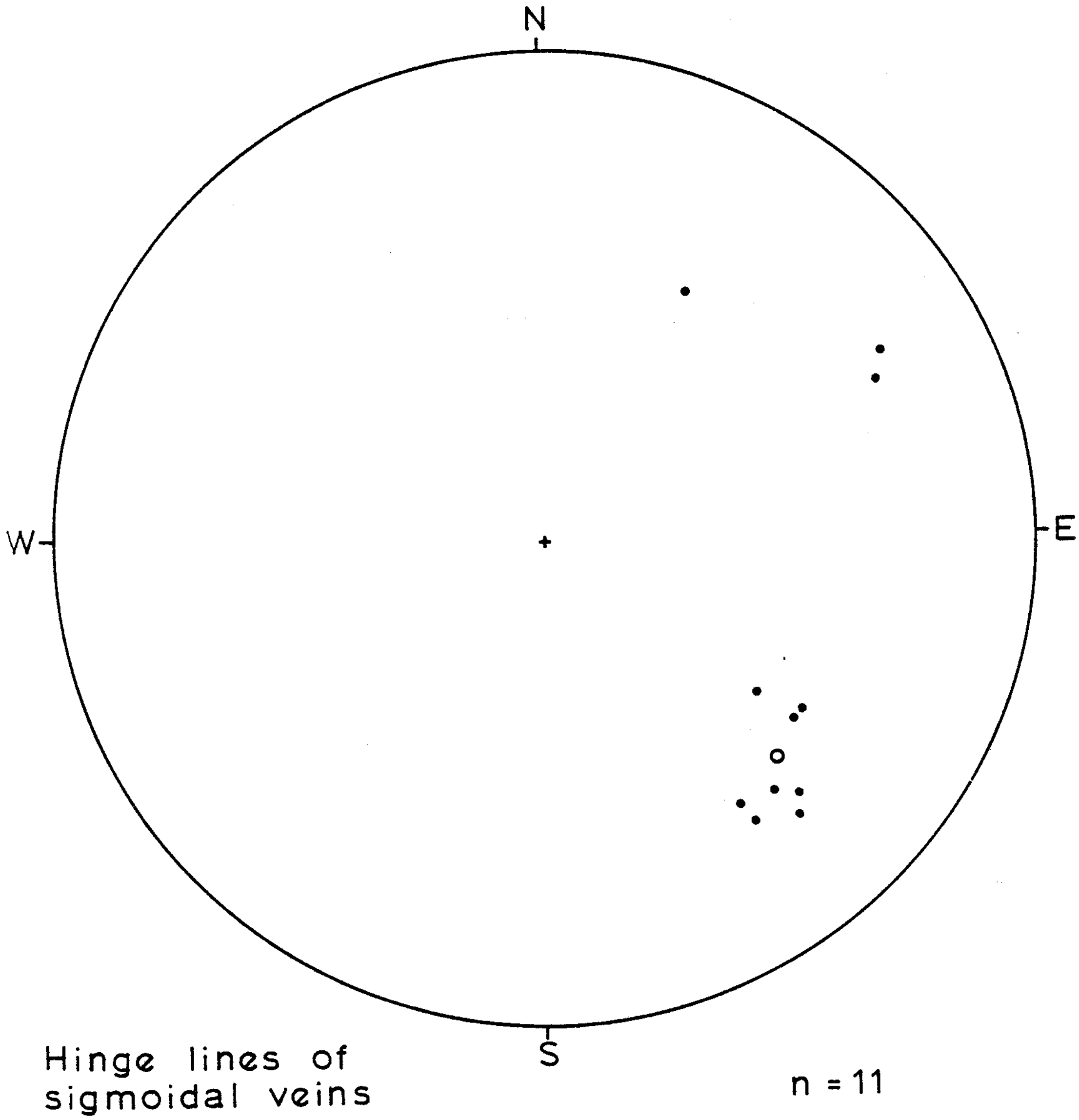
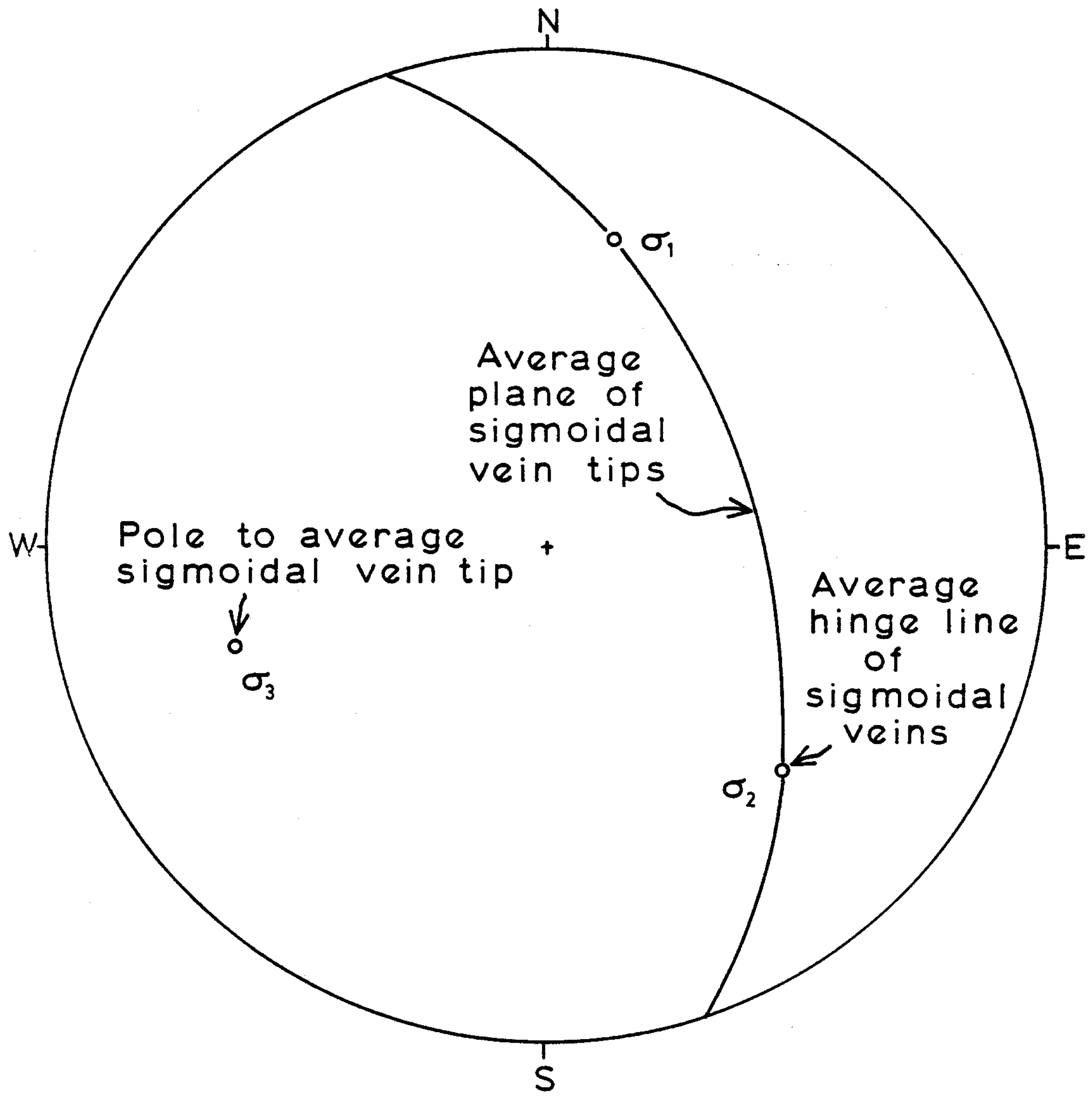


Figure 6.8. - Lower hemisphere equal area projection of the average orientations of σ 2 (from fig. 6.7) and σ 3 (from fig. 6.6), with σ 1 plotted 90 degrees away from both directions.

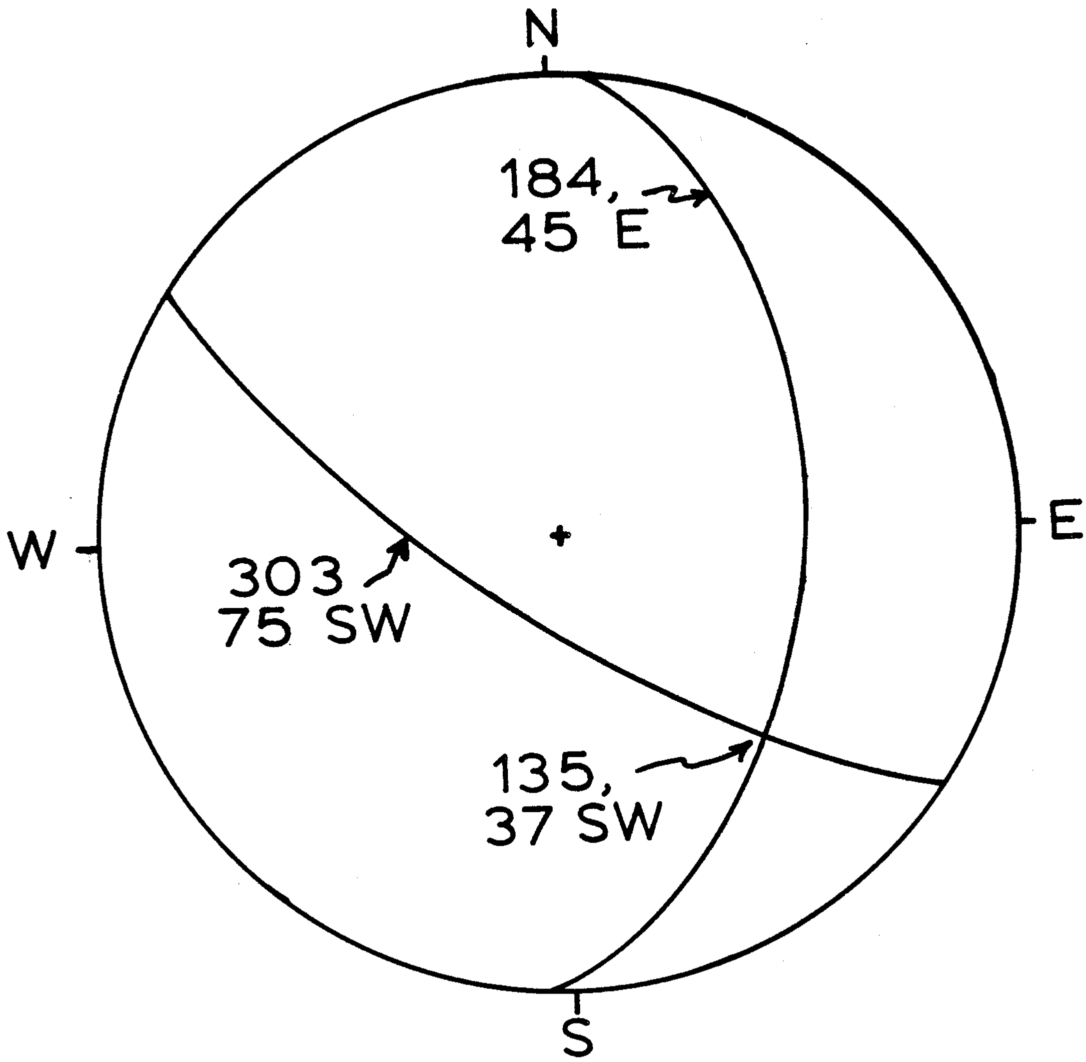


dipping 45E and striking 305 and dipping 75SW (fig.6.9).

Ramsay and Graham (1970) suggested that in areas where the incremental strain axes have remained constant throughout a deformation, the orientation of these conjugate shear sets can be correlated with axes of strain of the deformation. Under these conditions, the two directions of the conjugate shears will have a symmetric arrangement with respect to the principal finite bulk rock strains. The high symmetry of the conjugate shear pattern from the study area may indicate that the bulk strain was irrotational and that the orientations of λ 1, λ 2, and λ 3 can be determined from measurements of the conjugate shear set. If this indication is valid then the orientations of λ 1, λ 2, and λ 3 parallel the orientations of the principal stress axes.

<u>CORRELATION</u>. - Within the sandstone layers the approximate orientations of the principal stress and strain axes were determined. The elongate pebbles in the sandstone were used to determine the orientation of λ 1 and the orientation of a plane containing the λ 2 and λ 3 directions. The principal stress orientations were determined from the conjugate shear sets of quartz veins. The orientation diagrams for these two structures do not correlate exactly since λ 1 determined from the pebbles plunges more shallowly than the σ 3 determined from the veins. Although the orientations of stress and strain are not exactly superimposable, they are nearly so, suggesting that they are

Figure 6.9. - Lower hemisphere equal area projection of the average orientations of the two shear zones in the conjugate shear set.



both the products of a single nearly homogenous deformation event within the sandstone beds. In the siltstone beds the pole to the slaty cleavage is taken to represent the approximate λ 3 orientation. The maximum elongation direction (λ 1) was determined by plotting the orientation of a mineral lineation on that surface. λ 1 in the siltstones is coincident with λ 1 in the sandstones (from the pebbles), suggesting that the siltstone layers were deformed in a similar manner as, and during the same deformation event as the sandstones.

In the mudstones the slaty cleavage is well developed, but is very inconsistent in orientation and appears to have been disrupted during the formation of the quartz veins in the beds. These veins appear to develope synchronous with veins in the sandstone (see Chapter 5). There orientation is such that they appear to have opened at some moderate angle to the maximum compression direction σ 1, and maximum shortening direction λ 3, as determined from the sandstones and siltstones. This orientation of quartz veins may have resulted from the opening of veins along preexsisting planes of weakness (slaty cleavage planes). Although the veins are usually fibrous, the fiber orientations vary greatly within individual veins. This, combined with the branching and splaying of these veins, may indicate that these layers the veins opened in response to a large component of simple shear along the cleavage planes.

Kink bands mentioned in Chapter 3 have two trends in the outcrop (see fig.3.12). These trends with their consistant sense of displacement are not incompatable with the structures mentioned above. However, the nature of the kink bands is not clearly understood so I have made no attempt to determine principal stress or strain orientations from there orientations.

A block diagram of the structures mapped at Mispec Beach is given in plate 2. The structures are presented in their proper geographic orientation and in layers representing the lithologys in which they occur. To the left of the block are reproductions of the orientation diagrams for each of the structures as seen earlier in the text. The upper layer (A) is a layer of sandstone and pebbly sandstone with veins and deformed pebbles in it. The middle layer (B) is a layer of siltstone with a well developed slaty cleavage and kink bands (crenulation cleavage is not shown). The third layer (C) is a mudstone unit with quartz veins (black) and disrupted slaty cleavage.

One other structure which must be considered is the crenulation cleavage, which, as was described in Chapter 4, is well developed in the siltstones but not in the sandstones and postdates both the slaty cleavage and the veins. The strike of the crenulation cleavage is very nearly parallel with that of the slaty cleavage (fig. 6.10). There is a weakly developed mineral lineation of the slaty cleavages (fig.6.11). Asymmetric crenulation folds of the

Figure 6.10. - Lower hemisphere equal area projection of 49 poles to crenulation cleavage in siltstone.

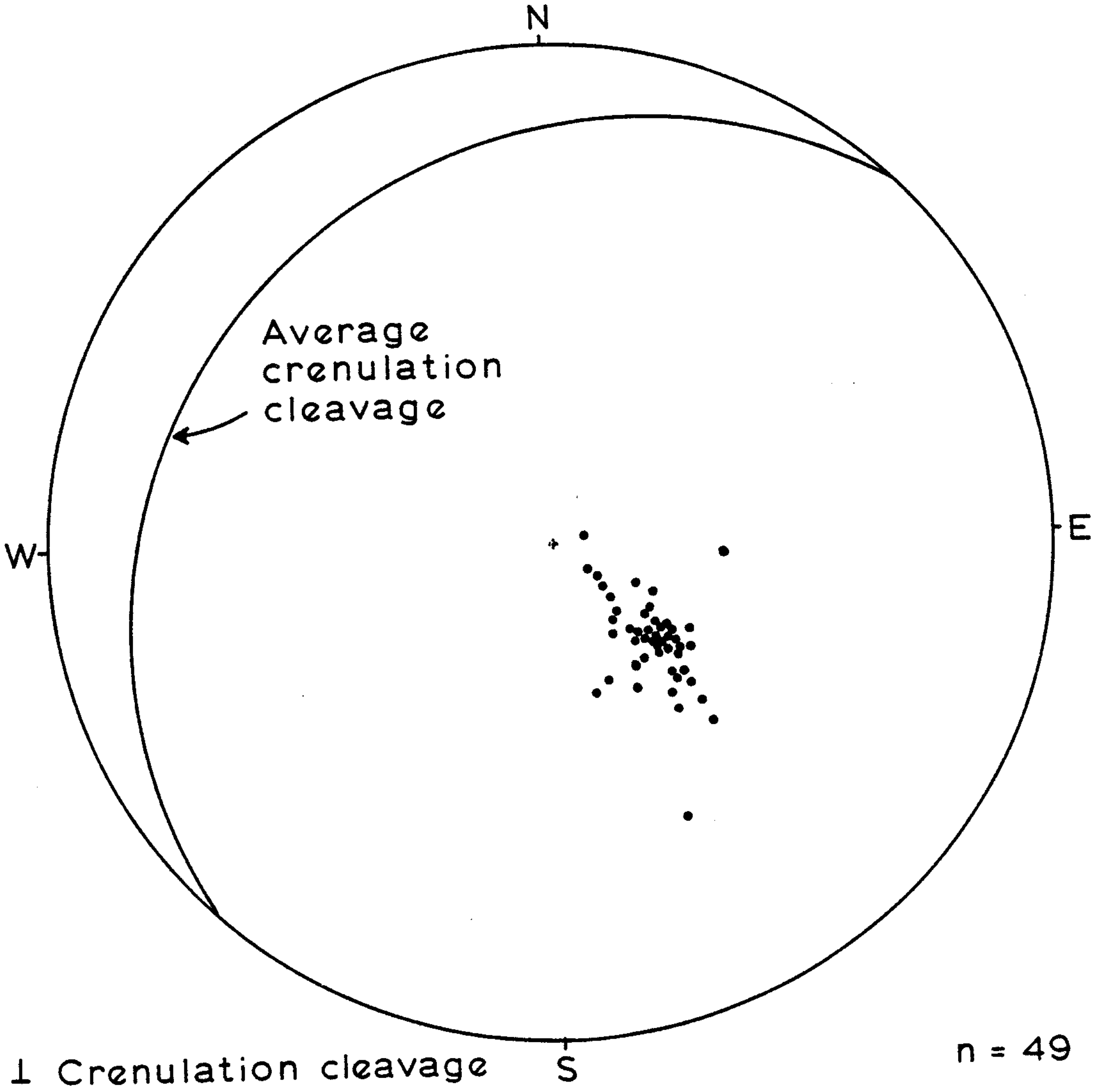
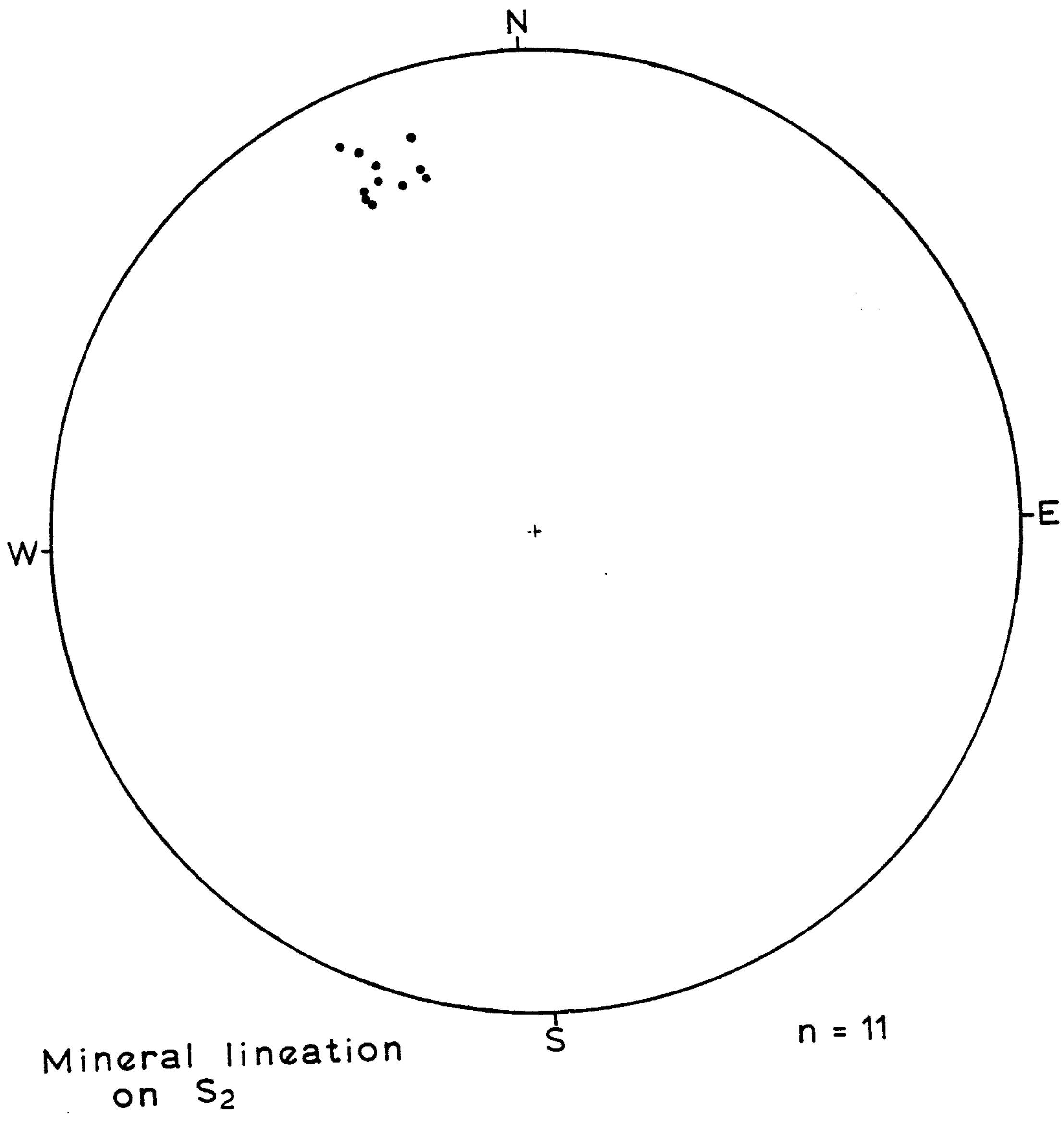


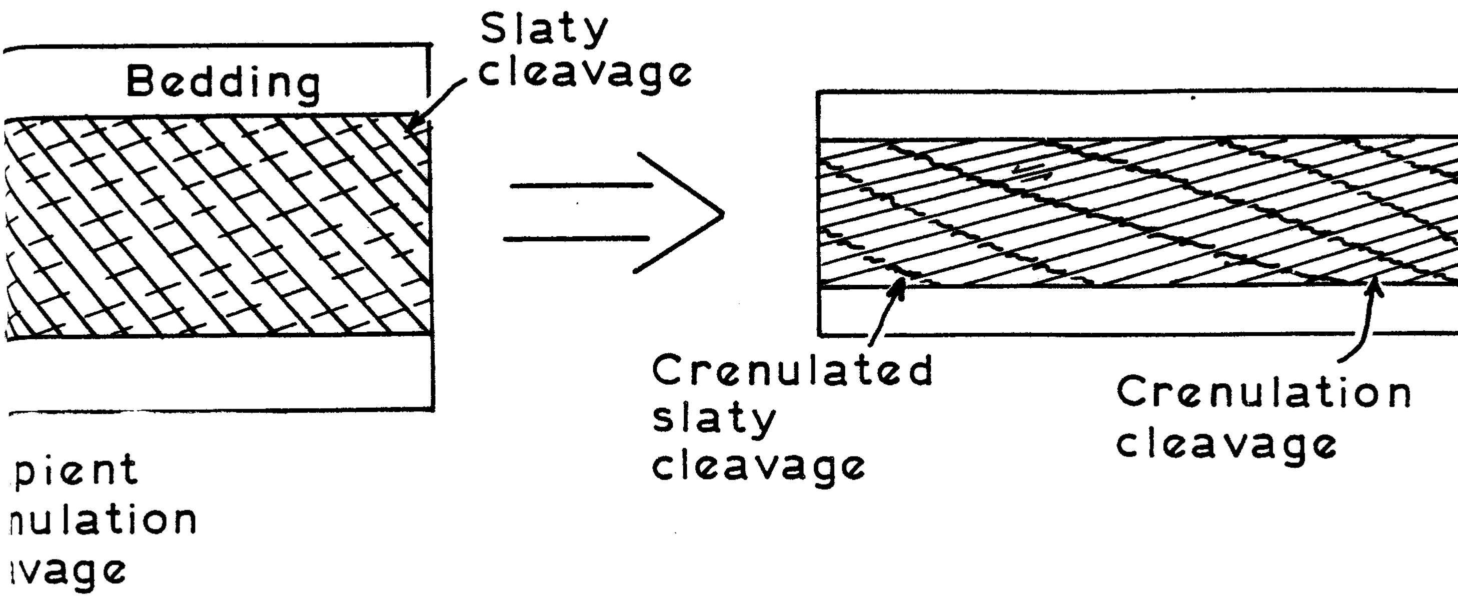
Figure 6.11. - Lower hemisphere equal area projection of 11 mineral lineations on crenulation cleavage planes.



slaty cleavage planes suggest some componant of normal type displacement along the crenulation cleavage planes. This sense of slip along the crenulation cleavage planes results in a shallowing of the dip of the slaty cleavage planes (fig.6.12).

By increasing the dip of the slaty cleavege planes (undeforming the rocks with respect to the crenulation cleavage, see above) a stronger correlation between the orientation of principal strains in the siltstone and the orientations of the principal stresses and strains in the sandstones is produced. By steepening the average plane given in figure 6.3 and comparing that with figure 6.8, one can see that the $\lambda 2$ and $\lambda 3$ directions from the slaty cleavage in the siltstone move closer to the $\lambda 2$ and $\lambda 1$ orientations determined for sandstones. Without any quantitative data on the amount of slip and the amount of pressure solution which occured during crenulation cleavage formation, I have no way of determining how much steeper the slaty cleavage planes might have been dipping.

Figure 6.12. - Deformation of a slaty cleavage plane dipping 60 degrees to the southeast (on right) by normal type slip along crenulaion cleavage planes dipping 20 degrees to the northwest. This deformation results in the shallowing of the dip of slaty cleavage.



CHAPTER 7

SUMMARY

The structures mapped at Mispec Beach include a penetrative slaty cleavage (S_1), quartz veins, deformed pebbles, and a crenulation cleavage (S_2). These structures correspond with structures mapped elsewhere in the Mispec Nappe. As mentioned in Chapter 2, Ruitenberg and others (1973) mapped a penetrative slaty cleavage (S_{1P}) that they associated with the local development of first generation folds (F_1). There are no F_1 folds at the Mispec Beach outcrop. Ruitenberg and others (1973) recognized a second generation of folds (F_2), which vary in magnitude from small crenulation folds of the slaty cleavage to larger folds several meters in amplitude. In the Mispec Beach area, these F_2 folds are crenulation folds of S_1, and they have an associated crenulation cleavage. A third generation of folds (F_3) affects Carboniferous rocks elsewhere in southern New Brunswick, but it has not been recognized in the Mispec Nappe in general (Ruitenberg et al., 1973)., and is not present at Mispec Beach.

The regional structures (large thrust nappes verging to the northwest) correspond well with the hypothesis that the Carboniferous strata were deposited in a strike slip basin that was closed as a result of its location in a restraining bend in the right lateral Chedabucto - Cobequid fault. Most of the structures mapped by previous workers (e.g. Rast and

Grant, 1973a, b) imply northwest transport of the thrust nappes.

My field work indicates a more northerly transport direction during deformation (see plate 2). However, the Mispec Beach area is very small compared to the whole zone of Carboniferous deformation, and the transport direction determined at this outcrop may not be of regional significance. It may only be a reflection of the outcrop within the nappe, or of the Mispec Nappe within the whole thrust pile. In order to understand the significance of the transport direction suggested by structures at Mispec Beach, detailed mapping elsewhere in the Mispec Nappe must be done to determine if observations at this one outcrop are consistent throughout the nappe, or if they vary in orientation at different locations.

Even a northward transport direction is not incompatible with the restraining bend hypothesis. The structures developed during Carboniferous deformation may have been strongly controlled by preexisting structures. This basement control of structures may also have contributed to the northward transport direction of the thrusts.

REFERENCES CITED

Alcock, F. J., 1938, Geology of St. John region: Geol. Surv. Canada, Mem. 216, 65 p.

Anderson, E. M., 1905, The dynamics of faulting: Edinburgh Geol. Soc., v. 3, part 3, p. 387-402.

Arthaud, F., and P. Matte, 1977, Late Paleozoic strike-slip faulting in southern Europe and northern Africa: a result of right-lateral shear zone between the Appalachians and the Urals: Geol. Soc. America Bull., v. 88, p. 1305-1320.

Bell, W. A., 1927,Outline of Carboniferous stratigraphy and geologic history of the Maritime Provinces of Canada: Roy. Soc. Canada, Proc. and Trans., 3rd Ser., v. 21, Issue Sect. 4, p. 75-108.

Borradaile, G. J.; M. B. Bayly; and C. M. Powell, eds., 1982, Atlas of deformational and metamrophic rock fabrics: Springer-Verlag, Berlin, 551 p.

Bradley, D. C., 1982, Subsidence in late Paleozoic basins in the Northern Appalachians: Tectonics, v. 1, p. 107-123.

_____, 1983, Tectonics of the Acadian Orogeny in New England and adjacent Canada: J. Geol., v. 91, p. 381-400.

_____, 1984, Late Paleozoic strike slip tectonics of the Northern Appalachians: Ph.D. Dissertation, State Univ. of New York at Albany, 285 p.

Currie, K. L.; R. D. Nance; G. E. Pajari; and R. K. Pickerill, , 1981, Some aspects of the pre-Carboniferous geology of St. John, New Brunswick: Geol. Surv. Can., Paper 81-1A, p. 23-30.

_____, and _____, 1983, A reconsideration of the Carboniferous rocks of Saint John, New Brunswick: Geol. Surv. Can., Paper 83-1A, p. 29-36.

Dostal, J., and D. F. Strong, 1983, Trace-element mobility during low-grade metamorphism and silicification of basaltic rocks from Saint John, New Brunswick: Can. J. Earth Sci., v. 20, p. 431-435.

Durney, D. W., and J. G. Ramsay, 191973, Incremental strains measured by syntectonic crystal growths: in K. A. DeJong and R. Scholten (eds.), Gravity and Tectonics, Wylie, New York, p. 67-96.

Eisbacher, G. H., 1969, Displacement and stress field along part of the Cobequid fault, Nova Scotia: Can. J. Earth Sci.,

v. 6, p. 1095-1104.

_____, 1970, Deformation mechanics of mylonitic rocks and fractured granites in the Cobequid Mountains, Nova Scotia: Geol. Soc. Amer. Bull., v. 81, p. 2009-2020.

Flinn, D., 1962, On folding during progressive three-dimensional progressive deformation: Quart. J. Geol. Soc. London, v. 118, p. 385-433.

Fyffe, L. R.; G. E. Pajari, Jr.; and M. E. Cherry, 1981, The Acadian plutonic rocks of New Brunswick: Maritime Sediments, v. 17, p. 23-36.

Garnett, J. A., and R. L. Brown, 1973, Fabric variation in the Lubec-Belleisle zone of southern New Brunswick: Can. J. Earth Sci., v. 10, p. 1591-1599.

Giles, P. S., and A. A. Ruitenberg, 1977, Stratigraphy, Paleogeography, and tectonic setting of the Coldbrook Group in the Caledonia Highlands of southern New Brunswick: Can. J. Earth Sci., v. 14, p. 1263-1275.

Greiner, H. R., 1962, Facies and sedimentary environments of the Albert shale: Amer. Assoc. Petrol. Geol. Bull., v. 46, p. 219-234.

Griggs, D., and J. Handin, 1960, Observations on fracture and a hypothesis of earthquakes: Geol. Soc. America, Mem. 79, p. 347-364.

Hayes, A. O., and B. G. Howell, 1937, Geology of St. John, New Brunswick: Geol. Soc. America, Spec. Paper 5, 146 p.

Hobbs, B. E.; W. D. Means; and P. F. Williams, 1976, An outline of structural geology: Wylie, New York, 571 p.

Howie, R. D., and M. S. Barss, 1975, Upper Paleozoic rocks of the Atlantic Provinces, Gulf of St. Lawrence, and adjacent continental shelf: Geol. Surv. Canada, Paper 74-30, p. 35-50.

Keppie, J. D., 1982, The Minas Geofracture: Geol. Assoc. Canada, Spec. Paper 24, p. 263-280.

Mann, W. P., 1984, Cenozoic tectonics of the northern Caribbean: structural and sedimentological studies in Jamaica and Hispaniola: Ph.D. Dissertation, State Univ. of N. Y. at Albany, 688 p.

McCutcheon, S. R., 1980, Revised stratigraphy of the Long Reach area, southern New Brunswick: evidence for major, northwestward-directed Acadian thrusting: Can. J. Earth Sci., v. 18, p. 646-656.

McKerrow, W. S., and A. M. Ziegler, 1971, The Lower Silurian paleogeography of New Brunswick and adjacent areas: J. Geol., v. 79, p. 635-646.

Mosher, S., 1983, Kinematic history of the Narragansett Basin, Massachusetts and Rhode Island: Constraints on late Paleozoic reconstructions: Tectonics, v. 2, p. 327-344.

O'Brien, B. H., 1977, The presistence and longevity of faults in the crustal evolution of southern New Brunswick: Maritime Sediments, v. 13, p. 93-106.

Olszewski, W. J., Jr., and H. E. Gaudette, 1982, Age of the Brookville Gneiss and associated rocks, southeastern New Brunswick: Can. J. Earth Sci., v. 19, p. 2158-2166.

Poole, W. H., 1976, Plate tectonic evolution of the Canadian Appalachian region: Geol. Survey Can., Paper 76-1B, p. 113-126.

Potter, R. R., J. B. Hamilton, and J. L. Davies, 1979, Geological Map of New Brunswick: New Bruns. Dept. Nat. Res.

Ramsay, J. G., 1967, Folding and fracturing of rocks: McGraw-Hill, New York, 568 p.

_____, 1980, The crack-seal mechanism of deformation: Nature, v. 2284, p. 135-139.

Rast, N., 1982, The Northern Appalachian traverses in the Maritimes of Canada: in N. Rast and F. M. Delany, eds., Profiles of orogenic belts, Amer. Geophys. Union, Geodynamics Ser. 10., p. 243-274.

_____, and K. L. Currie, 1976, On the position of the Variscan Front in southern New Brunswick and its relation to Precambrian basement: Can. J. Earth Sci., v. 13, p. 194-196.

_____, and W. L. Dickson, The Pocologan mylonite zone: Geol. Assoc. Can., Spec. Pap. 24, p. 249-261.

Rast, N., and R. J. Grant, 1973a, The Variscan Front in southern New Brunswick: N.E.I.G.C. Guidebook to Field Trips for 1973, p. 11.

_____, 1973b, The Variscan Front in southern New Brunswick: Am. J. Sci., v. 273, p. 572-579.

_____; _____; J. S. D. Parker; and H. C. Teng, 1978, The Carboniferous deformed rocks west of St. John, New Brunswick: Guidebook for field trips in southeastern Maine and southwestern New Brunswick, N.E.I.G.C., 70th Ann. Meeting, p. 162-173.

_____; M. J. Kennedy; and R. F. Blackwood, 1976, Comparison of some tectonostratigraphic zones in the Appalachians of Newfoundland and New Brunswick: Can. J. Eart Sci., v. 13, p. 868-875.

_____, and P. Stringer, 1974, Recent advances and the interpretation of geological structure of New Brunswick: Geoscience Canada, v. 2, part 4, p. 15-25.

_____, and _____, 1980, A geotraverse across a deformed Ordovician ophiolite and its Silurian cover, northern New brunswick, Canada: Tectonophys., v. 69, p. 221-245.

Rodgers, J., 1970, Tectonics of the Appalachians: Interscience, New York, 271 p.

Ruitenberg, A. A.; L. R. Fyffe; S. R. McCutcheon; C. J. St. Peter; R. R. Irrinki; and D. V. Venugopal, 1977, Evolution of pre-Carboniferous tectonostratigraphic zones in the New Brunswick Appalachians: Geoscience Canada, v. 4, p. 171-181.

_____, and S. R. McCutcheon, 1982, Acadian and Hercynian structural evolution of southern New Brunswick: Geol. Assoc. Canada, Spec. Paper 24, 131-148.

_____; D. V. Venugopal; and P. S. Giles, 1973, Fundy Cataclastic Zone: evidence for post-Acadian penetrative deformation: Geol. Soc. America Bull., v. 84, p. 3029-3044.

Shainin, V. E., 1950, Conjugate sets of en echelon tension fractures in the Athens Limestone at Riverton, Virginia: Geol. Soc. America Bull., v. 61, p. 500-517.

Stopes, M. C., 1914, The "Fern Ledges" Carboniferous flora of St. John, New Brunswick: Geol. Surv. Can., Mem. 41, 127 p.

Stringer, P., 1978, olding and cleavage in Triassic rocs of southern New Brunswick: Guidebook for field trips in southeastern Maine and southwestern New Brunswick, N.E.I.G.C. 70th Ann. Meeeting, p. 57-77.

_____, and E. Z. Lajtai, 1979, Cleavage in Triassic rocks of southern New Brunswick: Can. J. Earth Sci., v. 16, p. 2165-2180.

Strong, D. F.; W. L. Dickson; and R. K. Pickerill, 1979, Chemistry and prehnite-pumpellyite facies metamorphism of calc-alkaline Carboniferous volcanic rocks of southeastern New Brunswick: Can. J. Earth Sci., v. 16, p. 1071-1085.

Sylvester, A. G., and R. S. Smith, 1976, Tectonic transpression and basement-controlled deformation in San Andreas fault zone, Salton Trough, California: Amer. Assoc.

Petrol. Geol. Bull., v. 60, p. 2081-2102.

Webb, G. A., 1963, Occurrence and exploration significance of strike slip faults in southern New Brunswick: Amer. Assoc. Petrol. Geol. Bull., v. 47, p. 1904-1927.

_____, 1969, Paleozoic wrench faults in Canadian Appalachians: Amer. Assoc. Petrol. Geol., Mem. 12, p. 754-786.

White, J. C., 1983, Structural development along a section of the Cobequid fault, Nova Scotia (abs): J. Struct. Geol., v. 5, p. 553.

Williams, H., 1978, Tectonic lithofacies map of the Appalachian Orogen: St. Johns, Nfld., Memorial Univ. of Nfld., Map No. 1.

Wilson, J. T., 1962, Cabot fault, an Appalachian equivalent of the San Andreas and Great Glen fault systems and some implications for continental displacement: Nature, v. 195, p. 135-138.