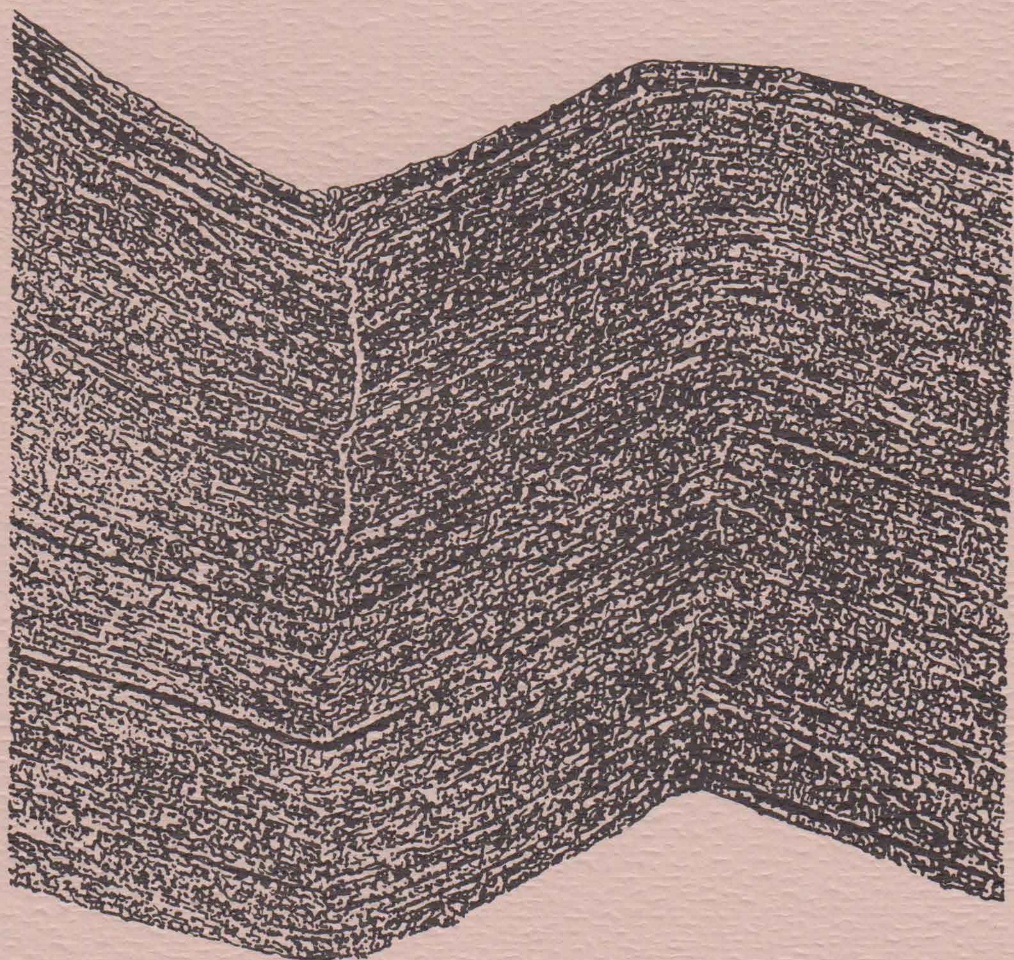
TRANSITION FROM DUCTILE TO BRITTLE DEFORMATION AT THE HEAD OF THE DEERFIELD BASIN, BERNARDSTON-LEYDEN AREA, MASSACHUSETTS

BY GERALD WILLIAMS



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TRANSITION FROM DUCTILE TO BRITTLE DEFORMATION AT THE HEAD OF THE DEERFIELD BASIN, BERNARDSTON - LEYDEN AREA, MASSACHUSETTS

A Thesis

by

GERALD WILLIAMS

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Contribution No. 35
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Major Subject: Geology

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ABSTRACT

Preserved Mesozoic sedimentary rocks of the Connecticut Valley - Deerfield Basin terminate northward in and near the Bernardston - Leyden area of Massachusetts and overlie a complex system of Paleozoic structures of the Connecticut Valley - Gaspe Synclinorium. This study examines the later stages of the structural history in this area in an attempt to place some limits on its stress orientation and deformational sequence. These later structural events span the period when the deformational style was changing from ductile to brittle. Detailed structural data were collected at 33 locations to determine the ductile sequence of deformation; 11 of these stations were analysed for transitional kink band structures; and a line of 67 additional fracture stations was established across the region from the Bronson Hill Anticlinorium to the Taconic Mountains. A total of 340 faults, 5830 joints, and 53 kink bands are analyzed.

Three periods of isoclinal folding are recognized and are probably a result of pre-nappe and nappe related deformation during the early stages of the Acadian orogeny. Two later periods of open folding are interpreted as related to gneiss dome emplacement toward the end of the Acadian orogeny. Two sets of kink bands bear

a conjugate relationship and indicate a σ_1 (maximum compressive stress) orientation of N 56°E, 32° NE at the time of their formation. These kinks probably are related to a late Paleozoic -early Mesozoic transition from ductile to brittle deformation and a reorientation of the principal stress trajectories involved in the formation of the Mesozoic Connecticut Valley Basin.

Faulting in and at the margins of the Mesozoic basin indicates a complicated combination of normal and strike-slip motions with possible pre-existing Paleozoic anisotropies controlling the configuration of the Paleozoic - Mesozoic boundary. Northeast, northwest, and east-west striking normal faults generally show components with south-side down relationships. Strike-slip components on northwest and northeast striking fault planes show right-lateral and left-lateral motion respectively, permissive of an interpretation of north-south or northeast-southwest compression.

The brittle fracture pattern of the detailed study area conforms to the regional pattern determined from 143 fracture stations along a traverse from the Bronson Hill Anticlinorium to the Taconic Mountains (including data collected by Williams (1976) and Silverman (1976) in the Bronson Hill area). The broader traverse area and the detailed study area are characterized by both normal faulting and roughly east-west jointing in the Paleozoic rocks and by both strike-slip and normal faulting with N 10°E trending joints in the Mesozoic rocks.

INTRODUCTION

The Problem

There has been a significant amount of petrofabric work done on the Paleozoic rocks in the vicinity of the study area. Recently there has also been much effort directed toward determining the brittle fabric of both the Paleozoic and Mesozoic rocks surrounding the area (Wise, personal communication). However, little or no work has been done on the transition from ductile to brittle behavior of these rocks or the orientations of the principal stresses operating at the time of this transition. The purpose of this study is to examine the stages of this transition and any possible changes in stress orientation by analysis of motion and stress indicators. Folds, dikes and veins, kink bands, and faults may give a good indication of the stress orientations involved during the transition. An appropriate place to study these features is in the vicinity of the Paleozoic-Mesozoic boundary. By examination of the nature of this unconformity a better understanding of the late tectonic evolution of the area can be reached.

A secondary purpose of this study is to define the brittle fracture pattern along the northwest border of Massachusetts. This pattern can be compared with the patterns found in other study areas in western Masachusetts to determine the domains of brittle fracture in the western part of the state and to establish where the more detailed study area fits into this regional pattern.

Location

The Bernardston-Leyden study area (Figure 1) is located in northwestern Massachusetts, just south of the Vermont state line and west of the Connecticut River Valley in the southern part of the Bernardston 7½' quadrangle. The area is about 20 kilometers south of Brattleboro, Vermont, 5 kilometers north of Greenfield, Massa-chusettts, and about 30 kilometers north-northwest of Amherst, Massachusetts. It lies in the axial region of the Connecticut Valley-Gaspe Synclinorium (Figure 2) at the northern end of the Mesozoic Deerfield Basin (Figure 1). A brittle fracture traverse, run in conjunction with this study, is located along the Massachusetts-Vermont border (Figure 3). It runs from the Taborton-Grafton area of New York state to the area of Northfield, Massachusetts-Hinsdale, New Hampshire.

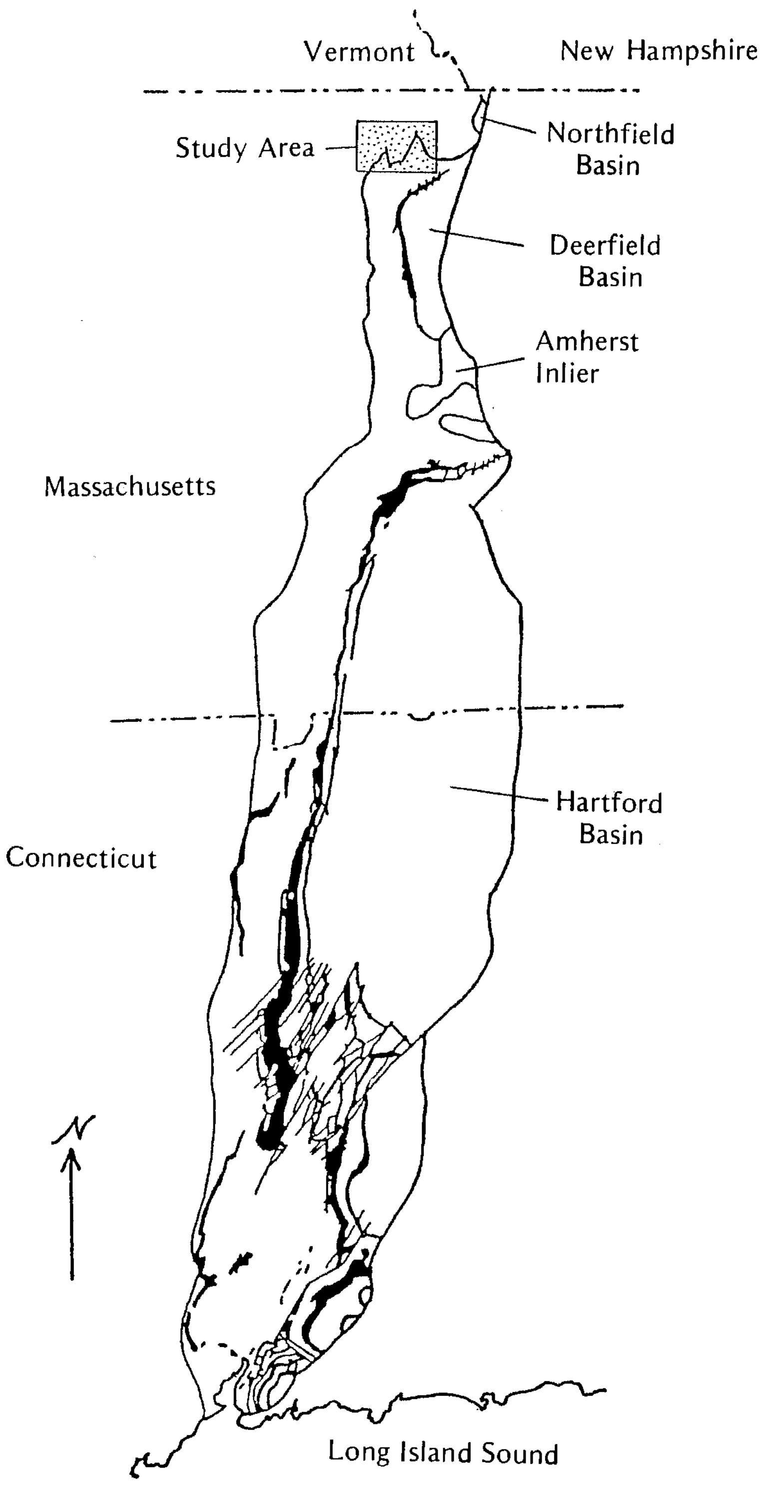

Figure 1: Connecticut Valley Mesozoic Basins.

Legend for Figure 2.

A - F Massifs and domes exposing Precambrian rocks of the Berkshire - Green Mountain Anticlinorium

A Green Mountain Massif	D Ray Pond Dome
B Chester Dome	E Sadawga Dome
C Athens Dome	F Berkshire Massif

1 - 6 Other domes of the Vermont line

1 Stratford Dome	4 Colrain Dome
2 Pomfret Dome	5 Shelburne Dome
3 Guilford Dome	6 Goshen Dome

a - k Domes of the Bronson Hill Anticlinorium

a Smarts Mountain Dome	g Keene Dome
b Mascoma Dome	h Vernon Dome
c Lebanon Dome	i Warwick Dome
d Croydon Dome	j Pelham Dome
e Unity Dome	k Monson Gneiss
f Alsted Dome	

BQ - Bernardston Quadrangle

Dotted line separates Connecticut Valley - Gaspe Synclinorium from Berkshire - Green Mountain Anticlinorium.

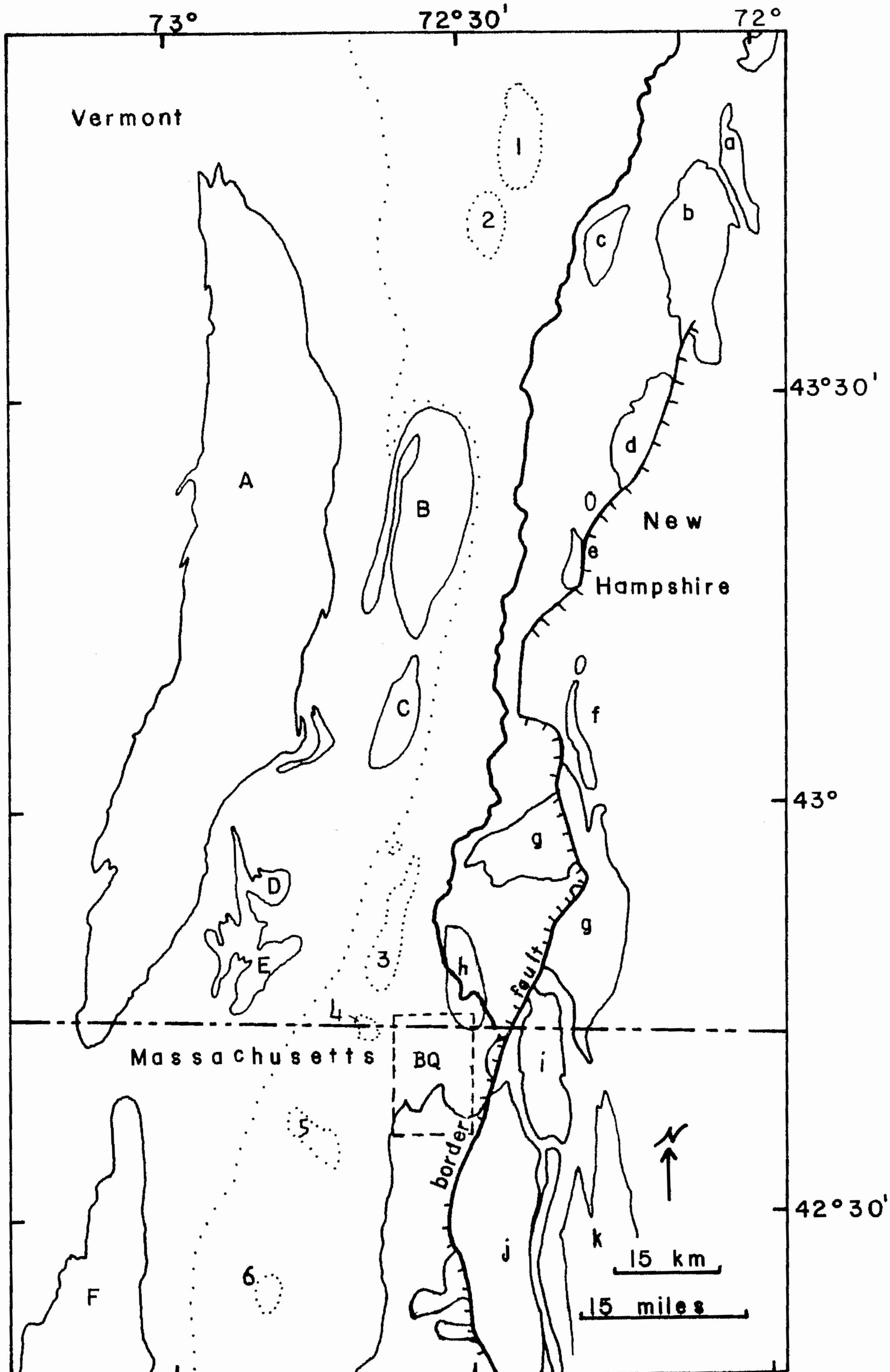

Figure 2: Regional Location Map.

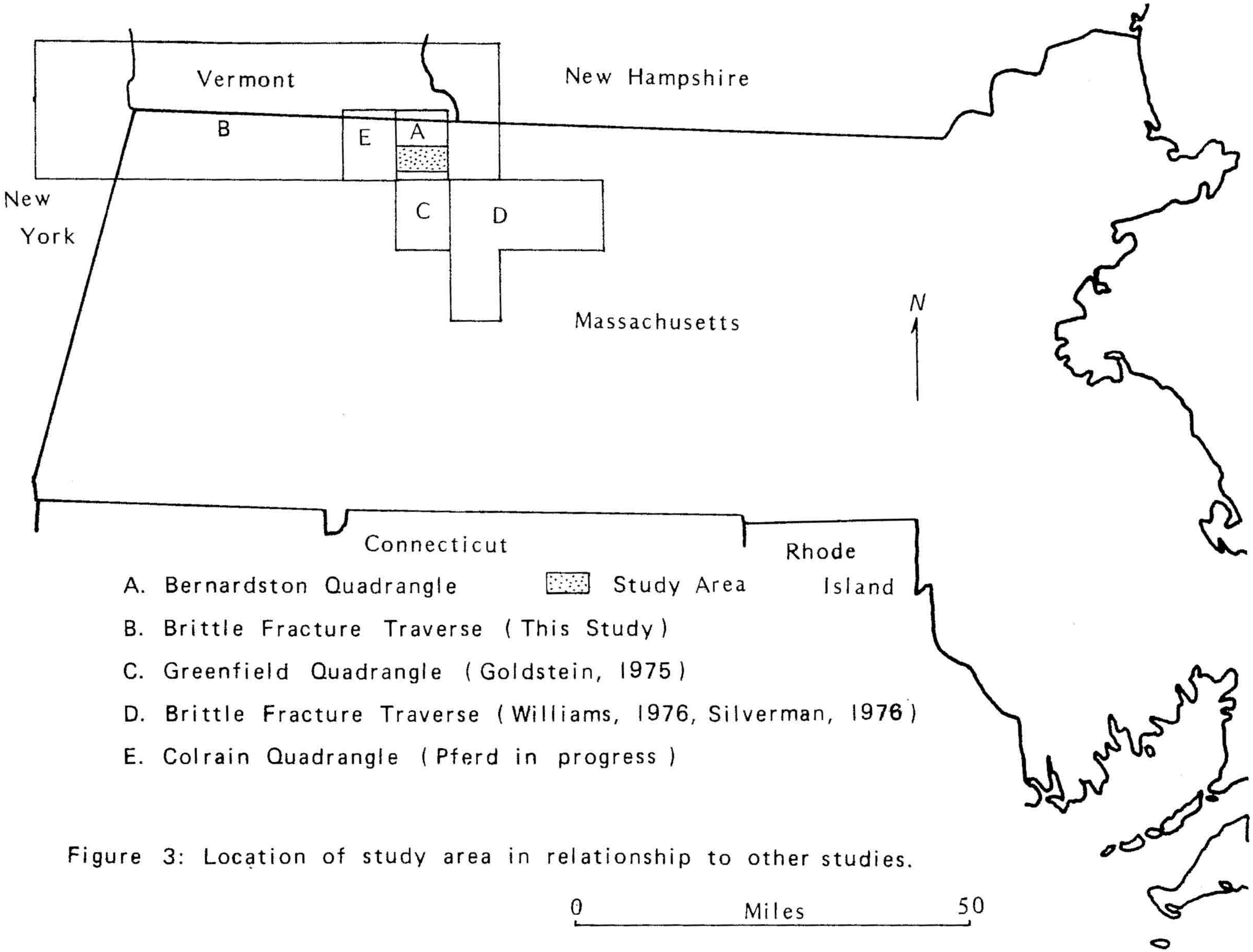

Figure 3: Location of study area in relationship to other studies.

Physiography and Culture

Topographically the study area can be divided into two regions. The northern highland region is underlain by low- to middle-grade metamorphosed Paleozoic rocks of the Lower Devonian Littleton Formation, Silurian Clough and Fitch Formations, and the Ordovician Partridge Formation. This region is characterized by rugged topography with steep east- and west-facing slopes and deeply incised stream valleys. The southern lowland region is underlain by the Mesozoic rocks of the late Triassic Sugarloaf Formation and is characterized by subdued topography and meandering stream valleys. Much of the topography over the area has been reshaped by Pleistocene glaciation, leaving the highlands streamlined in the direction of ice movement and the lowlands characterized by ground moraine, drumlins, eskers, kettle holes, and glacial lake deposits. Recent stream action has further modified the topography to the present configuration. Relief in the area is about 350 meters (1130 feet) from the top of Frizzell Hill (Figure 4) at over 400 meters (1310 feet) to the Fall River Valley in the south (Figure 4) at less than 55 meters (180 feet). The area includes the divide between the Connecticut and upper Deerfield drainage basins with the Green and Fall Rivers being the major streams of the area. Most streams flow from north to south with deep valleys in the highlands and meandering flood plains in the lowlands.

Explanation for Figure 4.

Jurassic	Jtc	Mount Toby Conglomerate
	Jdb	Deerfield Basalt
Triassic	TRs - Js	Sugarloaf Formation
Devonian	Dwr	Waits River Formation
	Dgm	Gile Mountain Formation
	Dsp	Standing Pond Volcanics
	Dpv	Putney Volcanics
	Dle - Dlw	Littleton Formation (eastern, western)
Silurian	Sf	Fitch Formation
	Sc	Clough Quartzite
Ordovician	Ops - Opf	Partridge Formation
	Oam	Ammonoosuc Volcanics
Probable Ordovician	Opa	Pauchaug Gneiss

A Frizzell Hill (high point)

B Fall River (low point)

C Otter Pond

D Old Fossil Locality

E New Fossil Locality

Study Area Boundary

Metamorphic Isograds

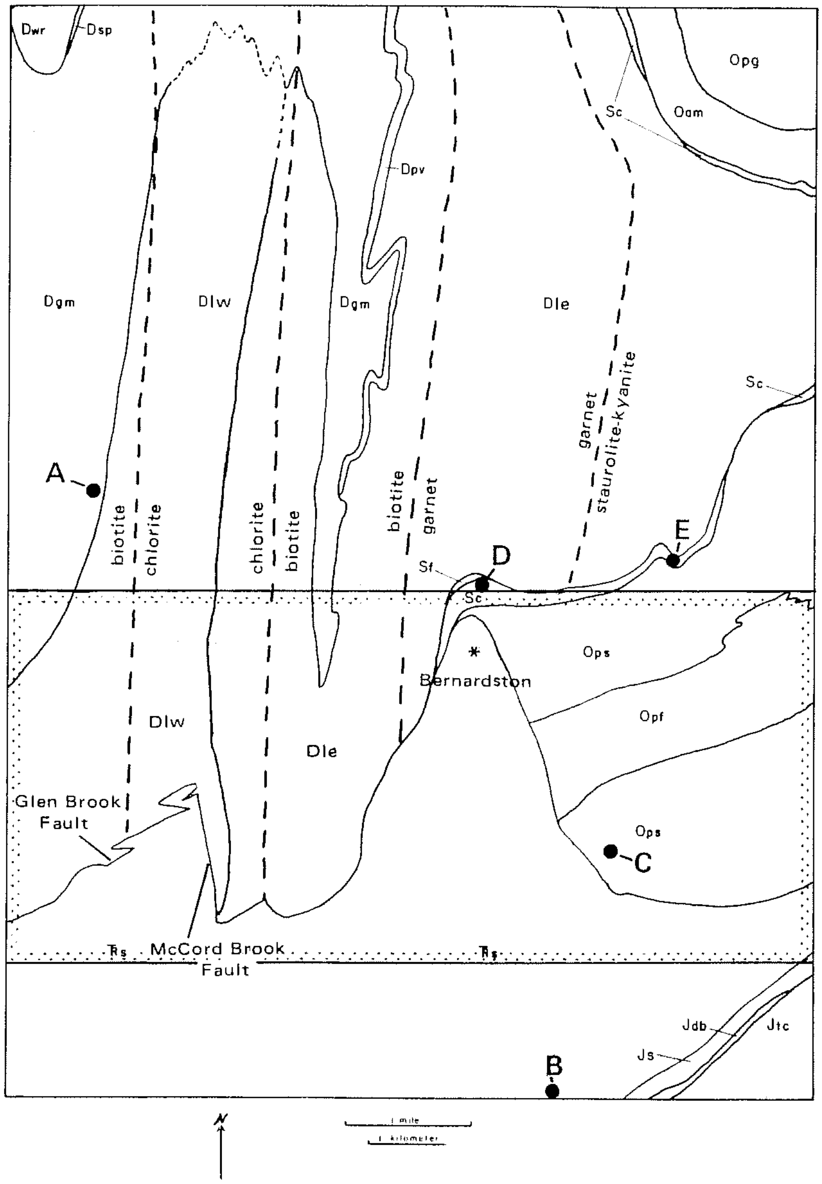

Figure 4: Geologic index map of the Bernardston quadrangle.

The area is sparsely populated with agriculture the major industry, although this has been declining in recent years. Public roads as well as numerous farm roads and logging trails provide relatively easy access to most of the area. Stream valleys as well as east- and west-facing slopes generally provide the best outcrop in the highlands, whereas most of the lowland outcrops are in stream valleys.

Previous Work

The bedrock of the Bernardston-Leyden study area was mapped in reconnaissance by Emerson (1898) as part of the geology of Old Hampshire County and as a part of the geologic map of Massachusetts in 1917. C. H. Hitchcock worked with the stratigraphy associated primarily with the fossil-locality at Bernardston (Figure 4) for correlation with the Stratford quadrangle in Vermont in 1912. The bedrock was later mapped by Balk (1956) but correlations with some of the better known units were not made at that time. Recently Boucot, et al. (1958), Trask (1964), and Hepburn, Hatch, Stanley, and Robinson have made the regional correlations and better defined the Paleozoic contacts. Hepburn (1975) has mapped the Brattleboro quadrangle in southeastern Vermont where he finds a sequence of folding similar to the sequence found in the Bernardston-Leyden area covered by this study.

Brittle fracture studies (Figure 3) in the Deerfield Basin (Goldstein, 1975) and northern Pelham Dome area (Williams, 1976, Silverman, 1976) as well as structural studies in the northern Pelham Dome area (Laird, 1974, Onasch, 1973, Robinson, 1963, 1967, 1976, Ashenden, 1973, Northeast Utilities, 1975) contribute to the understanding of the structural pattern of this region.

Stratigraphy

Formation names in the Bernardson quadrangle have been updated from those of Balk (1956), Emerson (1898), Hitchcock (1912), and others by Trask (1964), Trask and Thompson (1967), Hepburn (1972), and Stanley and Robinson (unpublished data) during the recent compilation of the new geologic map of Massachusetts (in progress). These updated names will be used throughout this paper.

Ordovician Partridge Formation. (Bernardston Formation of Balk, 1956, Hitchcock, 1912, Emerson, 1898, first defined as Partridge Formation in this area by Trask, 1964) This is the lowermost formation of the study area and is exposed on the eastern side of the Mesozoic reentrant underlying the village of Bernardston (Figure 4). It is composed of interbedded phyllites and amphibolites with lenses of metamorphosed rhyolite and diabase in some of the southern exposures (Balk, 1956). The thickness of this unit in the study area is on the order of one kilometer, although discontinuous exposure and folding prevent an accurate estimate.

Silurian Clough and Fitch Formations. (Formerly included in the Bernardston Formation by Balk, 1956) These are thin quartzite and quartz-conglomerate units with some fossiliferous calcareous quartzite and limestone layers (Boucot et al., 1958). The age of the Clough Formation is well documented by the occurrence of fossils of lower Silurian age 1.5 kilometers north of the intersection of Routes 5 and 10 in Bernardston (4728500 meters north, 700580 meters east, zone 18 of the Universal Transervse Mercator Grid System) and have also been reported (Beard, personal communication) at about 0.6 kilometers southeast of the intersection of Huckle Hill and Purple Meadows Roads (4728675 meters north, 703200 meters east, zone 18 of the Universal Transverse Mercator Grid System) (Figure 4). Estimates of the thickness of these units vary from about 150 meters (Emerson, 1898) to about 20 meters (Balk, 1956) but the thickness is probably closer to the lower end. These units extend only a short distance into the study area from the northeast and were not sampled.

Lower Devonian Gile Mountain Formation. (Formerly called the Conway Formation, Balk, 1956, Hitchcock, 1912, and Emerson, 1898) This formation is a quartz, muscovite, biotite schist with some inter-bedded quartzite and limestone layers (Balk, 1956). The stratigraphic relationship of this unit with the Littleton Formation is questionable. Near the contact between the two formations at Frizzell Hill graded beds in the Gile Mountain formation suggest that it underlies the Littleton, whereas in the east the Littleton Formation is in contact with the Silurian Fitch and Clough Formations suggesting the reverse relationship. These inconsistencies imply a relationship complicated by isoclinal folding or undetected faulting.

Lower Devonian Littleton Formation. (Formerly called Leyden Phyllite (Hitchcock, 1912), Leyden Argillite (Emerson, 1898, Balk, 1956), and Guilford Slate) Balk (1956) separated the Leyden into an upper and lower unit, although his stratigraphy is likely reversed from that interpreted for this study. This distinction is not made in the recent compilation of the state geologic map however, both units are described in this study due to their different mechanical behavior which will be discussed later. The western (lower of Balk, 1956) unit of this formation is composed of interbedded black, fine-grained, argillaceous, phyllitic layers with gray, fine-grained, quartzite layers at intervals of a few centimeters. The eastern (upper of Balk, 1956) unit of this formation contains the same phyllitic layers but is characterized by the lack of the numerous quartzite layers present in the western unit, although a few thin quartzite layers are present. Both sections are intruded locally by small granite dikes and/or sills which have been highly altered. Quartz veins of several generations are evident throughout the formation. Rocks of both units of this formation weather characteristically dark gray to silvery gray.

The thickness of the Littleton Formation in this area is uncertain. The intense folding and the location in the nose area of the Skitchewaug Nappe has probably created an apparent thickness much greater than the actual thickness. A thickness of at least one kilometer is required if the unit is doubled into one overturned limb and a maximum thickness of about three kilometers is possible in the unlikely event that the apparent thickness is close to the true thickness.

Triassic Sugarloaf Formation. This is the lowermost formation in the Deerfield Basin and the only Mesozoic unit in this study area. It is exposed at several contacts with Paleozoic rocks and locally within the southern part of the study area. At some of these contact locations the arkose which is red in general shows signs of leaching or bleaching along the contact. In the small basin just north of the Glen Brook fault contact (Figure 4) the generally red basal beds have irregular streaks of buff colored rock anastomosing through them. At the contact just south of Otter Pond (Figure 4) about 30 meters of stratigraphy at the base of the exposure is beige or buff color in contrast to the typical red color. The next 30 meters show mixtures of both the typical red and the buff colors. Above this level, only red beds are present.

Paleocurrent directions in the Sugarloaf Formation (Stevens, 1977) indicate that the sediments were transported into the basin from the north in the northern and western portions of the basin and from the northeast along the eastern margin.

The Sugarloaf Arkose dips generally to the south and southeast except along the eastern side of the Bernardston reentrant where it dips to the southwest. The dips vary from about 10° to about 50° and average about 25° to 30°. If the dip of the unconformity is reflected by the dip of this unit the thickness of the Sugarloaf Arkose at its contact with the Deerfield Basalt (Figure 4) in the southeastern corner of the Bernardston quadrangle is in the range of 0.8 kilometers if the basin resulted from several faults, or up to 2.5 kilometers if controlled by one master fault.

Jurassic Sugarloaf Formation, Deerfield Basalt, Turners Falls Sandstone, and Mt. Toby Conglomerate. These lavas and fresh water sediments comprise the remainder of the formations of the Deerfield Basin but are not present in the study area.

Paleozoic Setting and Structure

The Paleozoic section of the study area is located in the Connecticut Valley low-grade zone between the eastern Vermont and

Table 1A: Pertinent times and events of the region.
(ages in millions of years B. P.)

Era	Period	Age (boundary)	Scale	Tectonic Activity	Metamorphic and Igneous Activity	Erosion and Deposition
Cenozoic		65				
Mesozoic	Cretaceous		100		White Mt. Magmas	Regional Erosion
		136				
	Jurassic				Basalt Flows; Oldest Atlantic Seafloor; Diabase Dikes	Connecticut Valley Sedimentation
		190		Taphrogenesis Rifting Begins		
	Triassic		200			
		225		Probable Kinking		
Paleozoic	Permian				Allegheninan Thermal Resetting	
		280				
	Carboniferous		300			
		345				Erosion
	Devonian			Open Folds; Domes; Nappes; Isoclinal Folds (Acadian Orogeny)	New Hampshire Magma Series; Metamorphism	Marine Sedimentation
		405	400			
	Silurian					Intermittent Erosion and Sedimentation
		430				
	Ordovician			Taconic Orogeny	Ordovician Granites; Regional Metamorphism West of Berkshires	
		500	500			
	Cambrian					Marine Sedimentation
		570			Late PЄ of Pelham Dome	Erosion to Granville Basement

Table 1B: Time Relationships

Method	Age in M.Y.	Geologic, Structural Events and or Lithologic Units	Reference
K-Ar	115 ± 15	Lamprophyre dikes, younger White Mt. Magma Series, New England	McHone (1978)
	Jurassic	Differential subsidence and tilting of basin and gradual warping of Holyoke Range to present configuration	Chandler (1978)
K-Ar	140 - 180	Major age grouping from gouge and mylonite in normal faults	N.E. Utilities (1975)
Stratigraphy Petrology	Lower Jurassic	First erosional unroofing of Belchertown Intrusive Complex	Hall (1973)
K-Ar	143 - 182	Mylonite, gouge at fault contact Erving and Upper Gile Mt. Fm. (Mt. Toby site)	N.E. Utilities (1975)
K-Ar	154 ± 6	Fault in Putney Volcanics (Gill site)	N.E. Utilities (1975)
K-Ar	177 - 186	Sericitized feldspar in Gile Mt. Fm. (Mineral Hill site)	N.E. Utilities (1975)
K-Ar	189 ± 9	Deerfield Basalt (Poet's Seat site)	N.E. Utilities (1975)
K-Ar	171 - 196	All Connecticut Valley Basalts (Holyoke, Talcott, Hampden, Deerfield)	Reeseman et. al. (1973)
K-Ar	187 ± 20	Diabase dikes, New England, older White Mt. Magma Series	McHone (1978)
K-Ar	186 - 190	Major age grouping Deerfield Basalt, sericitized feldspar	N.E. Utilities (1975)
	Lower Jurassic	Initial formation of Amherst Block (isolation of Hartford and Deerfield Basins)	Chandler (1978)
		Spores, pollen, and fishes in Portland and Shuttle Meadow Fms. of Hartford Basin	Cornet et. al. (1973)
		Formation of Hartford and Deerfield Basins (?) spores and pollen	Cornet (1975)
Stratigraphy Petrology	Upper Triassic 210	Erosion to level of nappes and high grade metamorphics of Bronson Hill Beginning of Connecticut Valley Sedimentation	Krynine (1950)
K-Ar	211 ± 8	Ultramylonite at contact Erving - Lower Gile Mt. Fm. (Mt. Toby site)	N.E. Utilities (1975)
K-Ar	206 - 233	Mylonite in Gile Mt. Fm. (Mineral Hill site)	N.E. Utilities (1975)
K-Ar	200 - 235	Major age grouping of mylonites, breccias, and cataclaized gneisses in Gile Mt. Fm.	N.E. Utilities (1975)
K-Ar	230 ± 15	Oldest White Mt. Plutonic activity	McHone (1978)
K-Ar	227 - 242	Slickensides in Gile Mt. Fm. (Mineral Hill site)	N.E. Utilities (1975)
K-Ar	259	Pseudotachylite in Ramapo zone, New York	Ratcliffe (1977)
K-Ar	284 ± 10	Shear zone in Standing Pond Volcanics (Gill site)	N.E. Utilities (1975)
Rb-Sr	340 - 344	Barre Pluton, Vermont	Naylor (1971)
Rb-Sr whole rock	349 ± 13	Standing Pond Volcanics (metamorphic date ?) (Gill site)	N.E. Utilities (1975)
K-Ar	349	Adamant Granite, Vermont	Faul (1963)
Rb-Sr	354 - 357	Echo Pond Pluton, Vermont	Naylor (1971)
Rb-Sr whole rock	359 ± 11	Concord Granite, last of New Hampshire Magma Series	Lyons et. al. (1977)
Rb-Sr	363 - 380	Derby Pluton, Vermont	Naylor (1971)
Rb-Sr	380 ± 3	Barre type plutons (above)	Naylor (1971)
U-Pb	382 ± 5	Belchertown Intrusive Complex	Ashwal et. al. (in press)
Rb-Sr whole rock	380	Gile Mt. Fm. (metamorphic date ?)	N.E. Utilities (1975)
Rb-Sr	385 ± 20	Prescott Intrusive Complex (post-nappe formation, pre-doming)	Naylor (1971)
	Lower Devonian	Cardigan Pluton and Hardwick Granite (folded with nappes)	Robinson (1967)
Rb-Sr whole rock	402 ± 5	Blackwater Pluton of Spaulding Quartz Diorite	Lyons et. al. (1977)
Rb-Sr whole rock	405 ± 78	Bethleham Gneiss	Lyons et. al. (1977)
Rb-Sr whole rock	411 ± 19	Cardigan Pluton of Kinsman Quartz Monzonite	Lyons et. al. (1977)
Stratigraphy	Lower Devonian	Waites River Fm.	
		Standing Pond Volcanics - Putney Volcanics	
		Gile Mt. Fm.	
		Littleton Fm.	Boucot & Arndt (1960)
	Silurian	Fitch Fm.	Boucot (1961)
Fossils		Clough Fm. (brachiopods and corals)	Boucot (1961)
Stratigraphy	Ordovician	Partridge Fm.	Robinson (1967)
		Ammonoosuc Volcanics	
		Pauchaug Gneiss	Naylor (1967)
	Upper Precambrian	Monson Gneiss - Fourmile Gneiss	Robinson (1967)
		Poplar Mt. - Dry Hill Gneiss	Naylor et. al. (1973)

western New Hampshire metamorphic highs (Figure 4). Metamorphic grade ranges from chlorite zone in the center of the study area to biotite grade on the west side and staurolite-kyanite grade on the east side (Thompson et al., 1968). The rocks in the eastern part of the area belong to the Skitchewaug Nappe, the middle of three Acadian age structures of east over west transport. The Acadian orogeny took place during the closing of the proto-Atlantic (Iapetus Sea) as "Baltica" collided with paleo-North America in the Devonian (Ziegler, et al., 1977) (Table 1A). Later backfolding of the nappes (Robinson, 1967) west over east, was followed by the emplacement of the gneiss domes of the Bronson Hill Anticlinorium (Table 1A). These domes were emplaced during the peak of metamorphism (Trask, 1964)toward the waning stages of the Acadian Orogeny later in the Devonian.

To the west of the Connecticut Valley-Gaspe Synclinorium is the Berkshire Anticlinorium, the basement core of the western part of the northern Appalachians. The synclinorium is modified on the western side of its axis by a line of gneiss domes extending from Connecticut through Massachusetts into Vermont. The Guilford and Vernon Domes of southeastern Vermont lie just north of the study area and the Colrain and Shelburne Falls Dome lie to the west and southwest of the study area (Figure 2).

Structurally the Paleozoic rocks of the study area are quite complex. In the low grade part of the Littleton Formation at least seven fold sets have been recognized in the present study. These have contorted the bedding and schistosity into remarkable small-scale features. On a large scale the general north-south trend of the unit is maintained throughout the study area although, the location with respect to the Skitchewaug Nappe and the multiphase folding locally complicate this trend. The Partridge Formation is generally garnet grade or above and much more massive than the Littleton, preventing or obscuring development of the many complex small-scale folds, although it has a well developed schistosity and some small-scale folding.

Geophysically the study area lies on the eastern flank of the main gravity high associated with the Berkshire-Green Mountain Anticlinorium and to the east of the great gravity gradient of the Appalachians (Diment, 1968, Bromery, 1967). The Connecticut Valley-Gaspe Synclinorium lies in a broad magnetic low between the highs of the Berkshires to the west and the domes of the Bronson Hill Anticlinorium to the east. Domes within the synclinorium disrupt this broad low as local highs (Harwood and Zietz, 1977, Andreasen and Zandle, 1963).

Mesozoic Setting and Structure

The Mesozoic section of this study area is located in the Deerfield Basin (Figure 1), near the present northern limit of Mesozoic rocks in the Connecticut Valley Basin. The Deerfield Basin is separated from the Hartford Basin to the south by the Amherst Inlier (Figure 1). The basin is bounded on the east side by a major border fault and on the north and west by a complexly flexed and faulted unconformity. The strata near the western boundary may have originally extended to the Newark Mesozoic Basin in New Jersey, where the major border fault is located on the western side of the basin. This broad terrane hypothesis was first proposed by Russell (1878, p. 230) and has been summarized most recently by Hubert et al., (1978).

The development of the basin began in late Triassic time when Africa began to separate from North America in the development of the present Atlantic Ocean (Table 1A). During this initial "pull apart" the Connecticut Valley Mesozoic Basin was formed to the west of its major border fault and remains today as a failed arm, possibly an alaucogen, in the abortive attempt to form an ocean basin here (Dewey and Bird, 1970).

The Deerfield Basin filling consists of fresh water sediments of late Triassic age and volcanics and fresh water sediments of Jurassic age (Hubert, et al., 1978). In general the units of the basin dip moderately to the east and southeast toward the major border fault. The basal Sugarloaf Formation consists typically of red arkosic sandstone and conglomerate beds deposited by braided streams and mud flows. The Deerfield Basalt overlies the arkose and is locally a pillow basalt. Overlying the basalt is the Turners Falls Sandstone consisting of sandy red beds with some local lake beds. The Mt. Toby Conglomerate is the uppermost unit in the basin and in part a facies equivalent of the Turners Falls Sandstone. It consists of coarse conglomerate and breccia typically red or beige in color.

Igneous intrusions continued through Mesozoic time as shown by dike swarms of several ages (McHone, 1978). The culmination of igneous activity resulted from the latest phase of the White Mountain Magma Series in the Cretaceous, Mt. Ascutney, Vermont, being the best example in the region.

There appears to be little or no significant effect of the Mesozoic Deerfield Basin on the regional gravity gradient east of the Berkshires. Volcanic rocks of the basin are easily distinguishable on the magnetic map. They appear as extreme lows due to the nearly horizontal dipole associated with them (Harwood and Zeitz, 1977). Overall there is only slight magnetic contrast between the Paleozoic and Mesozoic rocks of the study area (Andreasen and Zandle, 1963), although locally magnetic units do show contrast.

Acknowledgements

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EARLY FOLDING

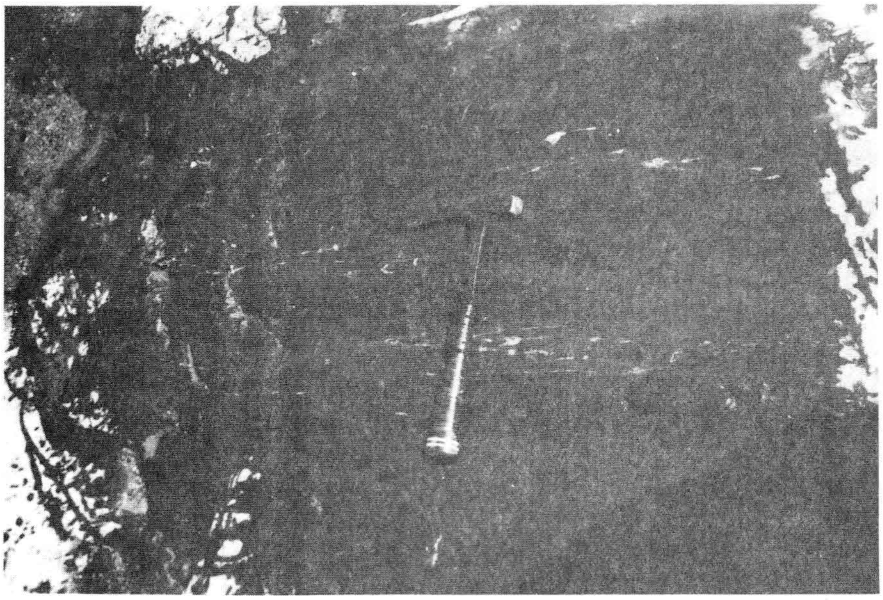
General

The folding of the Littleton Formation in the study area is complex. Multiphase deformation has resulted from the interference of at least seven sets of folds creating remarkably contorted structures. In 1911 C. H. Hitchcock appropriately characterized this area when he stated, "It is a region of great disturbances so that any interpretation is liable to correction". This is still an accurate description today although, there has been considerable progress recently.

Deformation 1

The earliest recognized folding (F_1) consists of isoclinal folds in the original bedding (Figure 5A). This folding produced the pervasive schistosity (Figure 6A) which parallels bedding throughout most of the area and a bedding-schistosity intersection lineation (L_1) locally present in the hinge areas of the few F_1 folds. The best example of this F_1 folding is located beneath the covered bridge on Eunice Williams Road in the town of Greenfield, Massachusetts (4724080 meters north, 695070 meters east, zone 18 of the Universal Transverse Mercator Grid System).

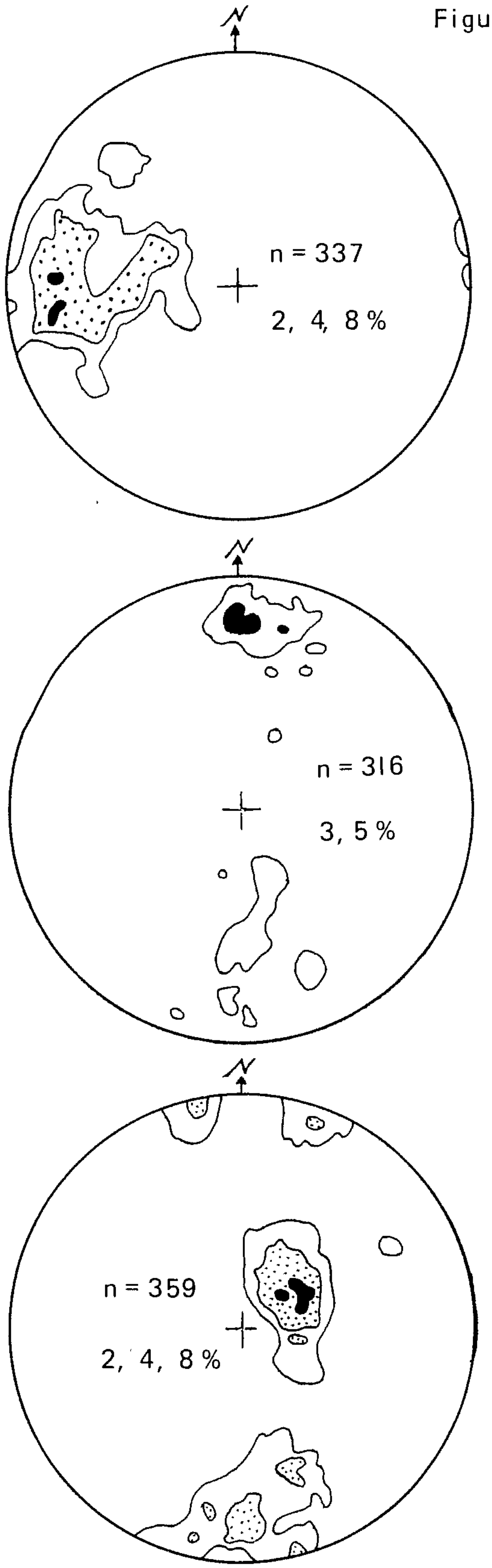
Figure 5: Early isoclinal folds.



D_1 isoclinal folding from under the covered bridge on Eunice Williams Road in Greenfield.

D_2 isoclinal folding from stream valley 0.65 kilometers east-northeast of dam, lower Greenfield Reservoir.

Figure 6: Schistosity, fold axes, and intersection lineations.
(contours in % per 1% area)



a. Poles to schistosity Bernardston-Leyden area.

b. All fold axes Bernardston-Leyden area.

c. All intersection lineations between schistosity and axial plane cleavages, Bernardston-Leyden area.

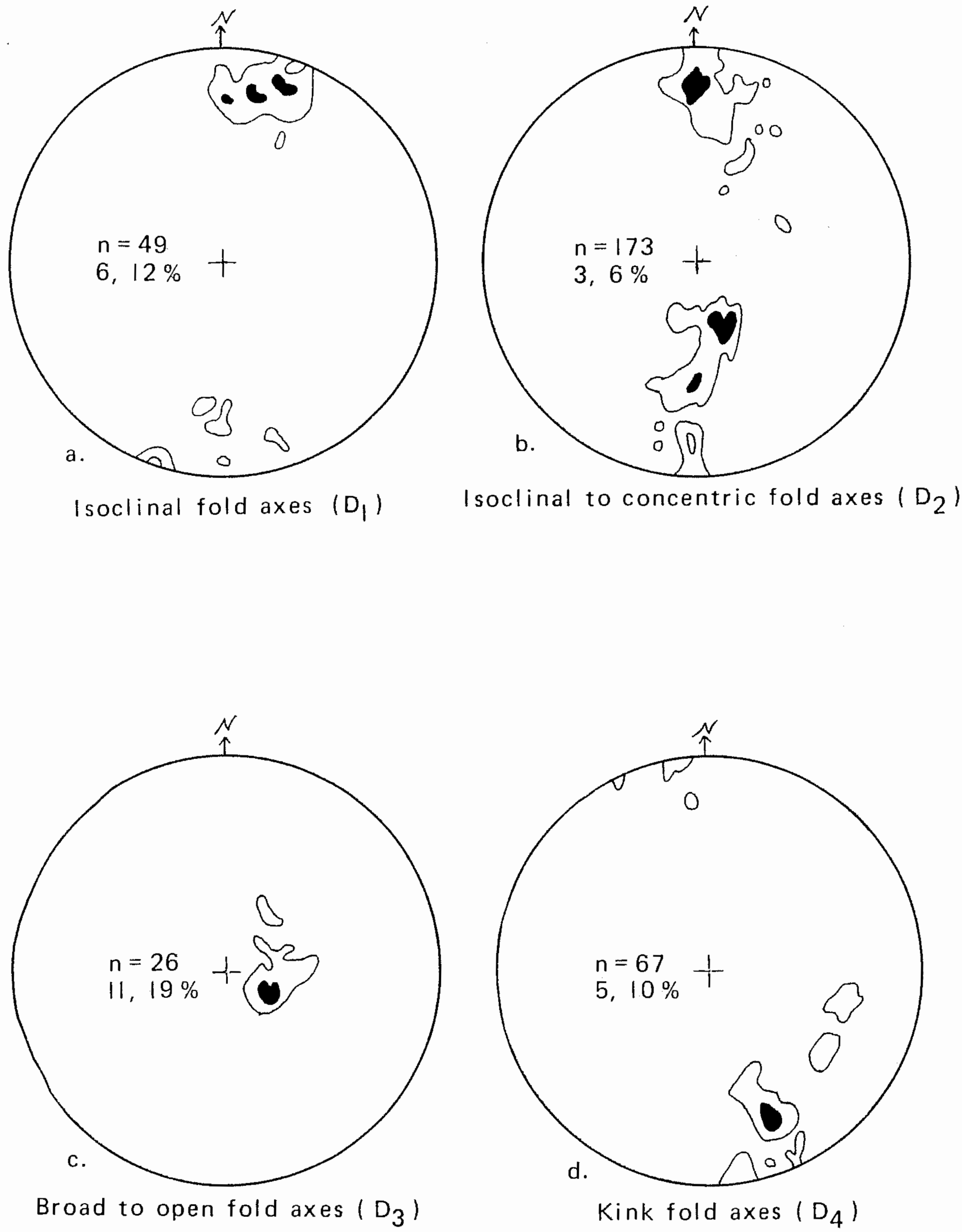


Figure 7: Fold axes by fold type. (contours in % per 1 % area)

Deformation 2

The second and third sets of folds (F_{2A} and F_{2B}) produced isoclinal to tight folds in the schistosity (Figure 5B) with well developed slip-cleavages intersecting the schistosity to form lineations L_{2A} and L_{2B}. These two fold sets are nearly identical in style and orientation and are distinguishable from one another only at locations of mutual interference. F_{2A} and F_{2B} folding are considered relatively syncronous as a second deformation period (D_2). These folds are interpreted as part of the east over west nappe formation during the Acadian Orogeny.

Deformation 3

The fourth and fifth sets of folds (F_{3A} and F_{3B}) are open folds in the schistosity ranging from concentric to broad warps and are generally at a high angle to the earlier coplanar isoclinal folds F_1, F_{2A}, and F_{2B}. These two sets of folds show associated prominent slip-cleavages, or locally, a secondary schistosity where they are tightly oppressed. They produce L_{3A} and L_{3B} intersection lineations with the prominent schistosity. As with F_{2A} and F_{2B} folds, the similarity in both style and orientation of the F_{3A} and F_{3B} prevent distinguishing between them except where they mutualy interfere. These folds are considered to be relatively syncronous and are interpreted as being related to dome emplacement during the peak of metamorphism (Hepburn, 1975, Trask, 1964) toward the end of the Acadian Orogeny.

Deformation 4

The sixth and seventh fold sets (F_{4A} and F_{4B}) are kink bands. Locally these folds have associated fracture or slip-cleavages and produce intersection lineations (L_{4A} and L_{4B}) with the prominent schistosity. These kink bands are interpreted as a synchronous conjugate system. They probably represent a late Paleozoic to early Mesozoic phase of deformation and will be discussed at length in a later chapter.

Although several episodes of folding are well represented, only those related to the nappe formation and the emplacement of the domes are of major significance in the regional structure or the resulting map pattern.

Correlation of Folding with Adjacent Areas (Table 2)

This study correlates well with that of Hepburn (1975) in the Guilford Dome area of southeastern Vermont. The D_2 of this study is represented by two sets of isoclinal folds of the schistosity whereas

Table 2 Regional Correlation of Folding

Conn. Valley	Conn. Valley	Conn. Valley	Berkshires	Bronson Hill *	Merrimack * Synclinorium
This study Bernardston, Ma.	Hepburn (1975) Guilford Dome	Woodland (1977) Royalton, Vt.	Pferd (Per.Com) Colrain, Ma.	Robinson (1967) Quabbin Reservoir Orange, Ma.	Tucker (1977) Field (1975) Barre, Ware, Ma.
Isoclinal F_1 Schistosity	Isoclinal F_1 Schistosity	Isoclinal F_1 Schistosity	Isoclinal F_1 Schistosity	Isoclinal 1A Foliation	Isoclinal 1A Foliation
Isoclinal F_{2A} I. Lineation Slip-cleavage	Isoclinal F_2 Slip-cleavage		Isoclinal	Recumbent 1B	
Isoclinal F_{2B} I. Lineation Slip-cleavage			?	Isoclinal 2A Backfolding Min. Lineation	Isoclinal 2A Backfolding Min. Lineation
Open to Broad F_{3A} I. Lineation Slip-cleavage	Open F_3 Slip-cleavage	Open to Flat F_2 Cleavage	?	Open to Isoclinal 2B b - Lineation	Open to Broad 2B b - Lineation
Open to Broad F_{3B} I. Lineation Slip-cleavage	Open F_4 Slip-cleavage	Open F_3	?		
Kink F_{4A} I. Lineation Cleavage	Broad Warp ** F_5		Kink		
Kink F_{4B} I. Lineation Cleavage	Kink		Kink		

* Not well correlated with this area. ** Possibly associated with kinks.

only one set was recognized in the Guilford Dome area. Hepburn's F_5 broad warps defined in map pattern have not been recognized in the Bernardston-Leyden area.

West over east backfolding of the early nappe structures as recognized by Robinson (1967), Field (1975), Tucker (1977), and Pferd (in progress) as the 2A folding of Robinson may be present in the Bernardston-Leyden area but has not been confirmed in this study.

In the Royalton area, Vermont, Woodland (1977) has not recognized many of the fold systems evident to the south. This may be a result of the higher grade of metamorphism and the more mechanically competent rocks present in the Royalton area.

In the Colrain quadrangle Pferd (personal communication) has recognized a conjugate set of kink bands localized along the eastern third of the quadrangle. In the northern part of this area northeast dipping kink bands show a clockwise rotation sense on east plunging axes and in the southern portion southeast dipping kink bands show a counter-clockwise rotation sense also on east plunging axes.

DIKES AND VEINS

Bernardston-Leyden Area

Highly altered (sericitized) granitic dikes and or sills intrude the Littleton Formation (Figure 8). They appear to be parallel with the schistosity in most cases, have a well developed schistosity themselves, and have been observed to be isoclinally folded, all indicating that they were intruded very early in the deformational history. They may be related to the intrusion of the Kinsman Quartz Monzonite (= Coys Hill Granite) or Hardwick Quartz Diorite to the east, at the onset of the Acadian Orogeny (Tucker, 1977).

Numerous quartz veins of several generations are present throughout the Littleton Formation (Figure 9a). The earliest and most numerous are oriented parallel to the prominent schistosity and are isoclinally folded along with the schistosity. Other isoclinally folded quartz veins cross-cut the schistosity. Later planar quartz veins of several orientations have been formed since the early intense deformations. These have been noted as gash fillings and joint fillings, and fill the dilated zones of the rotated limb in some kink bands. The earlier, generally north-south trending quartz veins are commonly quite large and massive (Figure 10). These probably formed the primary planes of weakness along which Mesozoic faulting took place. This is demonstrated by the large and numerous quartz

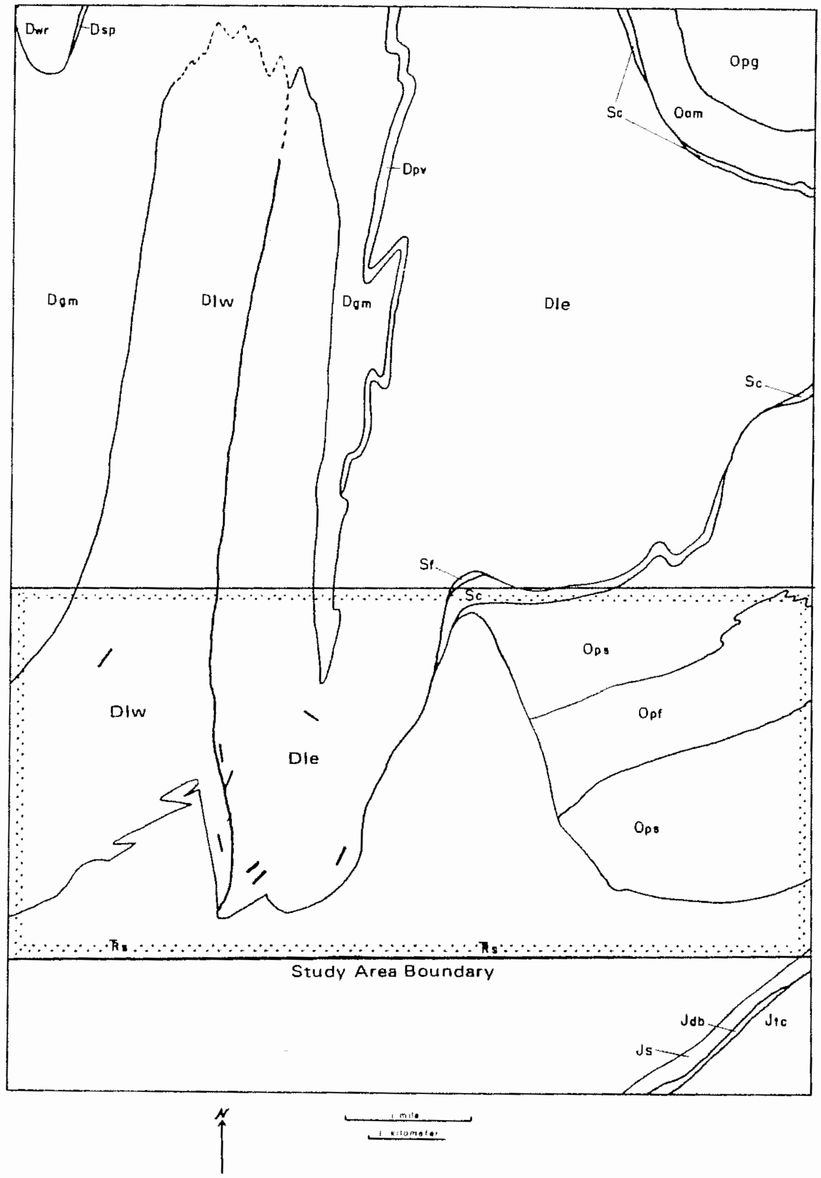
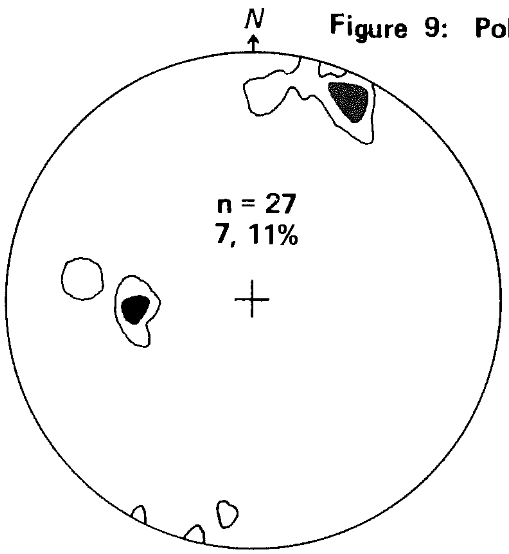
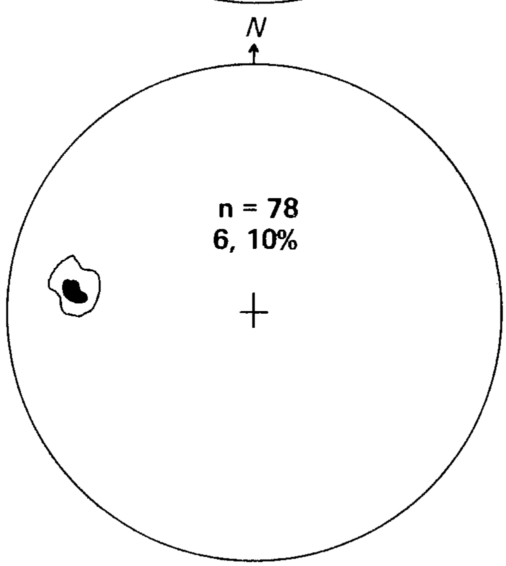


Figure 8: Location and approximate trend of pre-metamorphic dikes or sills in the study area (Note: Most are sill-like with major variability in strike and dip).

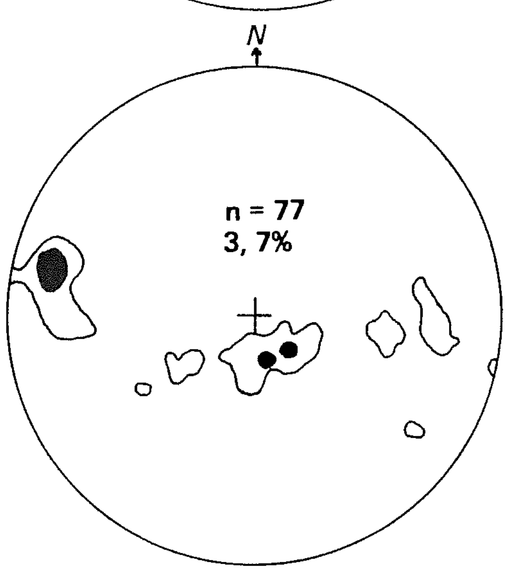
Figure 9: Poles to quartz veins (contours in % per 1% area).



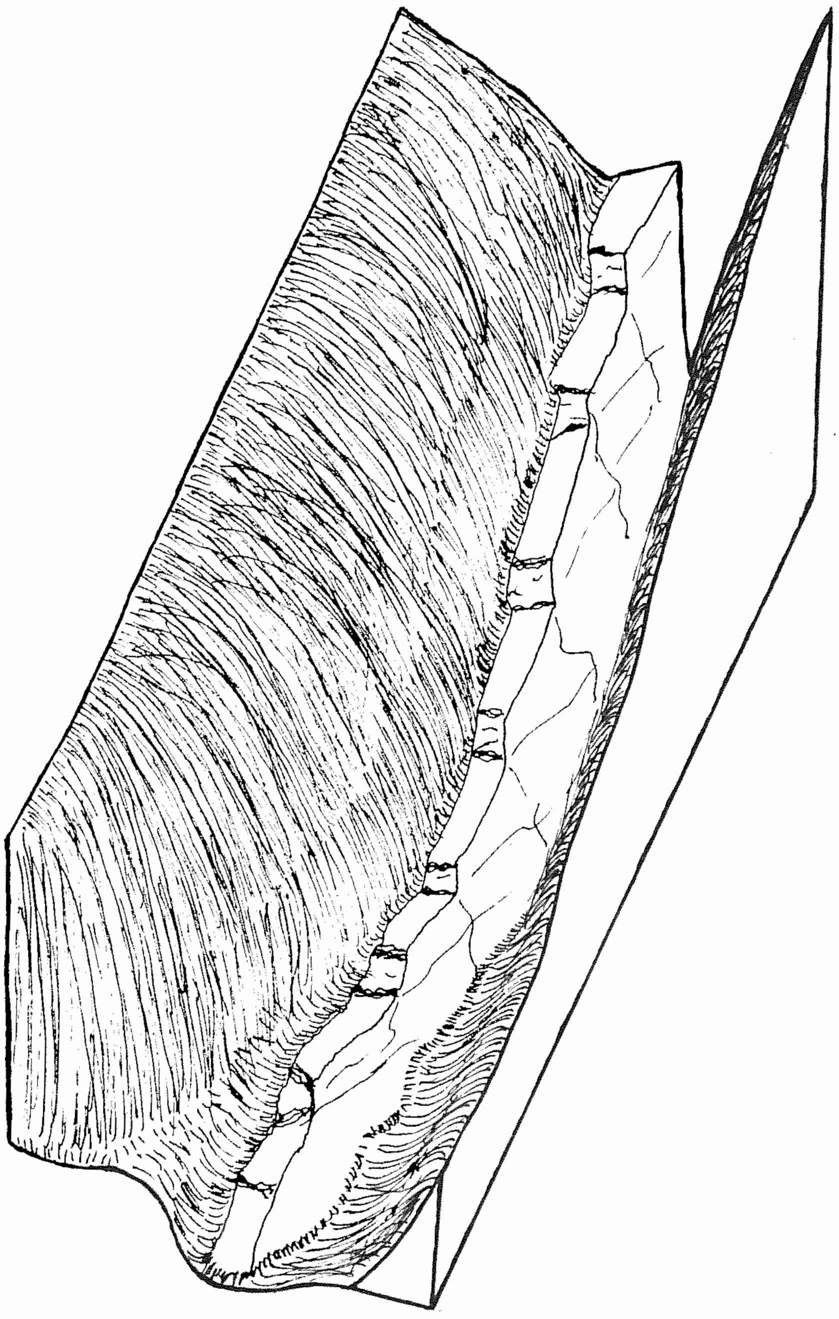
a. Poles to quartz veins, Bernardston - Leyden area.



b. Poles to quartz veins, brittle fracture traverse not including Figure 9a.



c. Poles to quartz veins, Bronson Hill Anticlinorium (Williams, 1976, Silverman, 1976)



Approximately 3 Meters

Figure 10: Sketch of a quartz vein in the center of a deeply eroded stream valley 0.65 kilometers east-northeast of the dam, lower Greenfield Reservoir.

clasts in the Mesozoic rocks adjacent to the fault contact at McCord Brook and by the ubiquity of exceptionally large angular, locally slickensided quartz boulders in virtually every north-south stream valley in the Paleozoic rocks. Sporadic calcite filled joints or small calcite veins were noted in the eastern portion of the Littleton Formation. Quartz veins, as such, were not observed in the Mesozoic rocks of the study area, although, several joints were noted with drusy quartz surfaces.

Brittle Fracture Traverse

Quartz veins along the northwestern Massachusetts traverse show a strong north-south to N 10°E dominance (Figure 9b) which coincides with the general trend of the foliation across the region. At most stations, early quartz segregation parallel to the foliation is a prominent feature. In the Taconic Mountains quartz veins show orientation maxima of N 50° W and east-west as well as the regional north-south trend. Relative ages of these veins have not been assigned as no cross-cutting relationships were observed. Quartz veins in the Berkshire area (Figure 9b) are dominated by N 10°E trends, parallel to the regional foliation. Other orientations are present locally and are consistently younger than the prominent north-south system. The Connecticut Valley and Bronson Hill areas are also dominated by north-south trending quartz veins parallel to the regional foliation (Figure 9c). Several other generations of quartz veins are also represented but no relative ages have been determined for these. The coincidence in the trend of the north-south system of quartz veins with a roughly north-south fault system in these areas may be a significant factor in the development of the Connecticut Valley Mesozoic Basin. The mechanical anisotropy created by some of the large quartz veins may have controlled the propagation of some normal faults along these planes of weakness as discussed above. Alternatively, some of the strains associated with early stages of the faulting may have controlled some of the quartz vein orientations.

LATE STAGE KINK BANDS

General

Kink bands are generally considered to be small-scale monoclinal folds with planar limbs and sharp angular hinge zones (Figures 11 and 12). Their development is usually restricted to rocks having a strong, closely-spaced, planar anisotropy (foliation or bedding) (Verbeek, 1978). They are a common feature in areas of multiphase deformation. Their occurence is controlled by two primary factors. First and most important, is the orientation of the local foliation with respect to σ_1 (the maximum compressive stress). Kink bands are developed with the foliation preferentially oriented parallel to σ_1 but, varying up to 30^o from it, conjugate sets being developed where σ_1 is oriented within 5^o of the foliation. Second, the composition of the layering plays an important part. Kink bands in the study area are prominently developed in the more homogenously mica-rich unit comprising the eastern section of the Littleton Formation (Figure 4). In the western section the interbedding of the mechanically stronger quartz-rich interlayers prevents the prominent kink band development seen in the more homogenous mica-rich eastern section.

Kink bands have been the subject of many detailed experimental studies (Donath, 1968, Anderson, 1968, Dewey, 1965, Fyson,

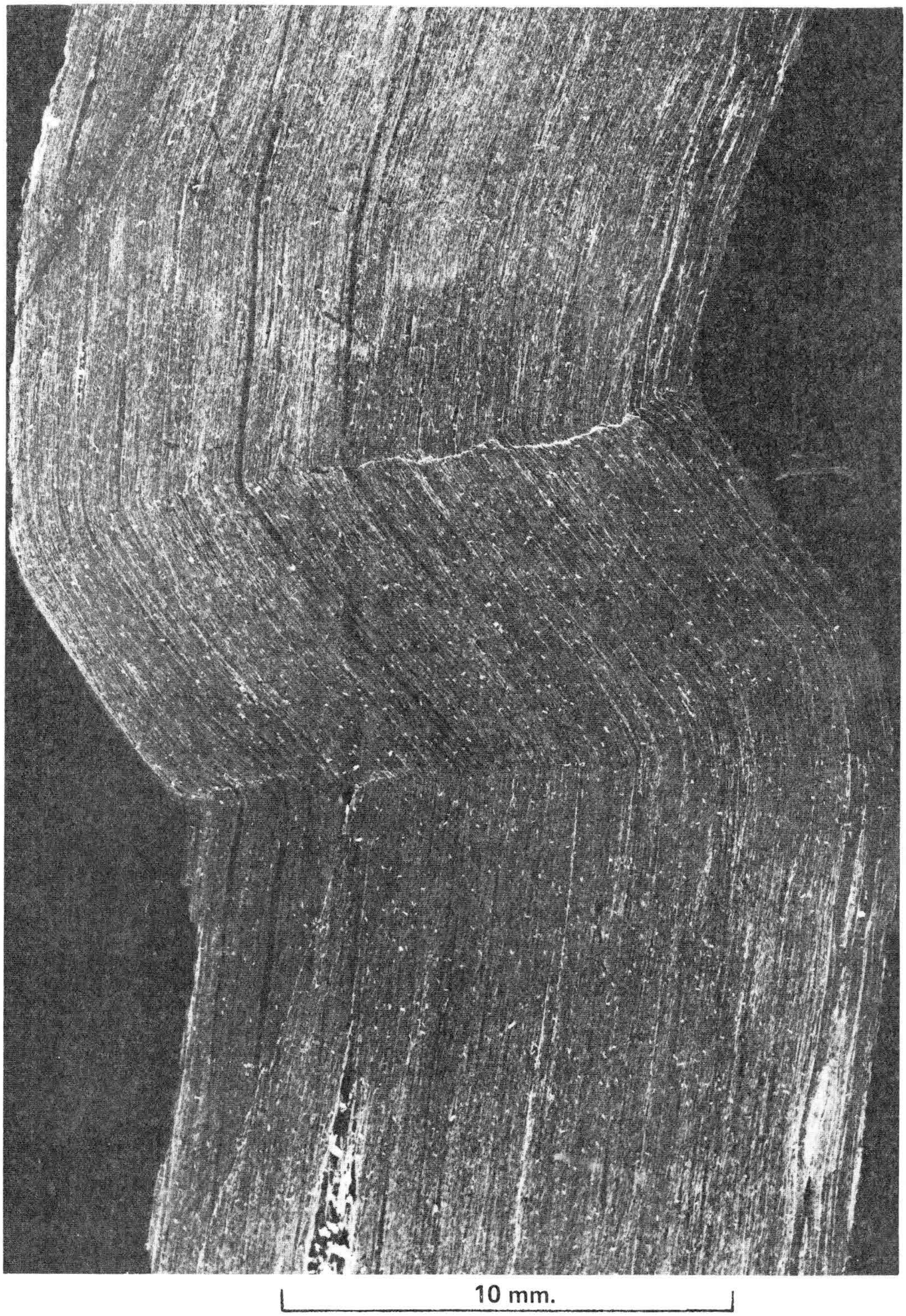


Figure 11: Photomicrograph of a kink band in thin section, from the lower member of the Littleton Formation. Phyllite at biotite grade.

Figure 12:
Kink bands from the lower member of the Littleton Formation. Phillite at biotite grade.

1968, Paterson and Weiss, 1966, and Gay and Weiss, 1974, and others) but until recently experimental results have not been applied to field investigations. Kleist (1972) used kink bands to determine the σ_1 orientation along a segment of the Denali Fault in Alaska, and Verbeek (1978) used kink bands for σ_1 determinations in the Pyrenees at the France-Spain border.

Ideal Kink Band Models

Verbeek (1978) summarizes the various ideal kink model end members which have been experimentally determined or observed in nature. He describes the processes of development and the characteristics associated with each ideal model. (See Figure 13 for angular relationships and kink geometry, Gay and Weiss, 1974).

Migration Model. Patterson and Weiss (1966) experimented with a phyllite similar to the phyllite of this study. They describe kink band development as initiating at a point (Figure 14) and growing by lateral migration of the kink boundaries into the adjacent undeformed rock during progressive strain. All of the angular shear strain, due to slip along foliation, takes place within the kink band. Outside the kink band the rock is virtually undeformed. As the kink band migrates into the undeformed region (expands in width with time) each point within the kink band has passed through the kink

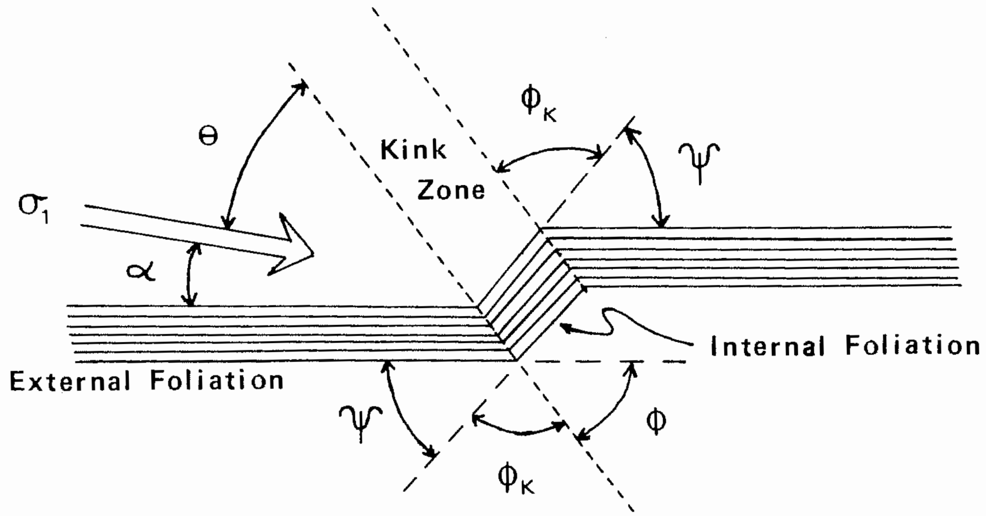
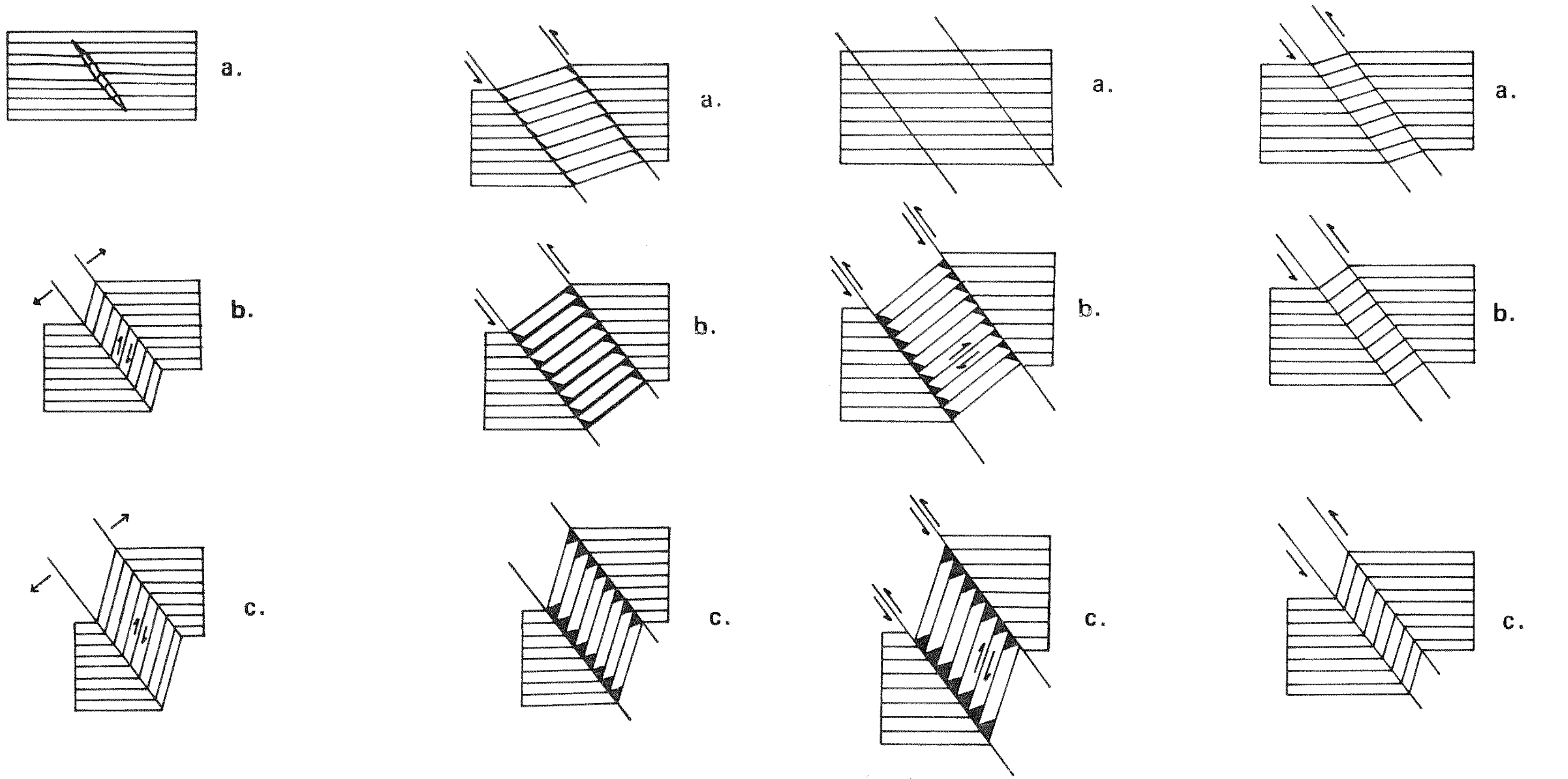


Figure 13: Angular relationships between foliation and kink zone used for σ_1 determinations. (Gay and Weiss, 1974)



Migration Model (Paterson and Weiss, 1966)

Rotation Model (Donath, 1968, Clifford, 1968)

Joint Drag Model (Flinn, 1952, Dewey, 1965)

Simple Shear Zone Model (Dewey, 1965)

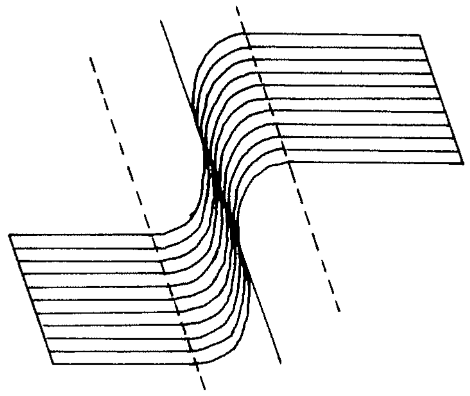
Figure 14: Ideal kink band models.

boundary and been reoriented parallel to the internal kink foliation, being first folded and then unfolded.

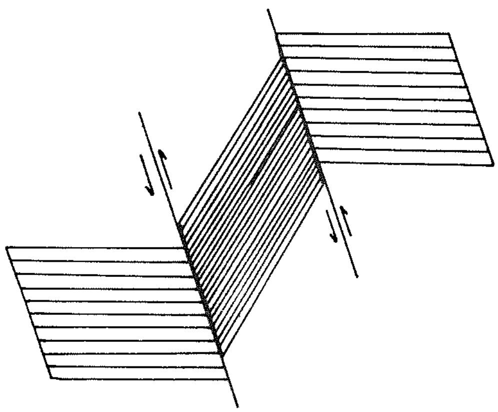
Rotation Model. In this model proposed by Donath (1968) and Clifford (1968) it is presumed that the positions of the kink boundaries become fixed early in the deformational history (Figure 14) with ϕ_k increasing to 90^o then decreasing with progressive deformation. Dilation of the kinked limb increases with rotation to a maximum at $\phi_k=90^o$. Further rotation reduces this dilation until $\phi_k = \phi$ where dilation is again zero, and theoretically can progress no further or becomes "locked". Commonly during the dilation phase minerals such as quartz or calcite are precipitated in the resulting voids preventing the preferred $\phi_k = \phi$ relationship from developing. Triangular voids also occur at the kink boundaries due to the brittle behavior assumed during rotation of the kinked limb.

Joint Drag Model. This model proposed by Flinn (1952) and Dewey (1965) is characterized by breakage and offset of the layers by shear along the kink boundaries (Figure 14). The boundaries are formed by breakage of the layers during initial strain as the layers are rotated, similar to the rotation model but with slip between the layers rather than dilation. The angular relationships between ϕ and ϕ_k are not restricted in this model due to the discontinuities at the kink boundaries and slip between the layers rather than dilation, allowing the kinked limb to rotate beyond $\phi_k = \phi$ to where $\phi_k < \phi$. This model may not represent true kinking in itself but rather parallel or conjugate shear zones. It should be noted however, that many kinks observed do display discontinuous layers across the boundaries which may result from continued strain on kinks of the rotation model. This rotation, upon reaching the maximum $\phi_k = \phi$ angular relationship, is continued by shearing along the boundaries allowing further rotation of the kinked limb, with ϕ remaining constant after shearing. Thus, this may be considered a special case of the rotation model.

Simple Shear Zone Model. The kink boundaries in this model (Dewey, 1965) are generally poorly defined (Figure 14). Simple shear parallel to and confined within a restricted zone, with or without mechanical participation by the internal foliation (slip between the layers), causes end effects at the boundaries. These end effects dissipate the shear effect in this region and result in rounded hinges at the margins with no discrete kink boundaries (Figure 15). This model may not represent true kinking but may help to explain the rounded nature of some kink boundaries and the common ambiguity of these boundaries. This model may however, be a mechanism for modifying some kink bands.



Modified Simple Shear
Zone Model
(Dewey, 1965)



Strain Band Model
(Dewey, 1965)

Figure 15: Other ideal kink band models.

Strain Band Model. This model proposed by Dewey (1965) is characerized by migration of silica through solution within the high strain area concentrated in the kink band and precipitation in the strain free adjacent rock (Figure 15). The resulting kinked limb, due to this silica migration, becomes more pelitic than the adjacent rock outside the kink boundaries and the internal layers are thinned allowing them to rotate beyond the otherwise limiting $\phi_k = \phi$ relationship of the rotation model, without the development of discontinuities at the boundaries. This process may be another modification of the rotation model, explaining some of the smaller ϕ_k's observed in some thin sections and the more thinly laminated kinked limbs observed in some kinks in the sections.

Discussion. These various models are of considerable importance in explaining the broader concepts of the mechanics involved in kink band development. However, kink bands observed in this study do not conform to any of the ideal models discussed above. The kink bands of this study area display components of many of these models. They are most similar to the migration and rotation models, lacking the triangular voids at the boundaries, indicating that they may be more closely related to the migration model or modified by the strain band model. They are characterized by sporadic discontinuities at the boundaries (Figure 11), migration of silica both into and out of the kinked zone, and rounded hinge areas (Figure 11). Thus the ideal models are limited when applied to naturally formed kinks. These models are further limited in that they do not consider pre- or post-kink strain which may be more important than the kinks themselves (Verbeek, 1978), and in some models the assumption of brittle behavior, as there is certainly ductile flow at the hinge areas. In addition cataclasis has been observed within kink boundaries (Donath, 1968). Early slip between layers is a ubiquitous and necessary prelude to kink band development, yet it is inherent in none of the ideal models (Verbeek, 1978). Later strain on the kink bands could result in further thinnng and lengthening of their inter-kink planes or flattening of the entire structure. This subsequent strain could alter the ϕ_k and ϕ angles and produce differential strain-induced solution and precipitation between the kinked and unkinked regions.

Kink Bands as Stress Indicators

The relationship between conjugate sets of kink bands and σ_1 (maximum compressive stress) is well established, σ_1 being the obtuse bisector of the conjugate kink boundaries. Recently, through experimental work, certain angular relationships have been recognized which allow the prediction of the approximate σ_1 orientation operating at the time of kink development when only a single kink set is present. Since the kink axis is only sub-parallel to the σ_2 orientation, only an approximation of the σ_1 orientation can be predicted (See figure 13 for definitions of angular relationships). Gay and Weiss (1974) determined that the angle Θ between the kink boundary and σ_1 varies predictably with a change in the angle α between the external kink foliation and σ_1. Θ varies from about 45^o at α's maximum value of 30^o to about 60^o at $\alpha = 0^o$. The α value is determined from the consistency of the measured angular relationships in experimentally formed kinks (Figure 16). The angle ϕ between the kink boundary and the external foliation varies from about 60^o at $\alpha = 0^o$ to about 80^o at $\alpha = 30^o$. ϕ bears the most reliable linear relationship to in the results of Gay and Weiss (1974) and is given the most weight in the prediction of α. The angle ϕ_k between the kink boundary and the internal kink foliation varies from about 70^o at $\alpha = 0^o$ to about 90^o at $\alpha = 30^o$. The angle ψ between the internal and external kink foliation varies from about 50^o at $\alpha = 0^o$ to about 12^o at $\alpha = 30^o$. These last two linear angular relationships (ϕ_k, ψ) are less reliable but can help in determining α more accurately. In that these relationships are not totally accurate, several kink bands should be measuresd at each area to provide a more reliable mean value for α and σ_1. There are several possible solutions using this technique. However, these solutions are limited by certain constraints: 1. σ_1 must lie within 30^o of being parallel with the external kink foliation for the development of kinks in preference to slip along the foliation. This slip occurs at α values greater than 30^o and probably before kink

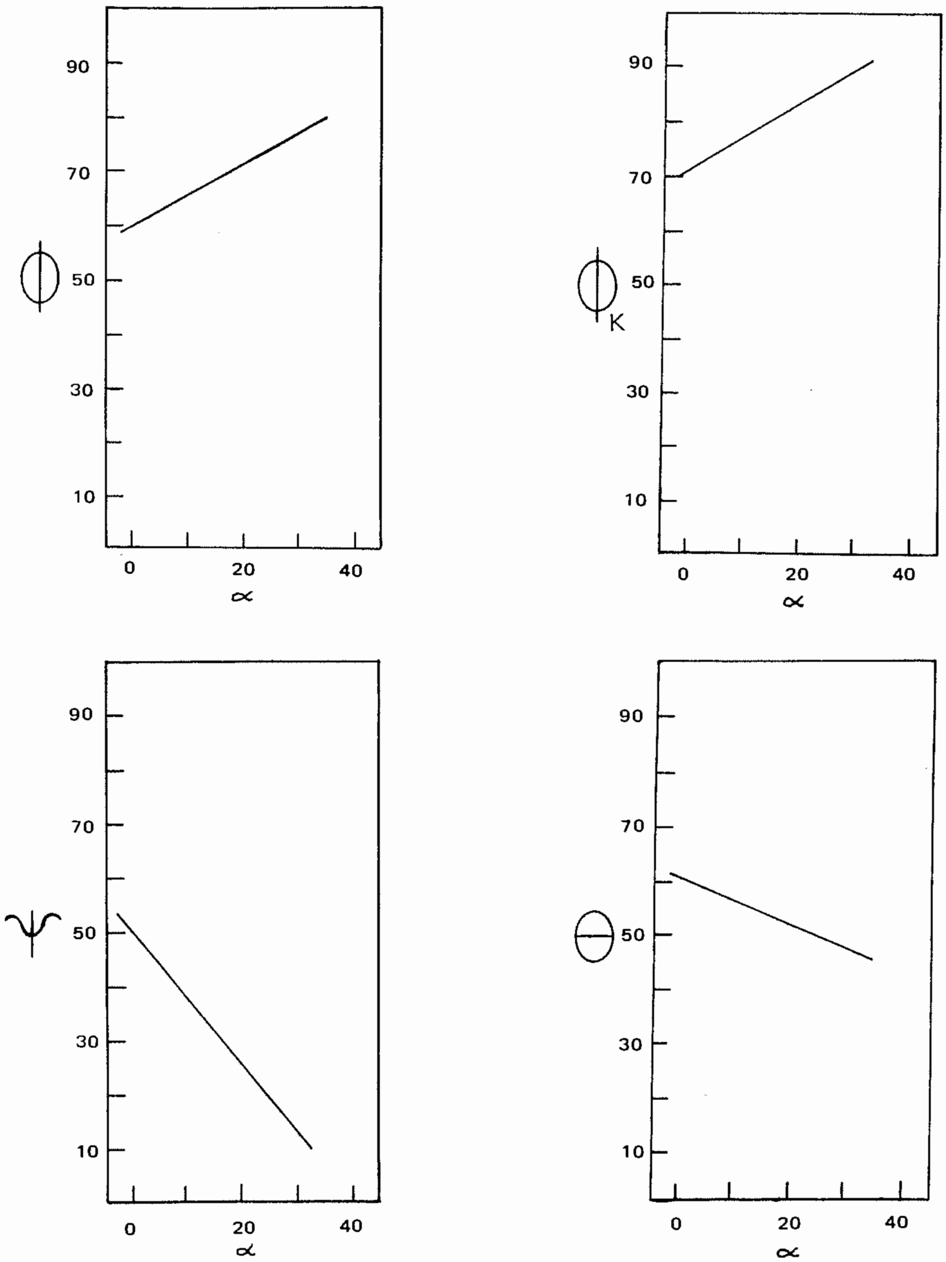


Figure 16: Best fit curves for kink band angular relationships with varying α's (Gay and Weiss, 1974) experimentally determined for the Migration model. For definitions of Φ, Φ_k, Ψ and Θ refer to Figure 13.

initiation. 2. σ_1 usually lies in the acute angle (ϕ) formed by the kink boundary and the external foliation, closer to the foliation. 3. σ_2 is assumed to lie sub-parallel to the kink axis with σ_1 lying in the plane approximately normal to the kink axis. These constraints usually limit the possible solutions to one. The solutions for each kink measured at a station are analysed for their maximum concentration which is taken to be the best approximation of the σ_1 orientation during kink band formation. The resulting solution should be compatable with the orientation of the average external foliation (within 30^{o}) in which the kinks are developed.

Gay and Weiss (1974) state that the principal stress directions at the instant of kink band nucleation appear to control the orientation of kink boundaries. But, subsequent growth of kinks involves first, a perturbation of the initial stress state, and second, a change in the angles in the kink system as a result of progresive "backward rotation" (rotation of the external foliation toward parallelism with σ_1) in the undeformed domains. In my opinion this "backward rotation" of Gay and Weiss may be due to edge effects caused by the confining cylinders used in their experiments or in nature by very closely spaced kink bands. Widely-spaced, narrow kink bands represent very small aggregate strains for the body of rock as a whole and probably are not subjected to any gross rotation, making these the most reliable for determination of the principal stress directions.

At the other extreme, as the lengths of the internal and external limbs approach being equal, the mechanical behavior may be drastically altered and "backward rotation" may be a result.

Areal Distribution of Kink Bands

Kink bands are present in the argillaceous rocks along the low-grade metamorphic core of the Connecticut Valley - Gaspe Synclinorium (Leo Hall, personal communication). In this study area the distribution of kink bands (Figure 17) is controlled primarily by the orientation of the prominent schistosity and also by the mechanical behavior of the layering comprising the rock. The orientation of the schistosity and the homogenously mica-rich composition of the eastern section of the Littleton Formation provide the proper conditions for kink development.

In the western section of the Littleton Formation the interbedding of mica-rich with quartz-rich layers and the relative absence of preferentially oriented schistosity, severely limit kink development. Locally, in the mica-rich layers, where the schistosity in the western section is preferentially oriented, kink bands do develop only to die out upon reaching the quartz-rich layers or to be dissipated as small-scale shears in the quartz (Figure 18). Due to their small size and the numerous other intersection lineations superposed on the

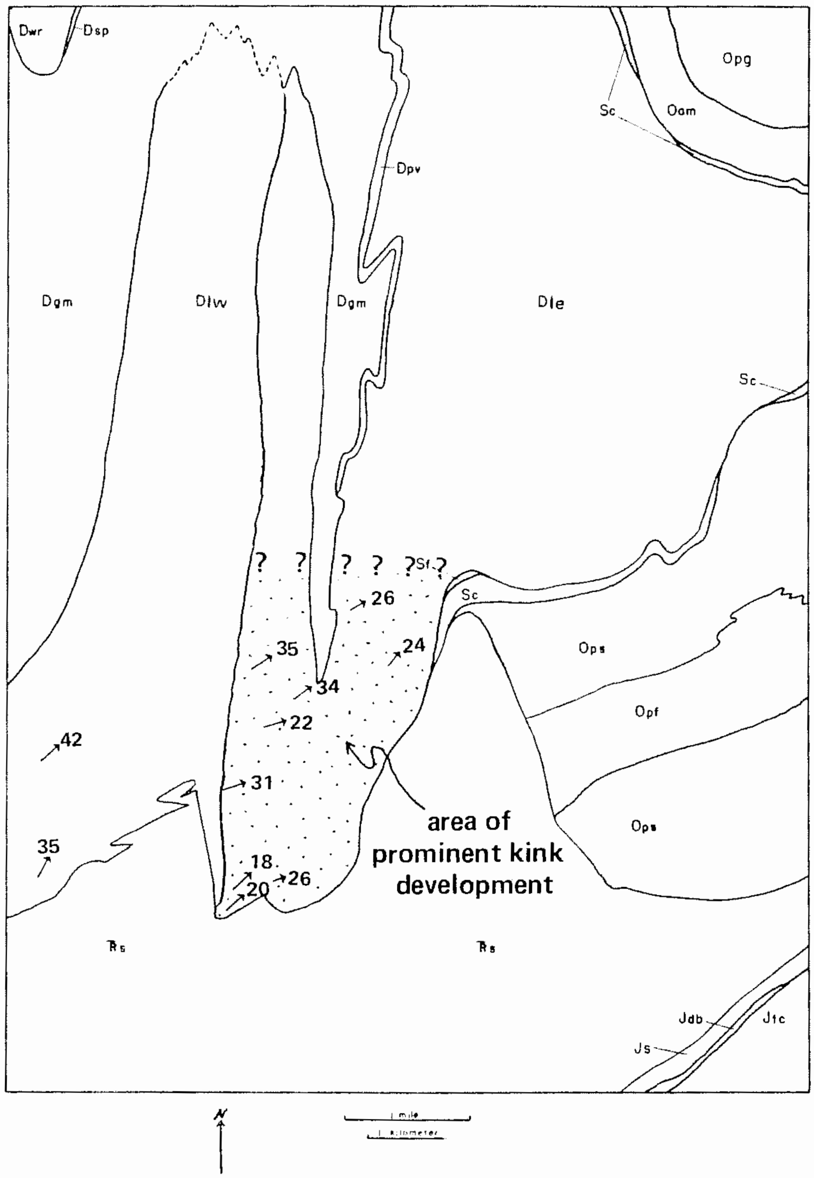


Figure 17: Bernardston quadrangle index map showing σ_1 determinations and area of prominent kink band development in the study area. (data are plotted on Figure 20)

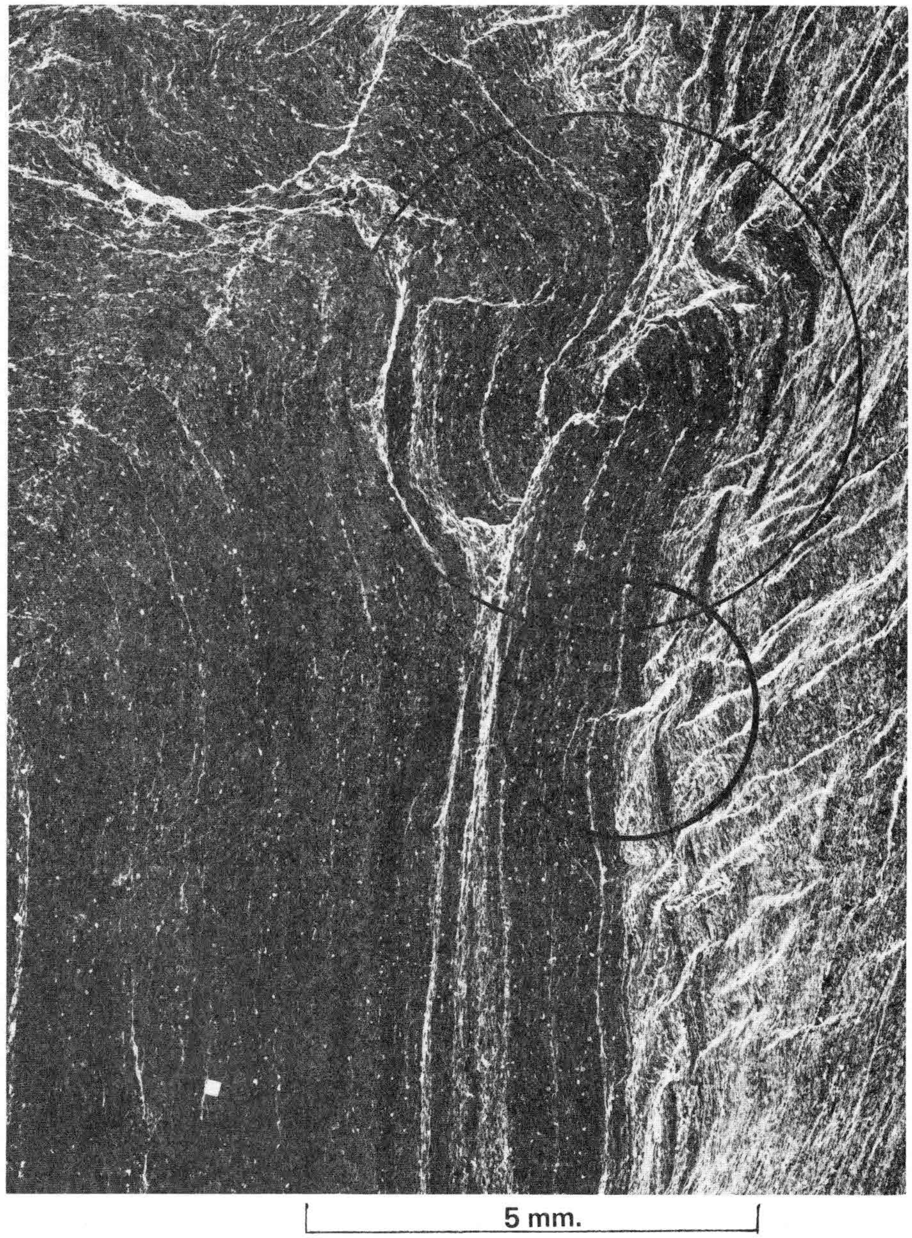


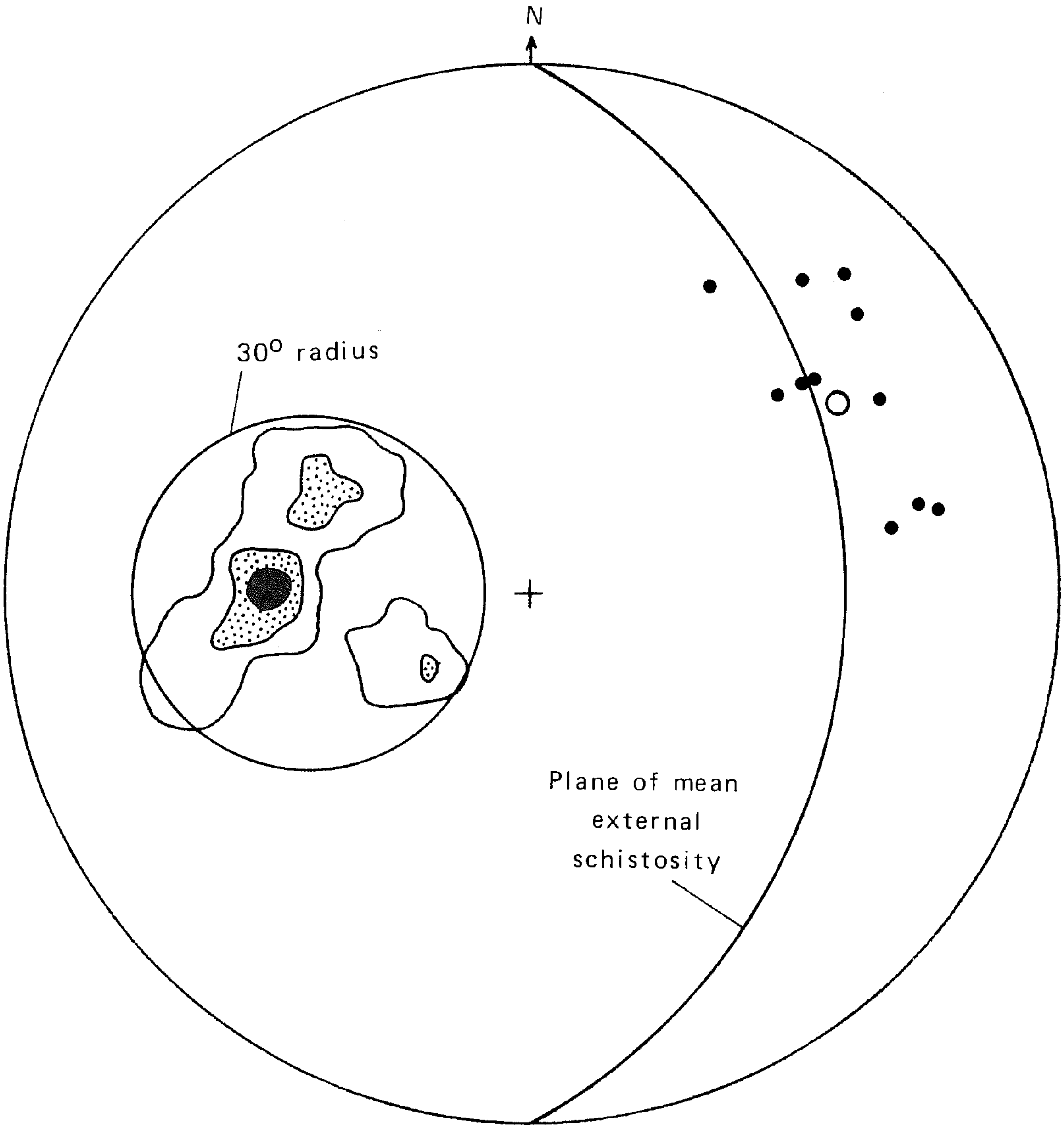
Figure 18: Photomicrograph of thin sectionshowing kinks in the mica-rich layers (right) being transformed into shears in the quartz-rich layers (left).

schistosity these kinks are not readily seen in the field but they are obvious in thin section. The orientation of schistosity in the western section is not particularly favorable with respect to σ_1 but, locally on many of the small-scale folds proper conditions are met for kink formation. σ_1 determinations from these kinks are difficult although, an approximate value can be determined from the limit of schistosity orientations on which the kinks are developed.

Principal Stress Trajectories from Kink Bands

Several kink bands were measured and analysed for each of eleven separate stations using conjugate relationships, where present, to determine the σ_1 orientation for each area. The methods of Gay and Weiss (1974), as described above, were used where only one of the potential conjugate pair was present (Figure 16). Where both sets were present, the σ_1 and σ_2 orientations were determined by the obtuse bisector of the kink band boundaries and the line of intersection of these boundaries respectively.

The overall results for the study area indicate a σ_1 orientation of about N56^{o}E plunging about 32^{o} to the northeast (Figure 19). The ideal schistosity orientation for the development of these kinks must be parallel to the σ_1 determination. The maximum of poles to the external kink schistosity (Figure 19) represents a plane striking north-



● σ_1 dterminations at individual stations (see Figure 17).

○ Mean σ_1 orientation = N 56° E, 32° NE.

Figure 19: σ_1 determinations from kink bands and related external schistosity. (contours on 52 poles to external schistosity = 7, 15, and 21 % per 1 % area).

south and dipping 38° to the east and is within two degrees of being parallel with the σ_1 orientation determined for the study area, as well as varying from the σ_1 orientation determined for any one station by no more than about 15°. Assuming that the σ_1 determination for each station is fairly accurate, the spread of the results (Figure 19) may be an effect of local structure deflecting the regional principal stress.

Depth and Timing of Kink Band Development

While the development of kink bands in the study area is the latest episode of ductile deformation, the depth and timing of their formation are only vaguely constrained. They are presumably low temperature and low pressure features which probably developed in the upper few kilometers of the crust. No mica recrystallization has been observed associated with the kinks, and the only kink associated mineralization is the differential solution and precipitation of quartz into and out of the kinked limbs. The rocks in the area of kink development reached only chlorite, biotitie, and garnet grade of metamorphism during the peak of Acadian metamorphism (Hatch, 1975). The kinks are certainly post-metamorphism features and occured at depths less than those required for metamorphic conditions. In some instances there is a transition from a fold to a fault along the strike of a kink boundary indicating ductile to brittle transition. The development of the kinks post-dates the dome emplacement, as open to concentric folds associated with the doming are deformed by kinks. Since the emplacement of the domes, conditions conducive to the development of these kink bands have existed at several times (Table 1A). Isostatic uplift and erosion following the doming may have produced suitable conditions for their development. Hepburn (1975) places the kinks in the Guilford Dome area (Figure 2) at the time of his latest F_2 folding (Table 2) or younger. In the Bernardston-Leyden area the kink bands are certainly the youngest folding event observed. Other possible times of kink band development in this area include: 1. Carboniferous, during the formation of the Carboniferous basins of eastern New England. 2. Early Mesozoic, as the Connecticut Valley Mesozoic Basin began to open, first in the south, then progressively northward. The center of rotation associated with this opening might progress northward along the resulting axis of the basin. The northward migration of this rotation center could reverse the σ_1 - σ_3 orientations as it passes through a point, leading from compressive kinking north of the center to extensional rifting south of it. Early Mesozoic dikes (McHone, 1978) also indicate a suitable stress field for the kink formation at this time. 3. Post early Mesozoic strike-slip faulting in the Mesozoic rocks of the study area cross-cutting the Jurassic Deerfield Basalt (Goldstein, 1975). These faults are of a compatable orientation for strike-slip motions with respect to the stress trajectories determined for the kink bands, and are similar to the relationship of kink bands to motion on the Denali fault in Alaska as demonstrated by Kleist (1972).

A late Paleozoic to early Mesozoic time is favored for kink band formation. Conditions would place the kinked Paleozoic rocks essentially at the surface by middle to late Triassic time and keep them there. The presence of alluvial fan deposits in the Mesozoic rocks with predicted locations of fan apexes and southerly directed paleocurrents at the head of the Mesozoic Deerfield Basin (Stevens, 1977) indicate that the present northern extent of the basin is close to its maximum extent.

These kink bands represent a transition from ductile to brittle deformation in this area and may well bridge the gap from more or less east-west compression during the Acadian deformations to the northwest-southeast extensions indicated by the Mesozoic dikes associated with the formation of the Connecticut Valley Basin.

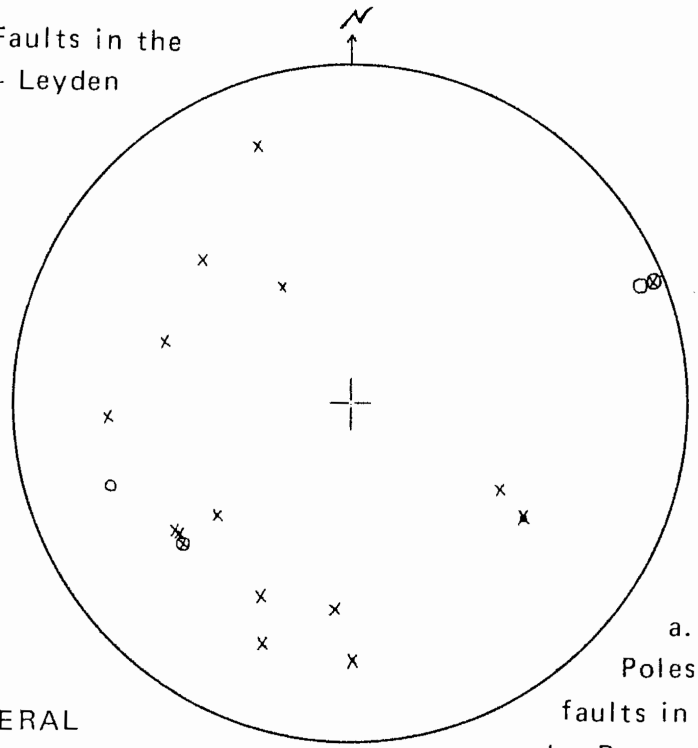
FAULT PATTERNS

Paleozoic Rocks

Faulting in the Paleozoic rocks of the Bernardston - Leyden area (Figure 20) is predominantly dip-slip normal faulting and conforms to the regional pattern defined by the brittle fracture traverse along the Massachusetts - Vermont border (Figure 3). These normal faults strike roughly northeast and northwest with the east side down on the northeast striking faults and the west side down on the northwest striking faults. Some right-lateral strike-slip faulting was observed in the vicinity of the fault contact with the Mesozoic rocks at McCord Brook (Figure 4) and in the valley of the east branch of McCord Brook further within the Paleozoic rocks. Balk (1956) mapped the contact between the upper and lower Leyden Argillite in this valley. These strike-slip faults strike about N 10°W and are believed to be of Mesozoic age, as similar right-lateral faults were observed in the Mesozoic rocks on the west side of McCord Brook.

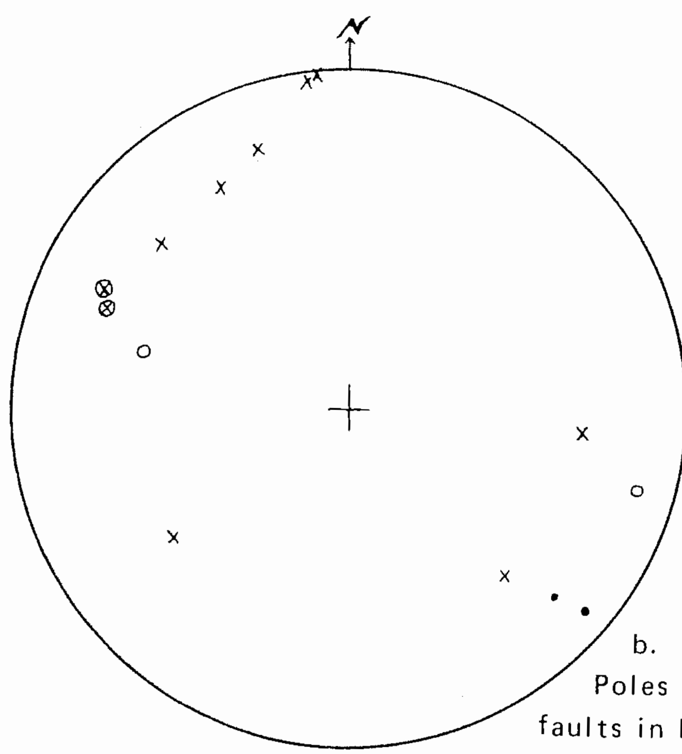
There is also a family of roughly north-south striking faults. The few slickensided surfaces present indicate these are normal faults. Virtually every north-south stream valley contains large quartz boulders, commonly slickensided. The valleys appear to follow large quartz veins which are parallel to the schistosity. These valleys

Figure 20: Faults in the Bernardston - Leyden study area.



× NORMAL
○ RIGHT - LATERAL
· LEFT - LATERAL

a. Poles to faults in Paleozoic rocks Bernardston - Leyden study area.



b. Poles to faults in Mesozoic rocks Bernardston - Leyden study area.

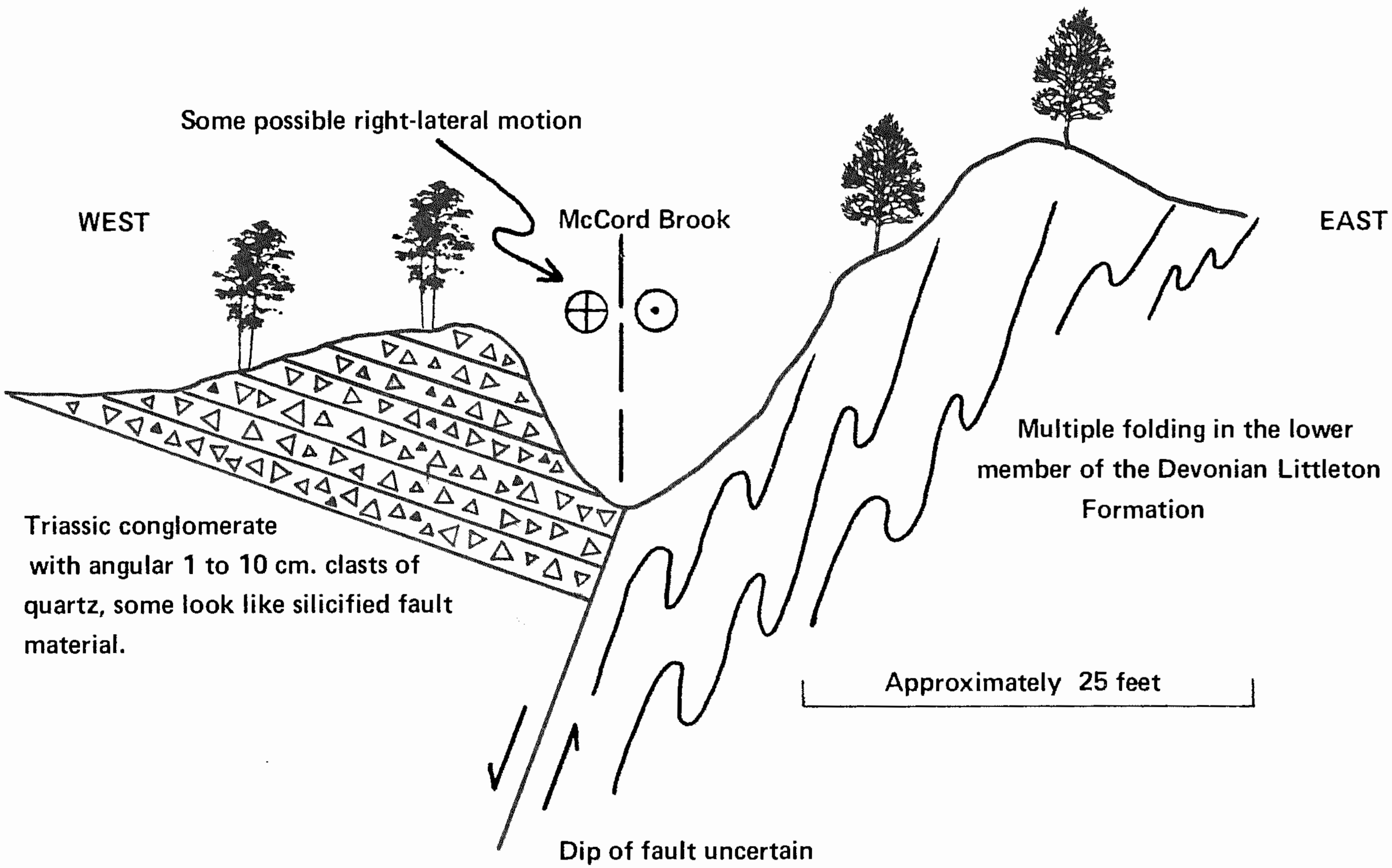
apparently are also fault controlled, with faulting concentrated in or at the margins of the quartz veins. The McCord Brook fault area has large (1 - 10 cm) quartz clasts throughout the Mesozoic rocks indicating faulting along an older quartz vein or silicified zone (Figure 21).

Mesozoic Rocks

Faulting in the Mesozoic rocks of the Bernardston - Leyden study area (Figure 20) is a combination of both normal and strike-slip faults. The normal faults are similar to those of the Paleozoic rocks, striking northeast, northwest or east-west. Sparse strike-slip faults show both right-lateral (six observations) and left-lateral (three observations) motions. Faults with right-lateral motions strike roughly north-south to north-northeast near the fault contact at McCord Brook. Left-lateral strike-slip faults strike about N 40^{o}E near the fault contact at Glen Brook (Figure 4). These strike-slip faults may represent a conjugate system similar to that proposed by Goldstein (1975) but are inadequate in number for reasonable statistical work. If they are parts of a conjugate set, the most probable σ_1 orientation would be about N 20^{o}E, or approximately parallel to the axis of the Mesozoic basin (Figure 22).

Illies (1975) found a similar relationship in the Rhine Graben. Using styolites as stress indicators he determined σ_1 to be hor-

Figure 21: Faulting at McCord Brook ½ mile east of the Greenfield water treatment plant.



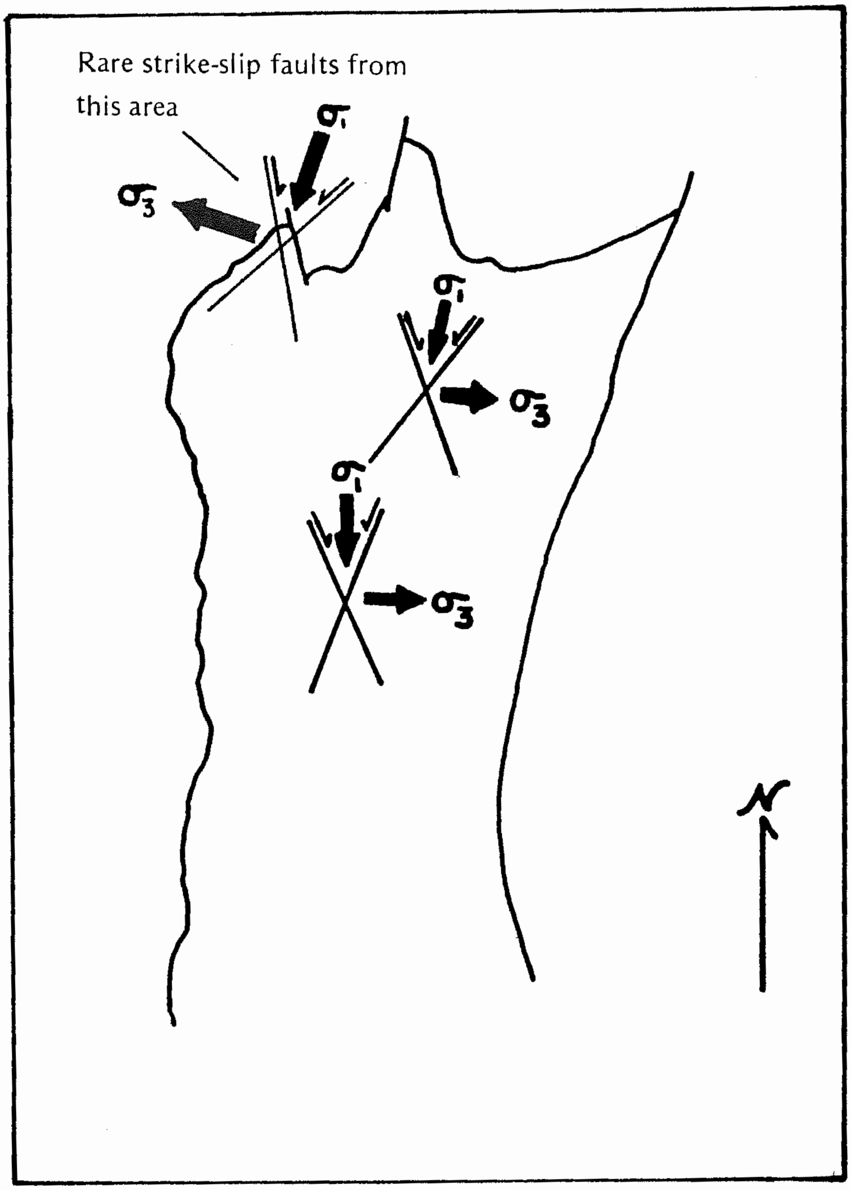
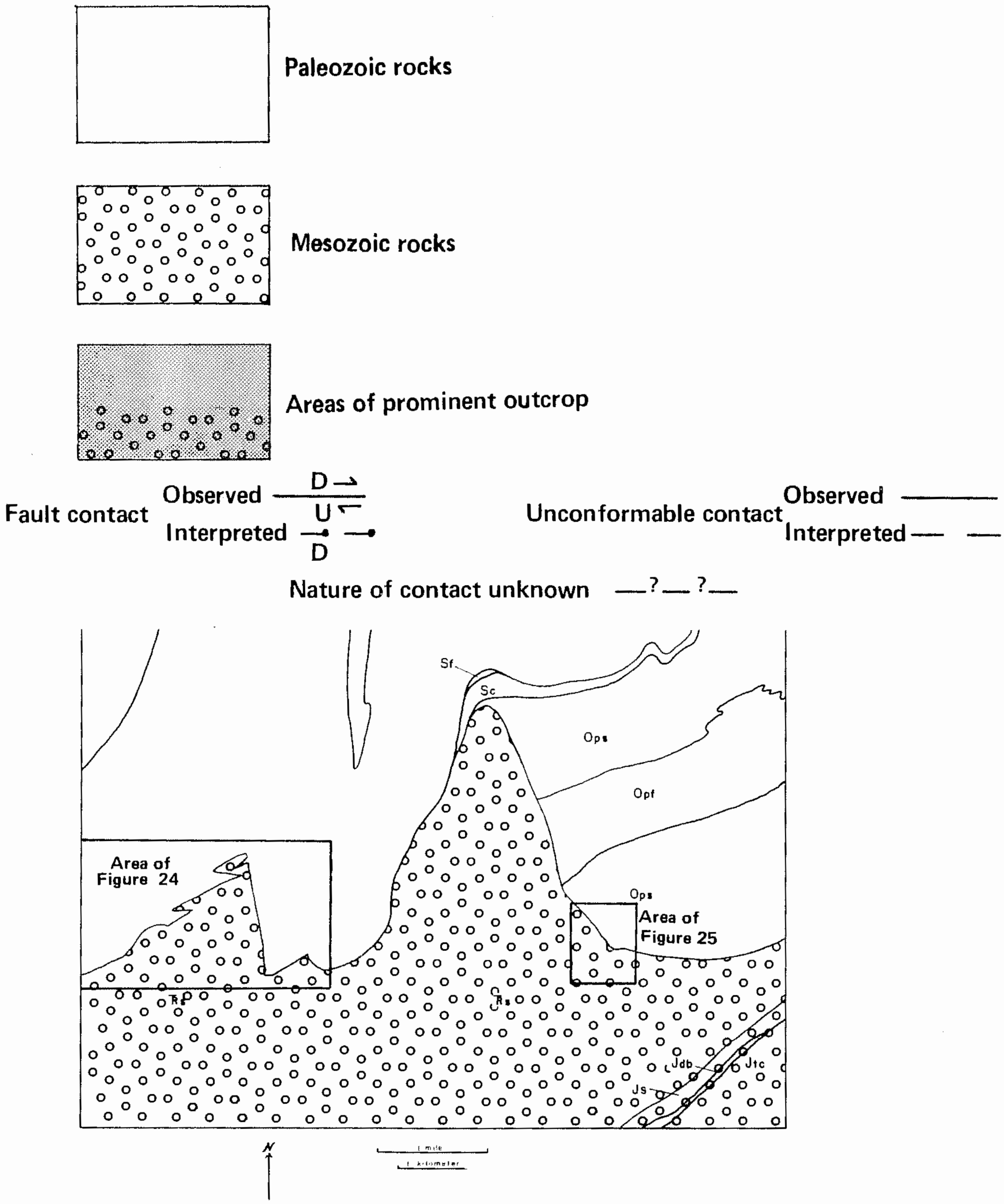


Figure 22: Horizontal stress directions in the Deerfield Basin after rotation about the strike of bedding (Goldstein, 1975).

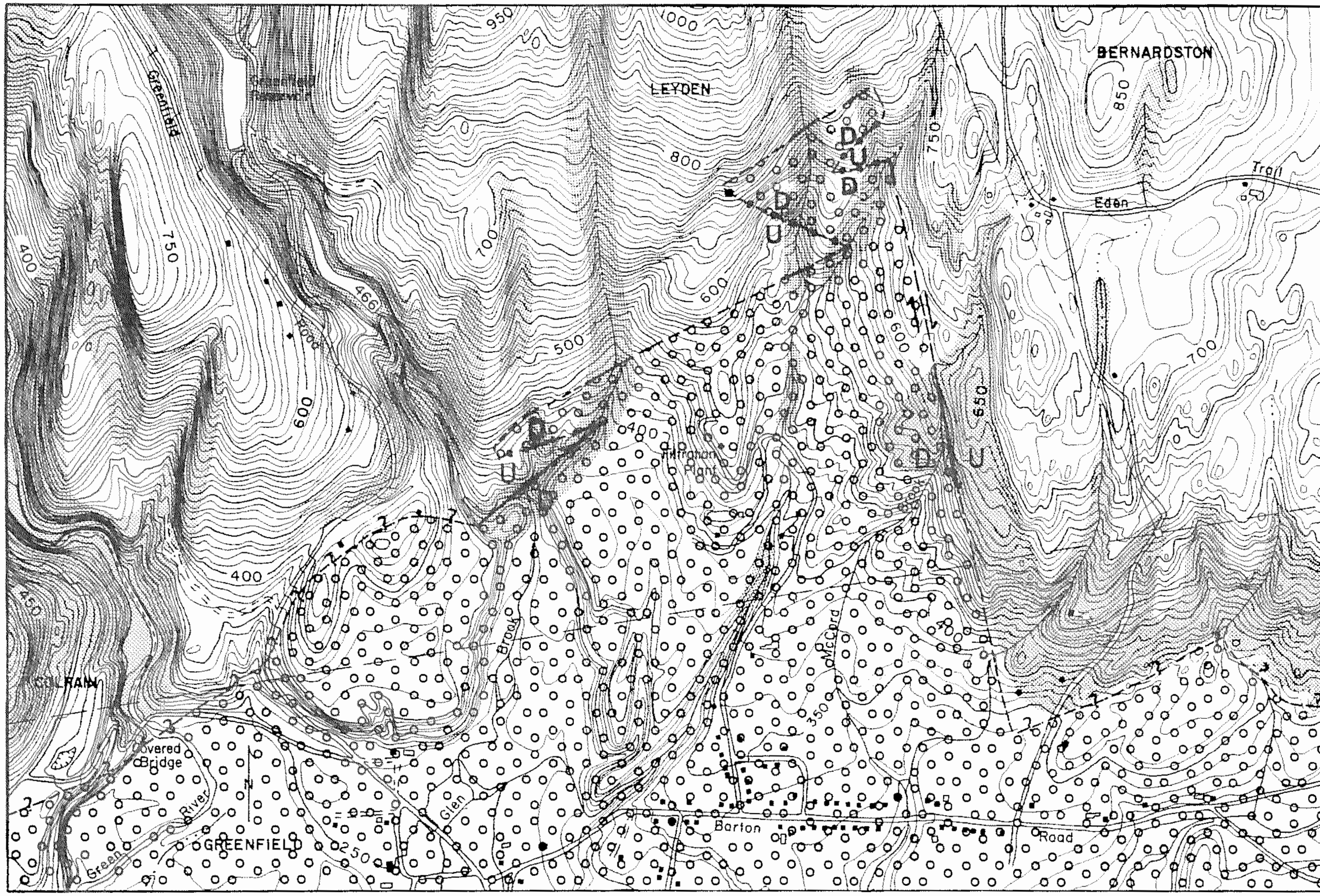
izontal and parallel to the graben axis during its formation. McHone (1978) has found a similar stress system for the Triassic -Jurassic through observation of dikes of this age throughout New England. In the case of dikes, σ_3 is taken to be approximately normal to their planar orientations but the σ_1 and σ_2 orientations are confined to lying only somewhere approximately parallel to the plane of the dike. Thus σ_1 may likely but not necessarily lie horizontally along the strike of the dikes, as a vertical orientation is also possible.

Paleozoic - Mesozoic Boundary

The Paleozoic - Mesozoic boundary at the northern end of the Deerfield Basin was examined to determine the nature of the contact with respect to structural control and topography of the unconformity. Unfortunately, most of this boundary is covered by glacial deposits and recent aluvium, especially on the eastern side of the study area. However, in the area north of the Greenfield Filtration Plant and east of the Greenfield Reservoir (Figure 23) the Paleozoic - Mesozoic contact is exposed well enough to obtain some idea of its nature. A detailed look here shows the contact to be much more complex than was mapped by Balk (1956). This area displays a combination of northeast, northwest, and east-west striking faults creating a very irregular pattern at the boundary. The east-west faults are normal with the south side down. The northeast faults



Southern half of Bernardston quadrangle showing location of Figures 23 and 24 including legend for Figures 23 and 24.



0.5 Miles

0.5 Kilometers

Figure 23: Paleozoic - Mesozoic contact east of the Greenfield Reservoir.

show both left-lateral and normal motions and may indicate two periods of motion. The north-west faults show both normal and right-lateral motions also suggesting two period of motion. The strike-slip motions on these faults apparently post-date the initiation of normal motions. Although no definitive evidence was found in this area, Goldstein (1975) found similar strike-slip faulting cutting the Jurassic Deerfield Basalt and considered these to be late in the structural development of the basin. On the east side of the Bernardston reentrant in the Mesozoic basin the contact is exposed to the east of Hoeshop Road and southeast of Otter Pond (Figure 24). The nature of the contact in this area also indicates the irregular appearance observed on the west side of the study area. Thus, where detail is available it appears that the Paleozoic - Mesozoic contact at the northern end of the Deerfield Basin is much more irregular than existing regional maps indicate.

These irregularities may reflect the influence of the pre-existing anisotropies of the schistosity, intensified by the presence of large quartz veins, and the dominant east-west joint system in the Paleozoic rocks. Although these anisotropies pre-date the Mesozoic stresses and the resulting fractures in the Mesozoic rocks, they are probably an important factor in the orientation and development of the Connecticut Valley Mesozoic Basin. This tectonic inheritance mechanism of Mesozoic basin formation, being controlled by the pre-

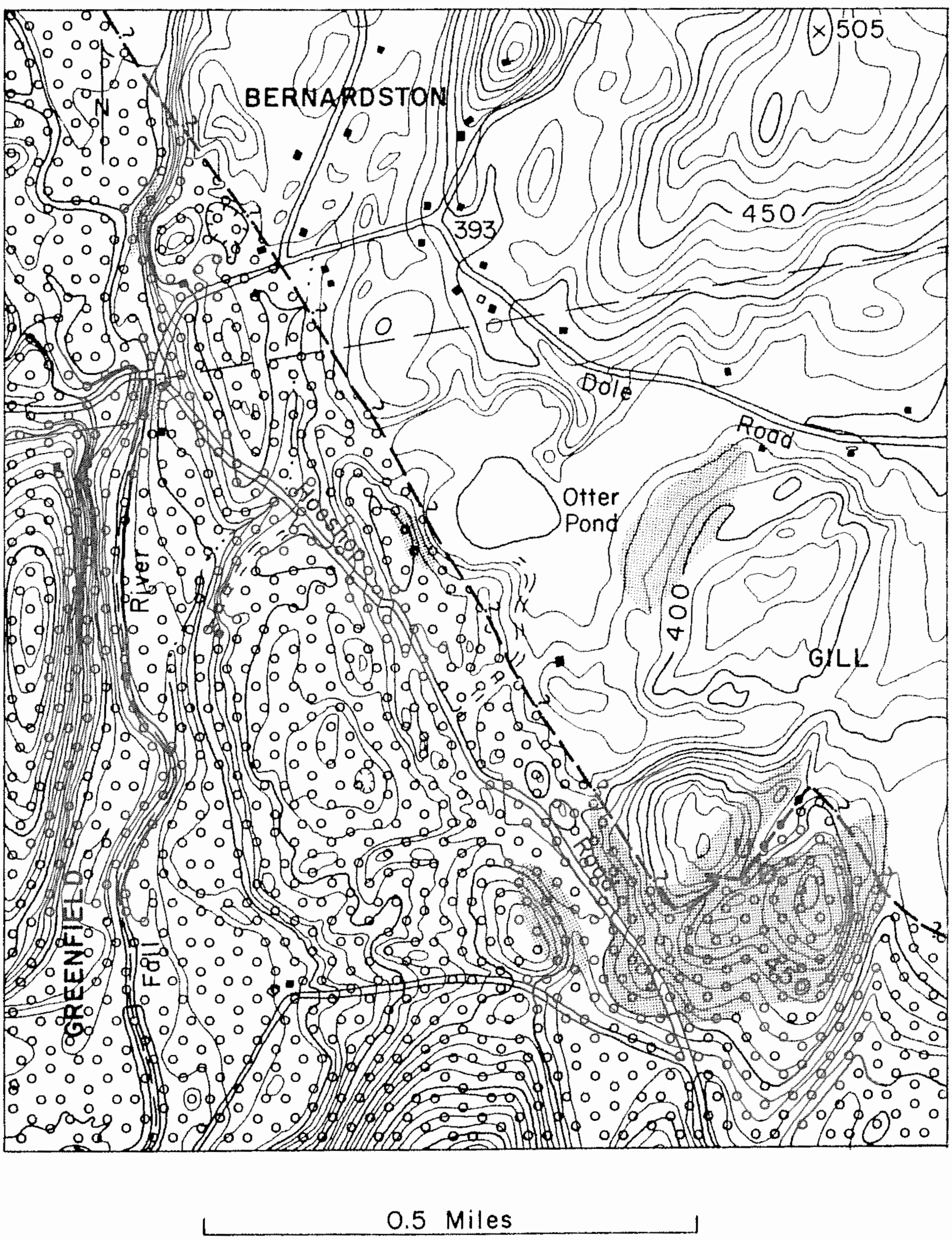


Figure 24: Paleozoic - Mesozoic contact south of Otter Pond.

existing structural fabric, is an old idea. It was suggested as early as 1907 by Lewis and since by many others, most recently Lindholm (1978). Data assembled by Lindholm from pre-existing maps of North American Triassic - Jurassic basins show that the border faults are parallel to the underlying foliation or structural grain in both strike and dip suggesting the strong control of these pre-existing planes of weakness in the development of basins. However, this does not hold true for many areas along the Connecticut Valley border fault.

Brittle Fracture Traverse

Faulting along the Massachusetts - Vermont border traverse (Figure 25) reflects the same north-south and east-west trending systems displayed by the major joint systems. In the eastern portion of the traverse dip-slip normal faulting dominates whereas, in the western and central portions scanty fault data combined with a wide variety of motion types and orientations on the few faults observed, prohibit any general fault classifications for these areas. The Taconic (Figure 26) and Berkshire (Figure 27) stations appear to be relatively unfaulted at the outcrop scale. The faults found in these areas are too sparse to fall into any pattern of orientation or movement sense. Faulting in the Paleozoic rocks of the Connecticut Valley area (Figure 28) is dominated by dip-slip normal motions on mostly northeast striking planes with secondary sets of normal faults

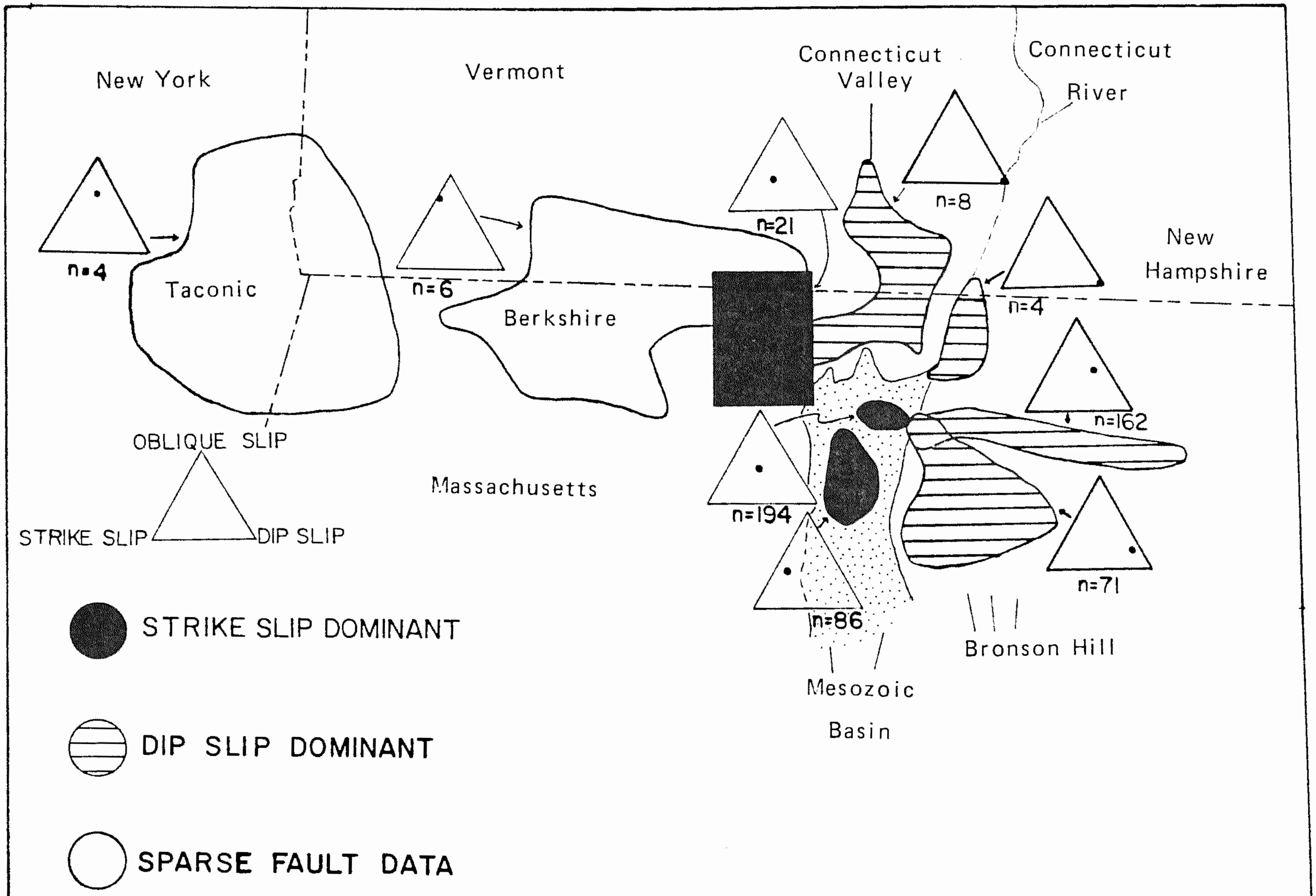


Figure 25: Fault type domains.

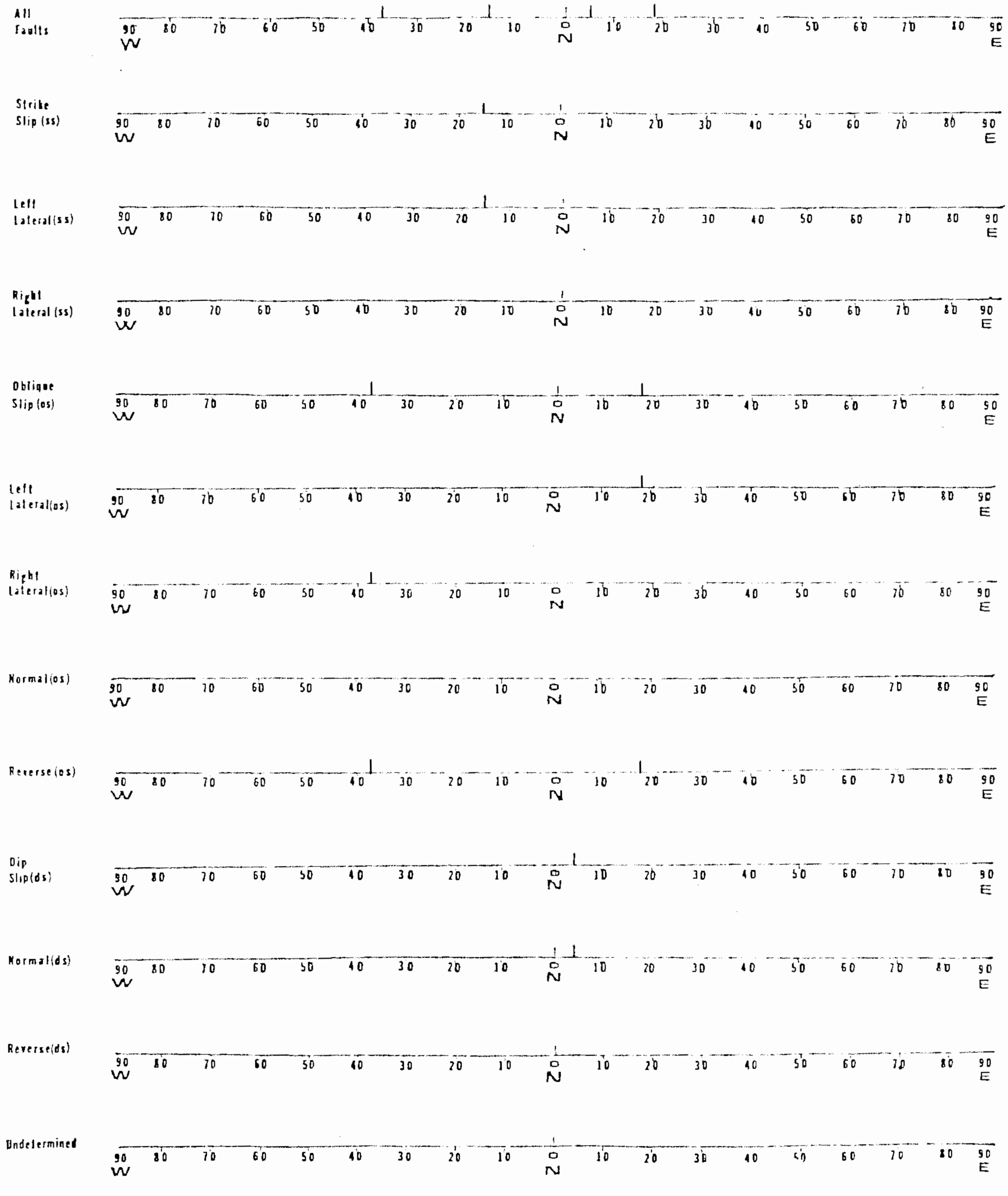


Figure 26: 4 faults observed at 21 fracture stations in the Taconic Mountains.

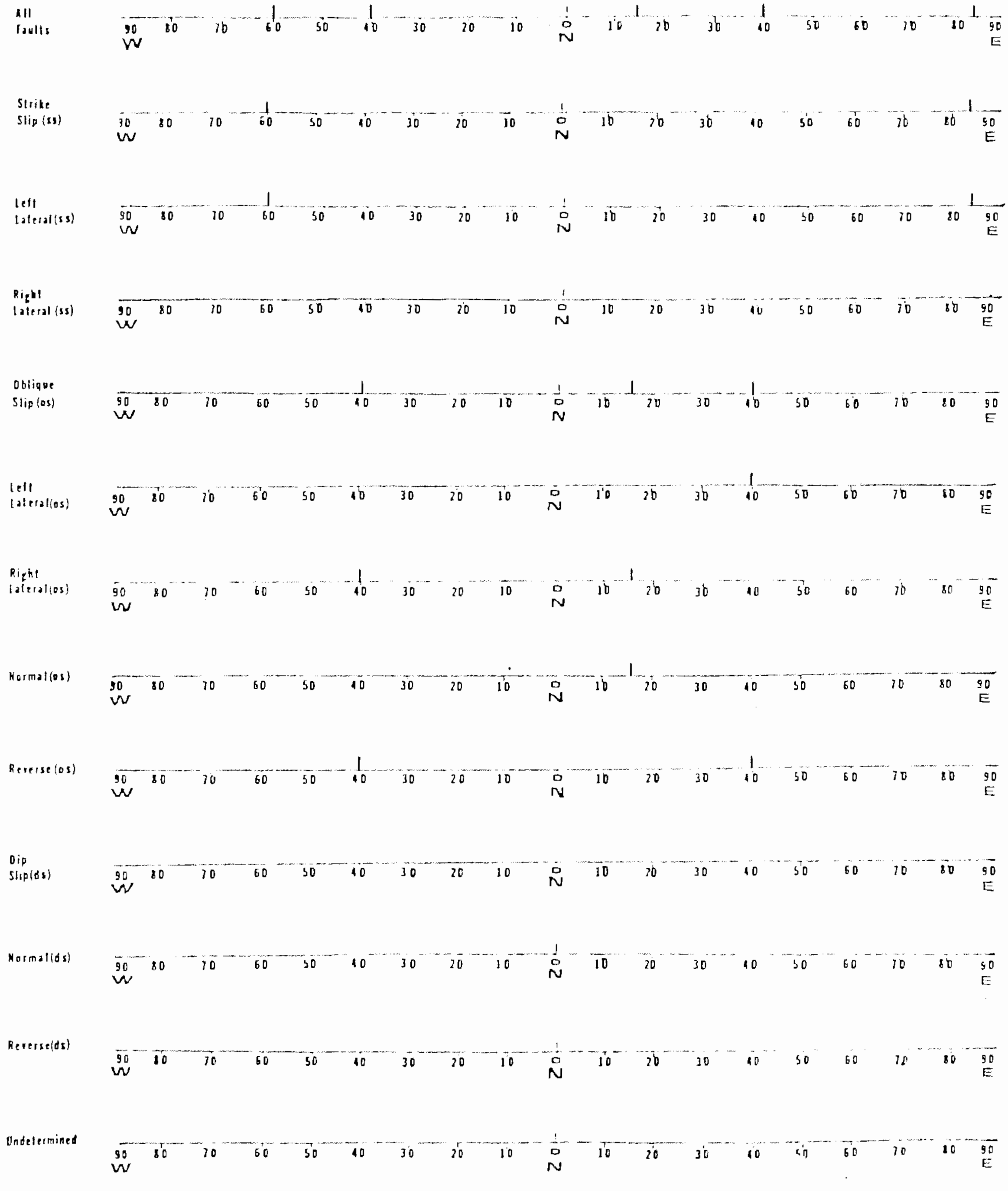


Figure 27: 5 faults observed at 29 fracture stations in the Berkshires.

trending nearly north-south and northwest. These faults are interpreted as being related to the development of the Mesozoic Connecticut Valley Basin to the south. Faulting in the Bronson Hill area (Figure 29) is similar to the Connecticut Valley area with northeast trending normal faults predominating and a secondary north-south trending set also evident. Silverman (1976) indicates the development of N 80^{o}E fault set on the east limb of the Bronson Hill Anticlinorium in the area of Athol, Massachusetts. The faults of this area are dominantly dip-slip and oblique-slip with the normal component predominant in both types.

Relative number of faults by motion type for each area are shown on Figure 30.

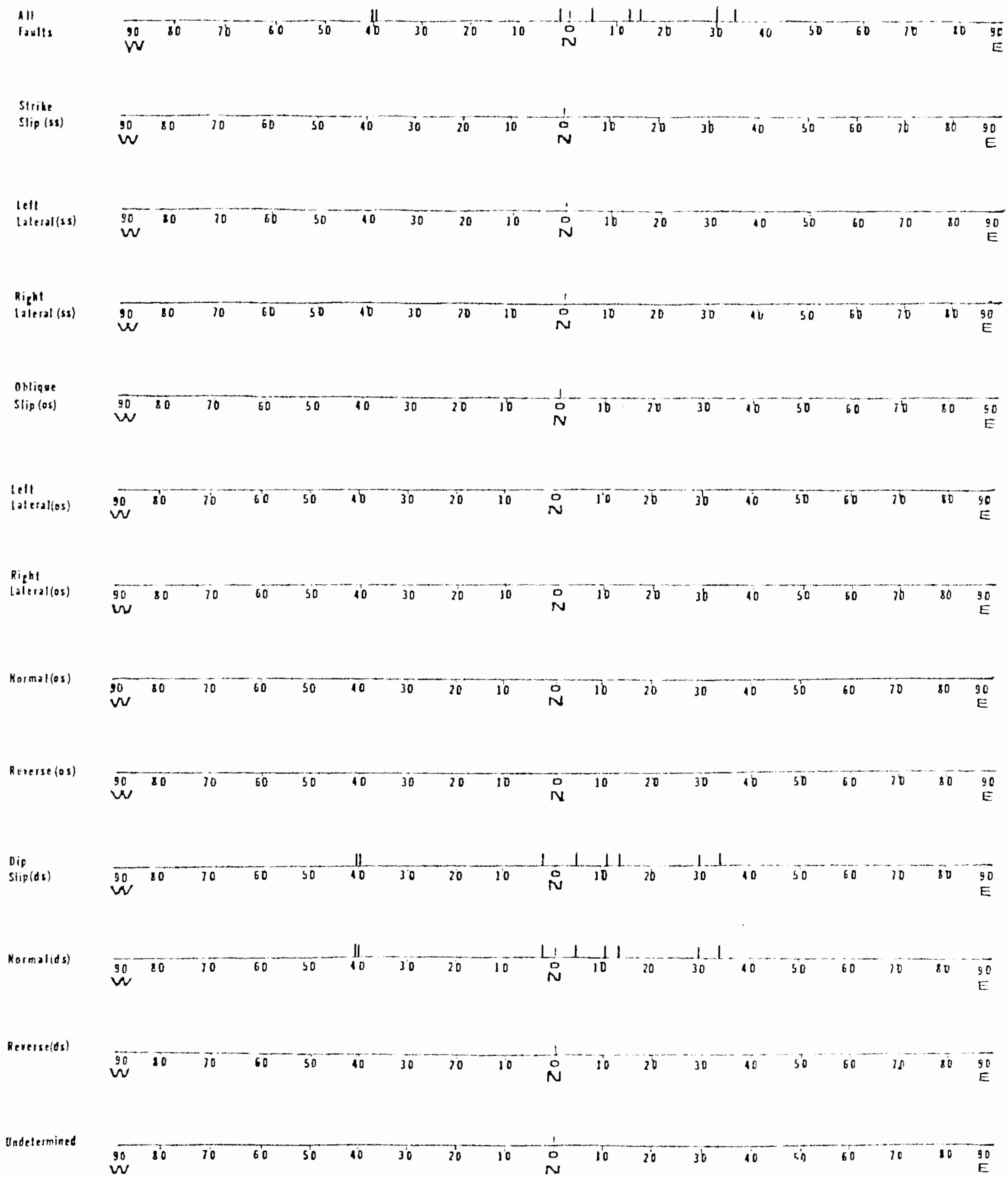


Figure 28: 8 faults observed at 16 fracture stations in the Connecticut Valley.

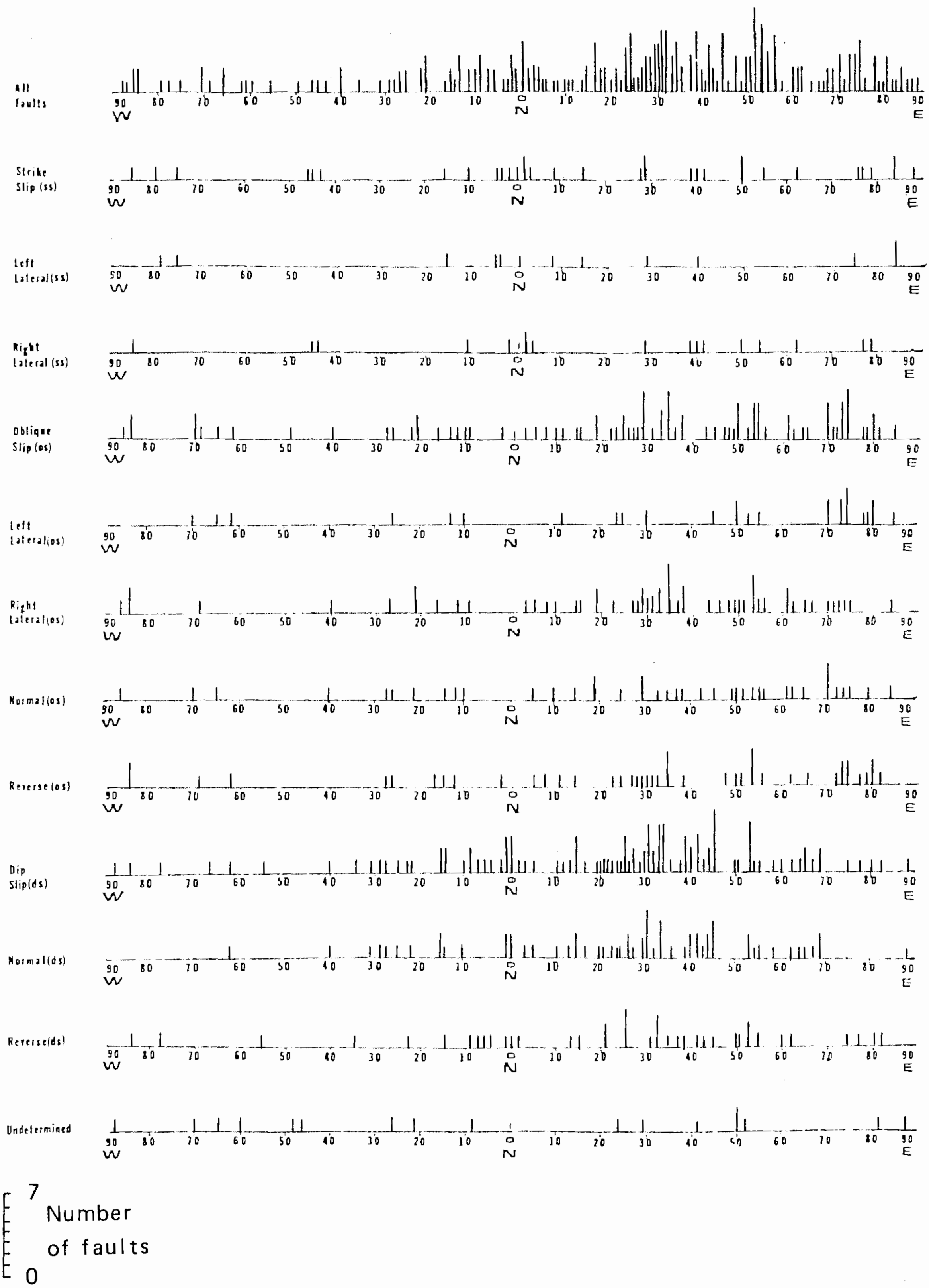
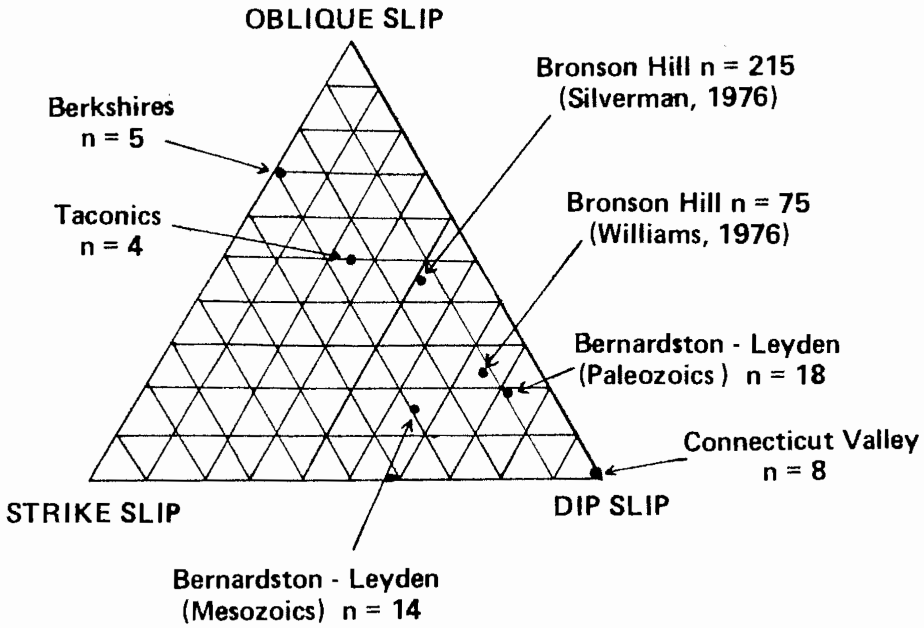


Figure 29: 290 faults observed at 42 fracture stations in the Bronson Hill Anticlinorium (Williams, 1976, Silverman, 1976).

Figure 30: Ternary diagram showing relative number of dip - slip, strike - slip, and oblique - slip faults for each area.



JOINT PATTERNS

Bernardston - Leyden Area

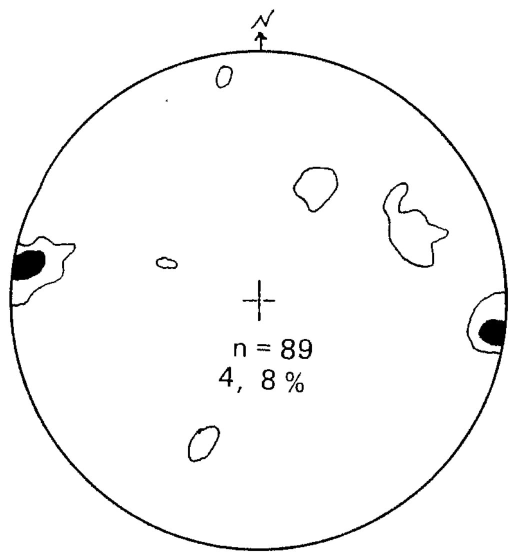
The Bernardston - Leyden study area lies within the Connecticut Valley - Gaspe Synclinorium region of the brittle fracture traverse along the Massachusetts - Vermont border. The fracture pattern in the Paleozoic rocks of this area corresponds well with the regional pattern defined by the traverse on the west side of the valley. The Mesozoic fracture pattern resembles that found by Goldstein (1975) to the south. Mesozoic and later stresses appear to have been dissipated along the pre-existing anisotropies of the regional north-south schistosity and the prominent N 80^{o}E to east-west fracture set.

Jointing in the Palezoic rocks of the study area (Figure 31b) is dominated by a strong N 80^{o}E to east-west, steeply dipping joint set. This joint set is preferentially oriented perpendicular to the prominent schistosity. Locally, where the schistosity has been folded out of the regional north-south orientation, a joint set parallel to the regional north-south trend is developed reflecting the strong influence of the regional tectonic grain and its control over the brittle fracture pattern of this area. Sorting of joints by mineralization, surface characteristics, or size showed no significant deviation from the total joint pattern.

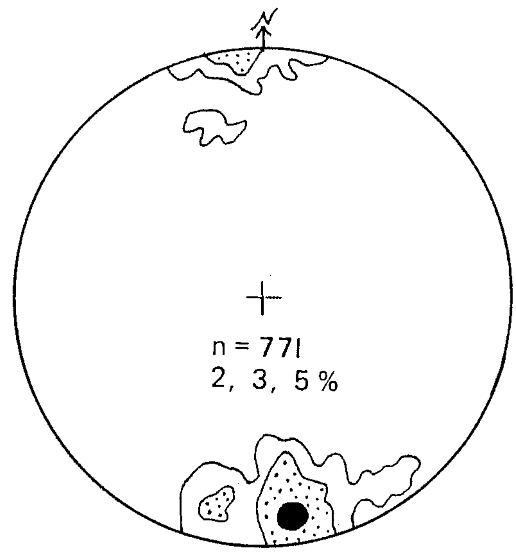
Jointing in the Mesozoic rocks of the study area (Figure 31a) is more complex than in the Paleozoic rocks. These rocks are dominated by a strong, steeply dipping, N 10°E set with a moderately dipping N 20° - 30°W set also well represented. The N 80°E to east-west set so strongly developed in the Palezoic rocks of the area is only sparsely represented at the Mesozoic stations sampled in this study but appears to be well developed in the Turner Falls area to the south (Goldstein, 1975). The strong N 10°E set may be inherited from the underlying regional schistosity. The N 80°E set, where developed, may be a result of vertical propagation of the Paleozoic fracture set of that orientation and/or younger deformation controlled by the Paleozoic grain producing jointing normal to that grain. The N 20° - 30°W set appears to be confined to the Mesozoic rocks and therefore may well represent a Mesozoic fracture set in which the stresses were redistributed in the Paleozoic rocks by strong pre-existing anisotropies.

Brittle Fracture Traverse

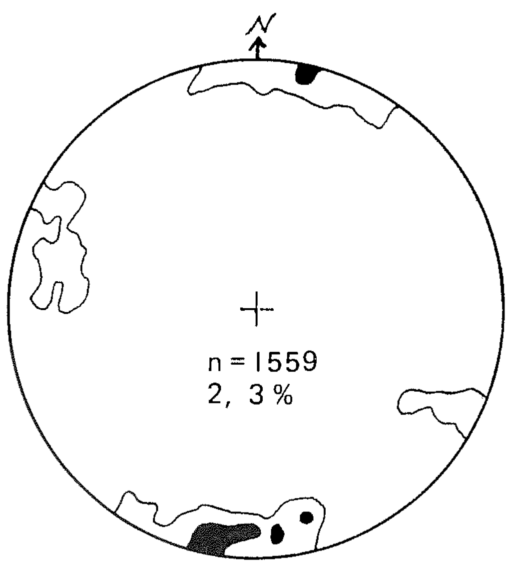
Jointing along the Massachusetts - Vermont border traverse is dominated by a roughly east-west set varying from about N 80°E to about N 70°W with a secondary set varying from north-south to about N 20°E, both steeply dipping (Figure 31C). Macrojoints (6 meters and over) across the area are dominated by a strong, steeply dipping,



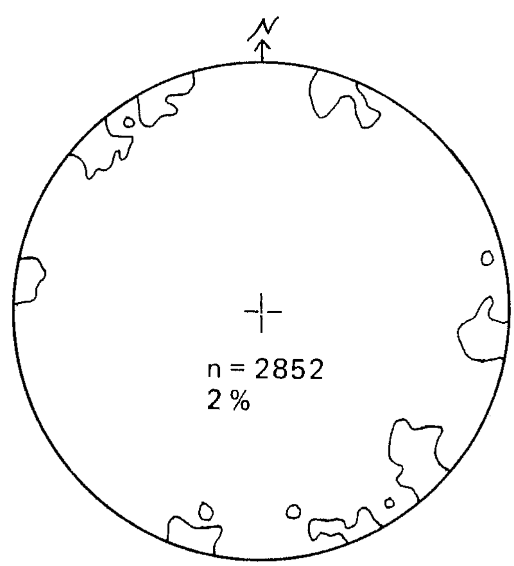
a. Joints in Mesozoic rocks, Bernardston-Leyden area.



b. Joints in Paleozoic rocks, Bernardston-Leyden area.



c. All joints brittle fracture traverse.



d. All joints Bronson Hill Anticlinorium (Williams, Silverman, 1976).

Figure 31: Poles to joints (contours in % per 1 % area).

east-west trending set. This trend is perpendicular to the regional foliation and structural grain over most of the area. In the Taconics jointing is dominated by an east-west trending set with a N 10^o -30^oE set and a N 70^oW set also evident over most of the area (Figure 32 and 33). Jointing in the Berkshire area (Figure 32) is dominated by this same east-west set which virtually disappears in the Heath quadrangle and in the south-central portion of the Wilmington quadrangle, where a strong N 50^oW set dominates (Figure 34). This N 50^oW set continues westward from here to the Taconics. West of the Heath - Wilmington area, where both sets are well developed, the east-west set again domiantes. The same east-west joint set is evident in the Connecticut Valley - Gaspe Synclinorium area of this traverse varying from N 80^oE to N 80^oW (Figures 32, 35 and 37). This set is not as well developed on the eastern side of the valley where a N 20^o - 30^oE trending, steeply dipping set is predominant.

Jointing in the Bronson Hill Anticlinorium area is quite complex (Figures 31d and 37). Unlike most of the region, there appears to be little or no correlation of joint maxima from station to station or in the overall pattern across the area. Nonetheless, at single stations, there is commonly a strong joint maximum. These data were sorted by rock type, mineralization, surface characteristics, and size without resolving any distinctive pattern for this area. Williams (1976) and Silverman (1976) in their fracture studies east and south of this

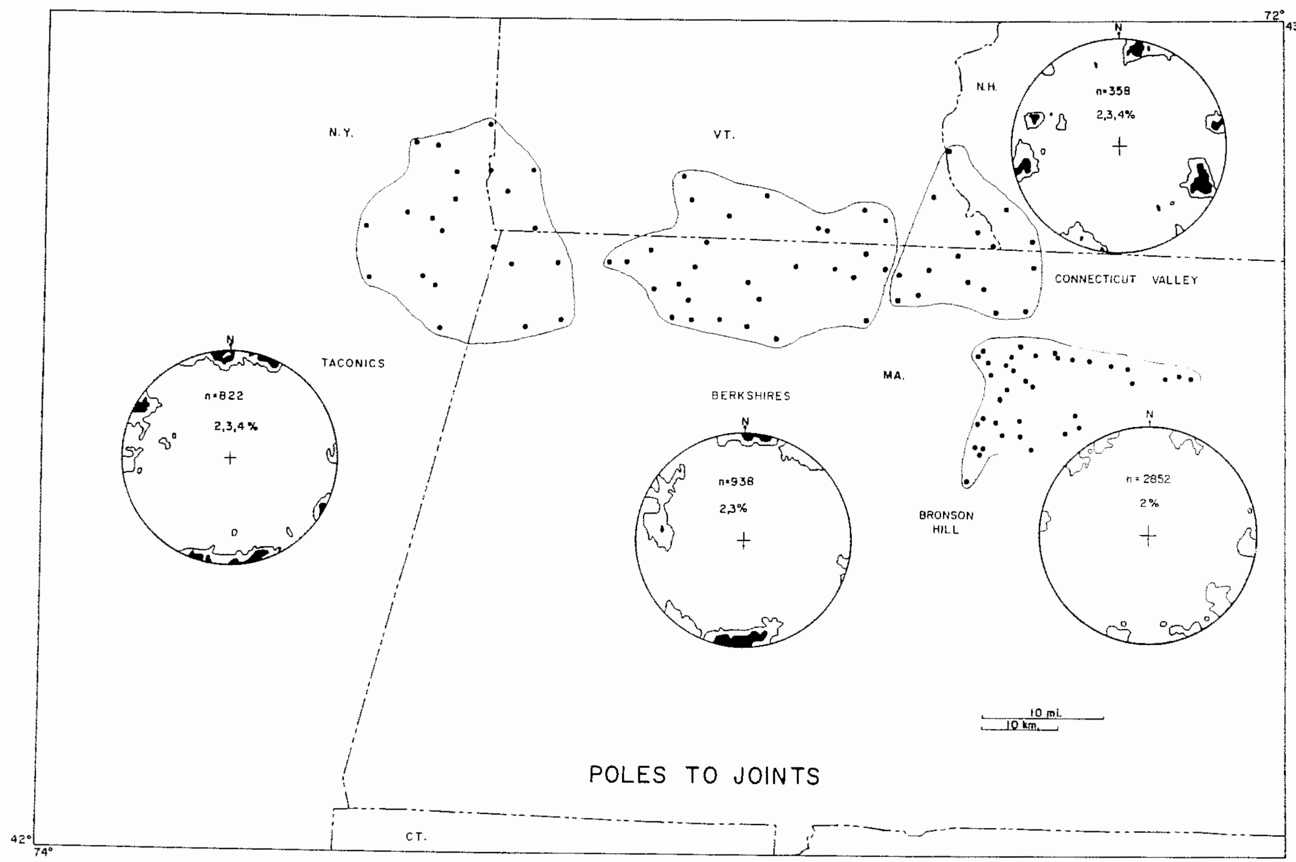


Figure 32: Poles to joints for each region of the Massachusetts - Vermont border traverse. (contours in % per 1% area)

a.	b.	c.	d.	e.	f.	g.	h.	i.
j.	k.	l.	m.	n.	o.	p.	q.	r.

a. Grafton
b. North Pownal
c. Pownal
d. Stamford
e. Wilmington - SW
f. Wilmington - SE
g. Brattleboro - SW
h. Brattleboro - SE
i. Keene - SW

j. Taborton
k. Berlin
l. Williamstown
m. North Adams
n. Rowe
o. Heath
p. Colrain
q. Bernardston
r. Northfield

Quadrangle locations for Figures 33, 34, and 35.

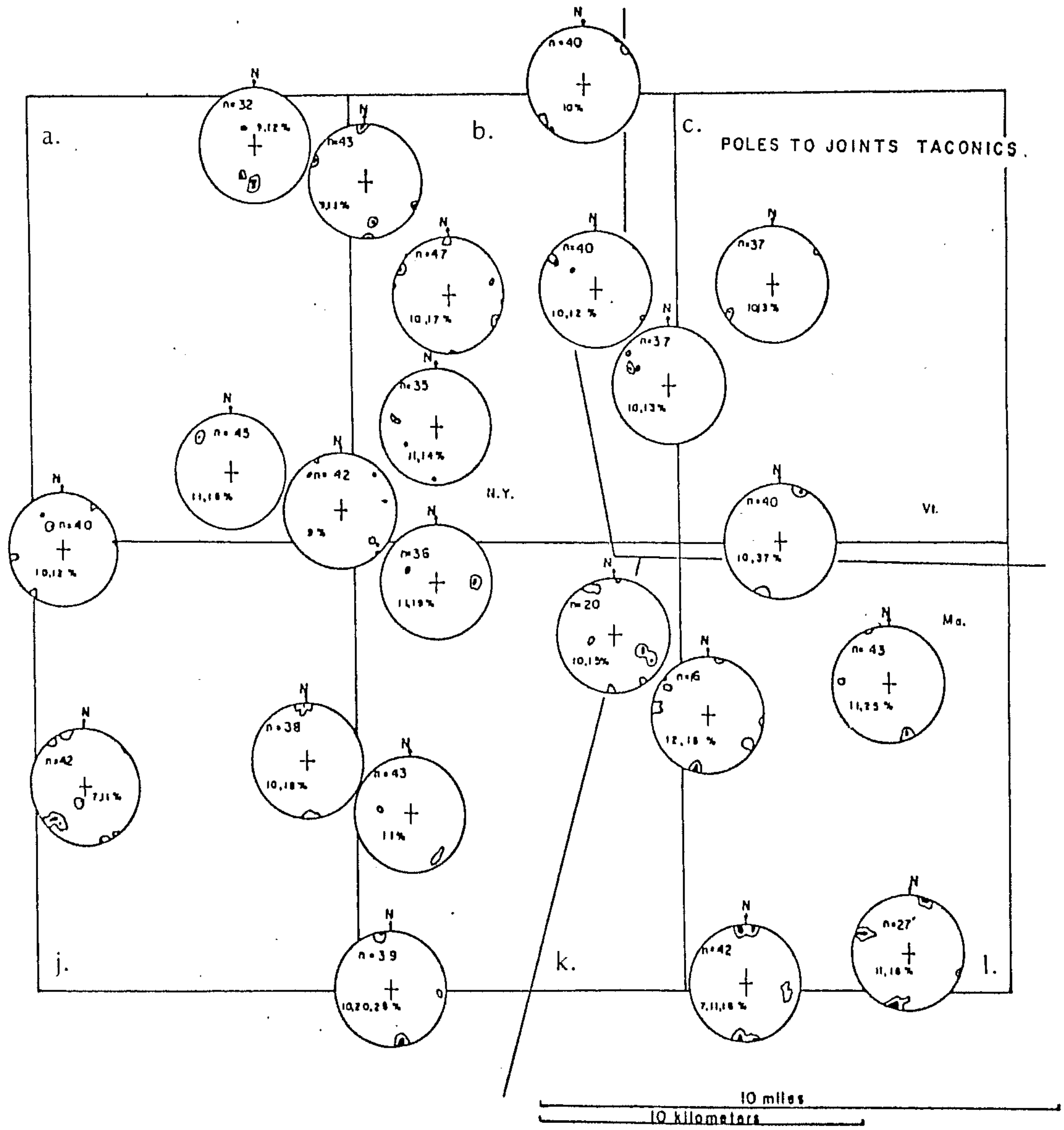


Figure 33: Poles to joints for each station in the Taconic region. (contours in % per 1% area)

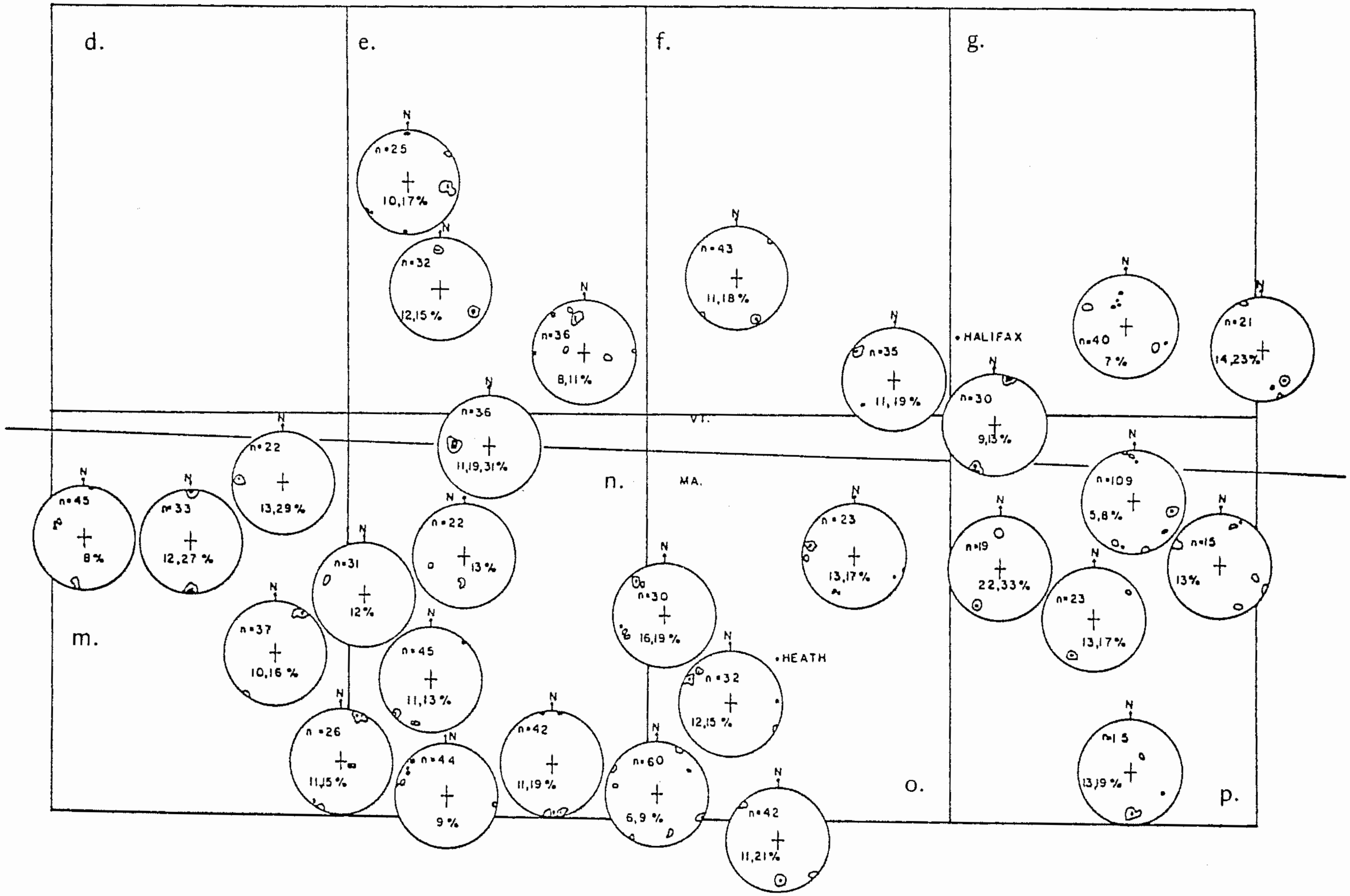


Figure 34: Poles to joints for each station in the Berkshire region. (contours in % per 1% area)

POLES TO JOINTS BERKSHIRES

10 miles

10 kilometers

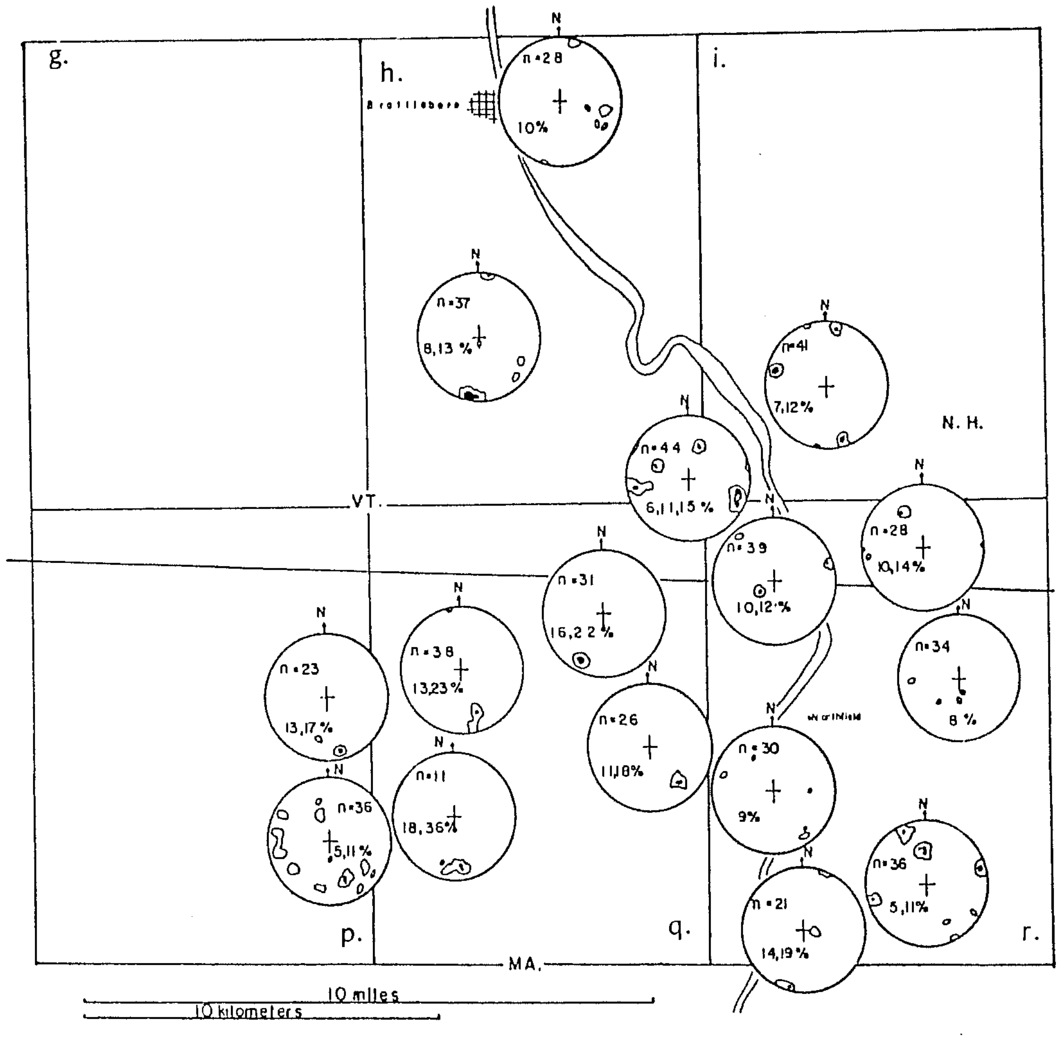


Figure 35: Poles to joints for each station in the Connecticut Valley region. (contours in % per 1% area)

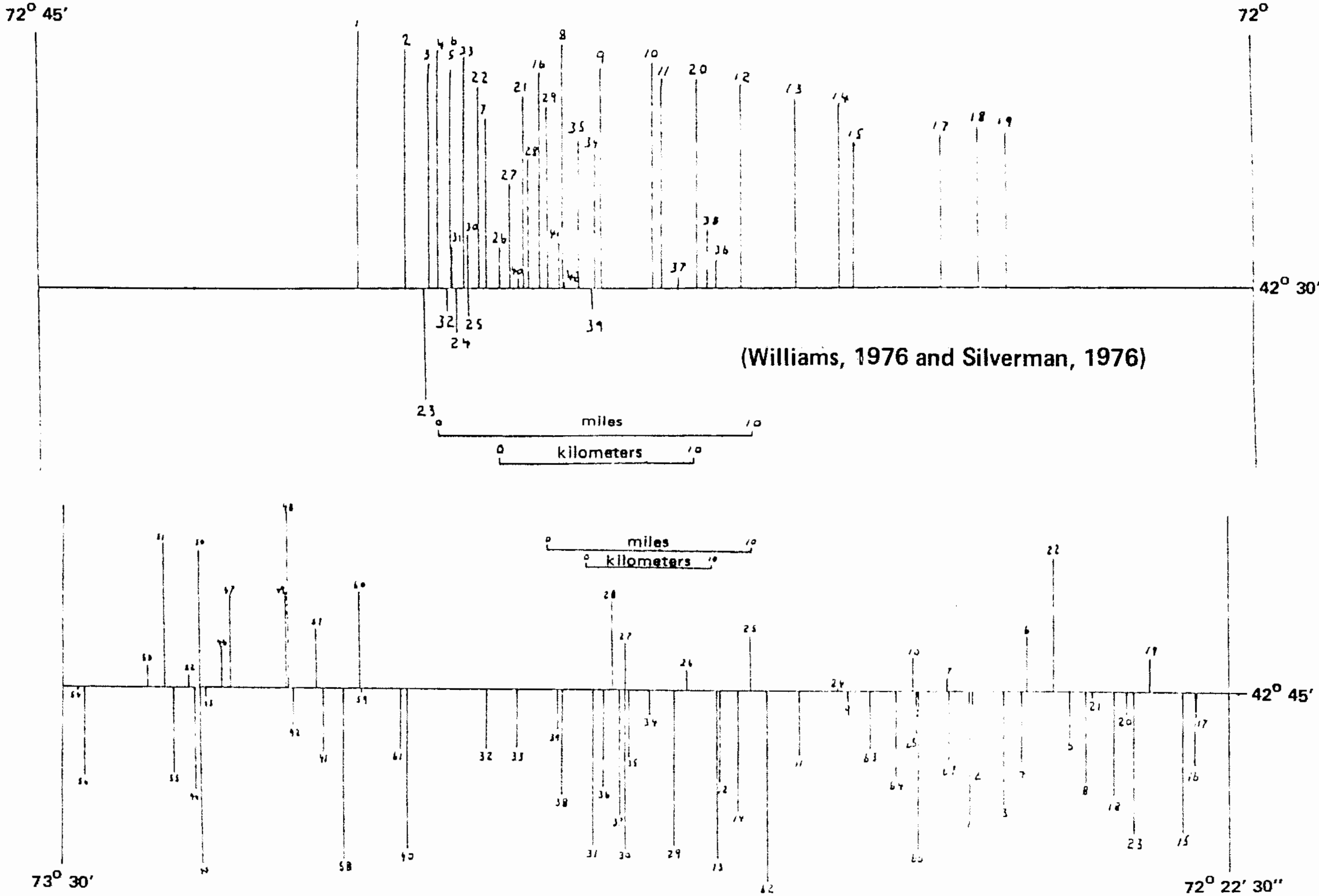


Figure 36: Station locations of brittle fracture traverses projected onto a single line. Williams, 1976 and Silverman, 1976 in Pelham Dome area (top), Massachusetts - Vermont border traverse of this study (bottom).

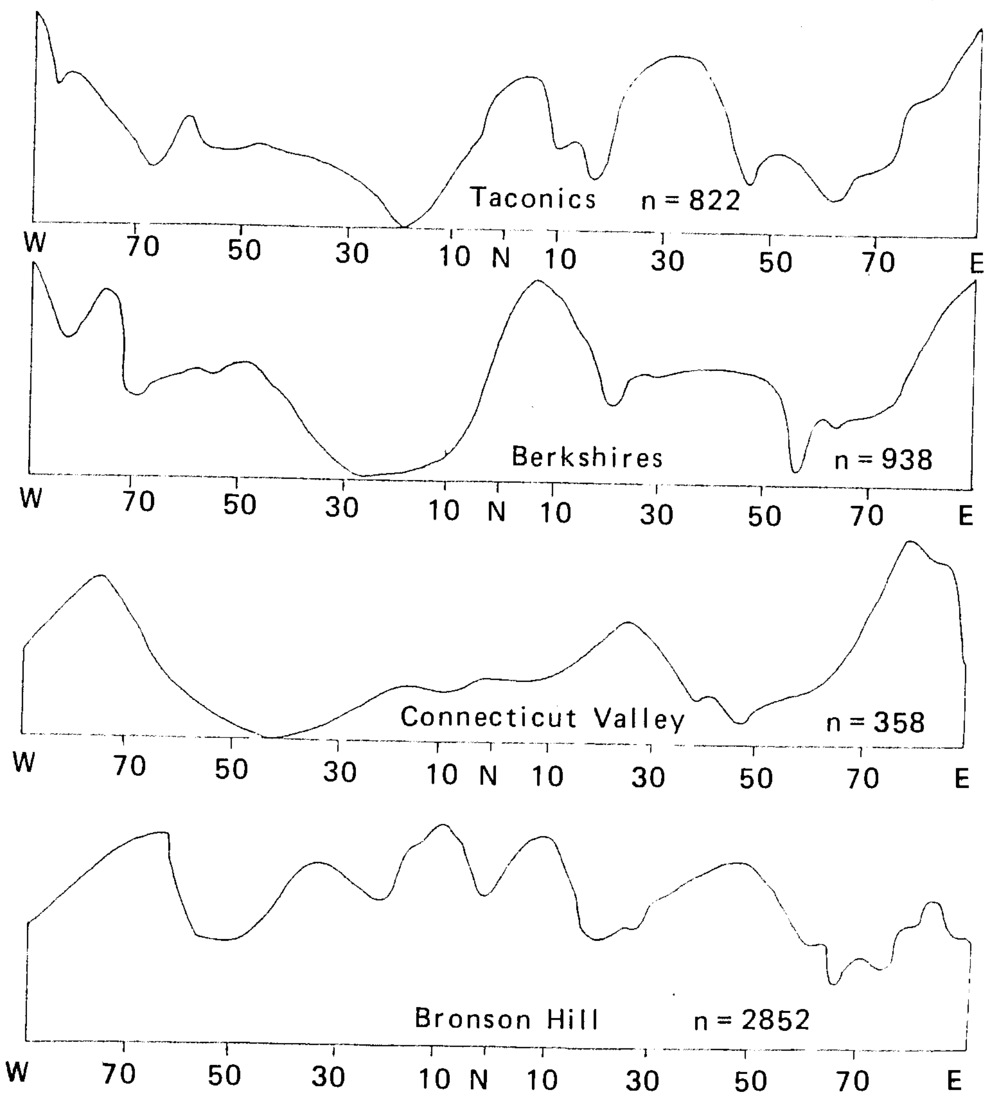


Figure 37: Joint azimuth histograms smoothed with a 5^{o} running average for each region.

traverse found little or no correlation of joint maxima from station to station in the northern portion of the Pelham Dome area of the Bronson Hill Anticlinorium. That area is anomolous in its joint diversity with respect to the more consistently oriented joint sets of the area west of the Connecticut Valley.

Brittle fracture data for each area are summarized in Table 3.

Table 3 Brittle Fracture Traverse Summary Table

Area	Taconics	Berkshires	Connecticut Valley	Bronson Hill
Joint Maxima	east-west N 10° - 30° E N 70° W	east-west N 50° W	N 20° - 30° E N 80° W N 10° W	Nothing Predominant
Fault Maxima	Insufficent Data	Insufficent Data	N 30° E north-south	N 30° - 35° E north-south N 80° E
Qtz. Vein Maxima	N 50° W north-south east-west	N 10° E	north-south	north-south

LINEAMENTS

Lineaments from satellite imagery (Figure 38) and shadowed relief maps (Figure 39) were analysed and compared to the observed fracture pattern along the Massachusetts - Vermont border.

Satellite lineaments across the region are dominated by north-south trends which probably reflect the structural grain throughout the area. Other trends are evident which correlate with the fracture trends in some areas but no good correlations can be made in most cases.

Lineaments from shadowed relief maps generally show better agreement with the fracture pattern observed across the area. However, as well as agreeing with the fracture patterns, other prominent lineament maxima are represented which probably do not relate to structural trends.

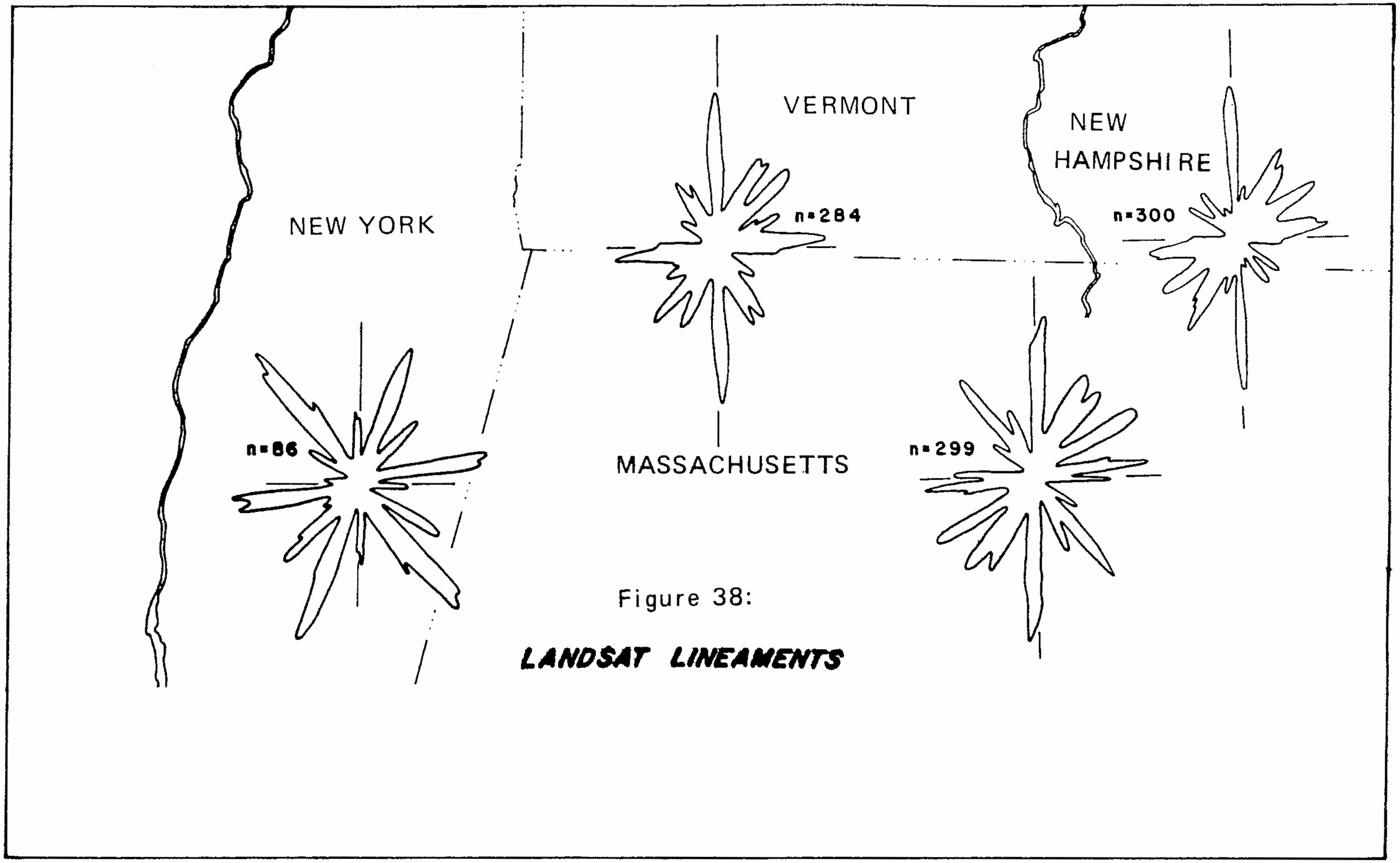
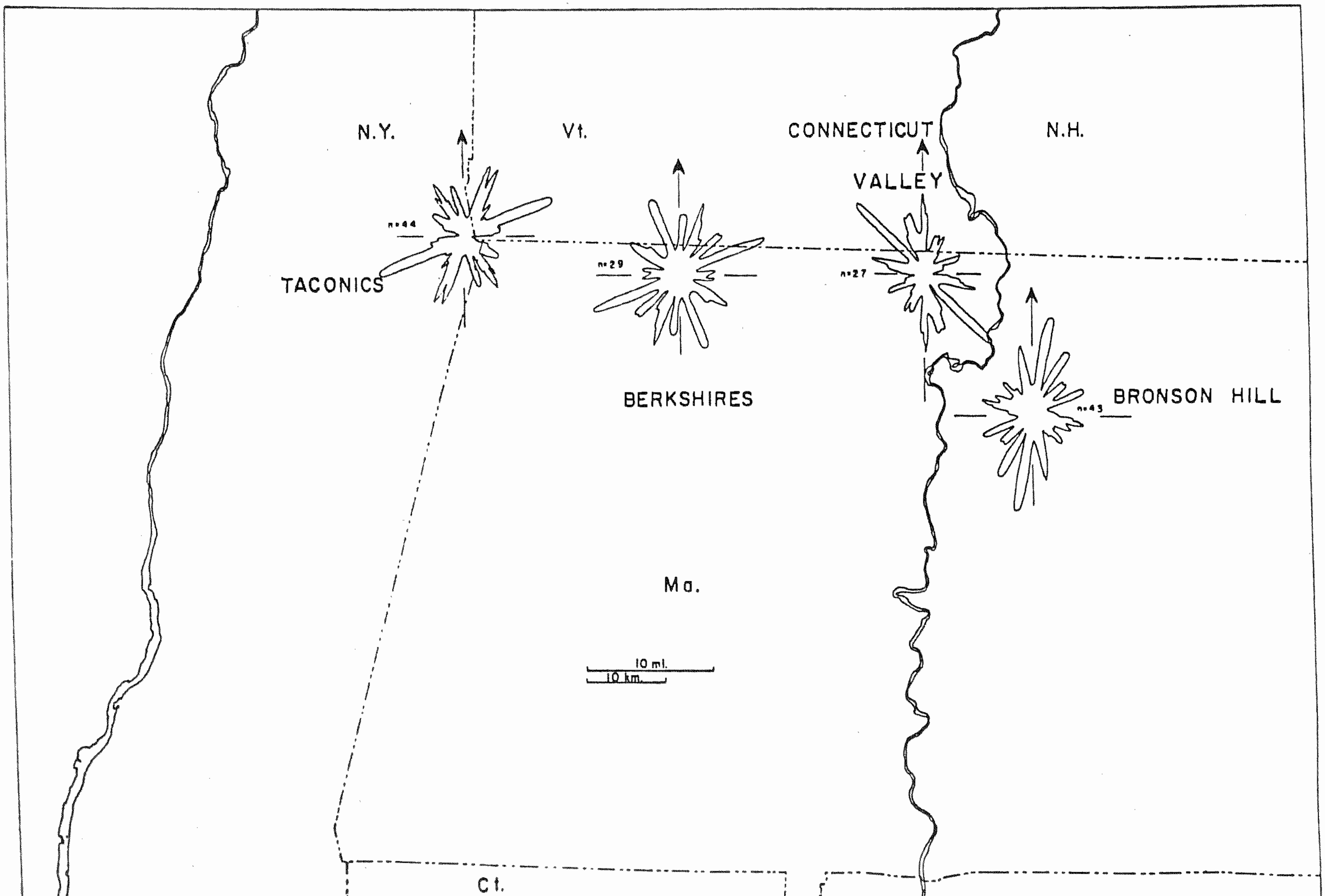


Figure 38:

LANDSAT LINEAMENTS

Figure 39: SHADOWED RELIEF MAP LINEAMENTS



SUMMARY

Folding

At least seven sets of folds are present in the study area but these probably represent only three distinct episodes of deformation. Three sets of isoclinal folds are probably related to nappe formation and possibly to pre-nappe stresses and post-nappe back-folding during the Acadian orogeny. Two sets of open folds seem to be related to dome emplacement during the peak of metamorphism toward the end of the Acadian orogeny. Two sets of kink bands (discussed below) are superimposed on the earlier Acadian folds.

Dikes and Quartz Veins

Granitic dikes and most of the quartz veins lie parallel to the prominant schistosity of the area. Their metamorphism and tight folding suggest they formed early in the structural history of the region. Later quartz veins, of quite variable orientation, are little deformed and cut many of the earlier features.

Kink Bands

Stress trajectories determined from kink bands indicate a mean σ_1 orientation of N 56^{o}E, 32^{o} to the northeast. These kink bands seem to represent the transition from ductile to brittle deformation which probably occured in late Permian or early Triassic time. Their stress orientation differs from the east-west orientation of mid-Paleozoic compression and may reflect a change in stress orientation of the later Paleozoic Alleghenian orogeny or stresses associated with the initiation of Mesozoic rifting to the south. If they are associated with the Mesozoic deformation, they may have important implications for the regional tectonics of basin formation.

Faults

Normal faulting associated with the formation of the Mesozoic Basin, followed by strike-slip motions, results in an irregular appearance of the Paleozoic - Mesozoic boundary. Normal faults are primarily northeast striking with some striking east-west. This dip-slip domain is consistent with the Bronson Hill area to the east but is bounded on the south and west by a predominantly strike-slip domain (Figure 25). The irregular map pattern of the Paleozoic - Mesozoic contact suggests that it has been controlled by small block motions rather than by one master fault.

Strike-slip motions in the Bernardston - Leyden area are permissive of conjugate interpretations similar to those of Goldstein (1975) to the south. These motions suggest a possible σ_1 orientation of about N 20°E during some later phase in the development of the Mesozoic basin or later, certainly after initial basin formation.

Joints

This study (combined with the Williams, 1976 and Silverman, 1976 data) representes the longest traverse of detailed fracture station data in New England to date, including 5830 joints measured at 143 stations. Jointing in the Bronson Hill Anticlinorium differs from the remainder of the traverse. In this region no overall orientation dominates the fracture pattern (Figure 31d). However, individual stations commonly exhibit strong maxima which may have nearly any orientation. An overall east-west trend dominates the joint pattern of the traverse (Figures 32 and 37) but is subdued at individual joint stations. This trend is most obvious in the macro-joints which are nearly all east-west trending and steeply dipping. This trend is lost in an area of the central Berkshires where a N 50°W set is dominant and in the Mesozoic rocks at the north end of the Deerfield Basin where a N 10°E trend dominates, possibly due to the influence of the underlying Paleozoic anisotropies.

Lineaments

No close correlation was observed between fracture station data and lineaments drawn from satelite imagery or shadowed relief maps. Specifically the lineaments here should not be arbitrarily considered as certain indicators of outcrop scale faulting.

Postulated Structural and Stress History

The postualted structural and stress history has been determined through analysis of folds, dikes, faults, and various radiometric dates. Much of this has been drawn from the existing literature and hopefully will be strengthened by future research.

The Devonian Acadian orogeny began with the intrusion of granitic bodies followed by east over west nappe formation with later backfolding of these nappes. The orientation of σ_1 during this time was more or less east-west. This recumbent folding was followed by dome emplacement during the peak of Acadian metamorphism with σ_1 still more or less east-west and σ_3 near vertical.

Kink bands most likely developed in Late Permian or Early Triassic time, possibly as the initiation of the continental "pull apart" between Africa and North America began to the south. σ_1

determined from these kinks indicates a N 56^{o}E direction of compression, plunging 32^{o} northeast. Triassic - Jurassic dikes throughout New England indicate that σ_3, if near horizontal, was oriented about N 60^{o}W suggesting a σ_1 orientation of about N 30^{o}E during the beginning of Mesozoic basin formation. Conjugate faults of probable Jurassic age indicate a σ_1 orientation of about N 20^{o}E as the basin continued developing. Late Jurassic dikes in northern New England indicate a σ_3 orientation of about north-south and possibly a σ_1 orientation east-west halting the formation of the Mesozoic Basin.

Cretaceous dikes associated with the White Mountain Magma Series indicate that σ_3 had rotated to about N 30^{o}E by that time with σ_1 possibly oriented about N 70^{o}W.

Contempory stress determinations indicate σ_1 is presently oriented roughly northeast in this area (Sbar and Sykes, 1973).

Suggestions for Further Study

During the research and writing of this paper several unsolved problems became evident, the solution of which could contribute to a better understanding of this area with respect to the tectonic history of the area and the development of the Mesozoic Connecticut Valley Basin.

1. A continuation of kink band stress determinations along the low-grade metamorphic axis of the Connecticut Valley - Gaspe Synclinorium to the north may help to determine the extent of the stress system at the time of kink development in the Bernardston - Leyden area.

2. A closer look at the Mesozoic rocks in the vicinity of the Paleozoic - Mesozoic contact with more extensive paleocurrent and paleoslope determinations may help to define the pre-Sugarloaf topography. Also, a better understanding of the nature of the unconformity may be gained from analysis of the areas of leaching and the source of the angular clasts at some of the fault contacts.

3. Analysis of drill hole data and use of geophysical methods could help to better define the Paleozoic - Mesozoic contact in areas of extensive glacial cover.

4. Fission track dating of minerals in the Paleozoic rocks would better define the timing of uncovering and put tighter constraints on the timing of kink band development, as well as giving some idea of the rate of uplift of this area.

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