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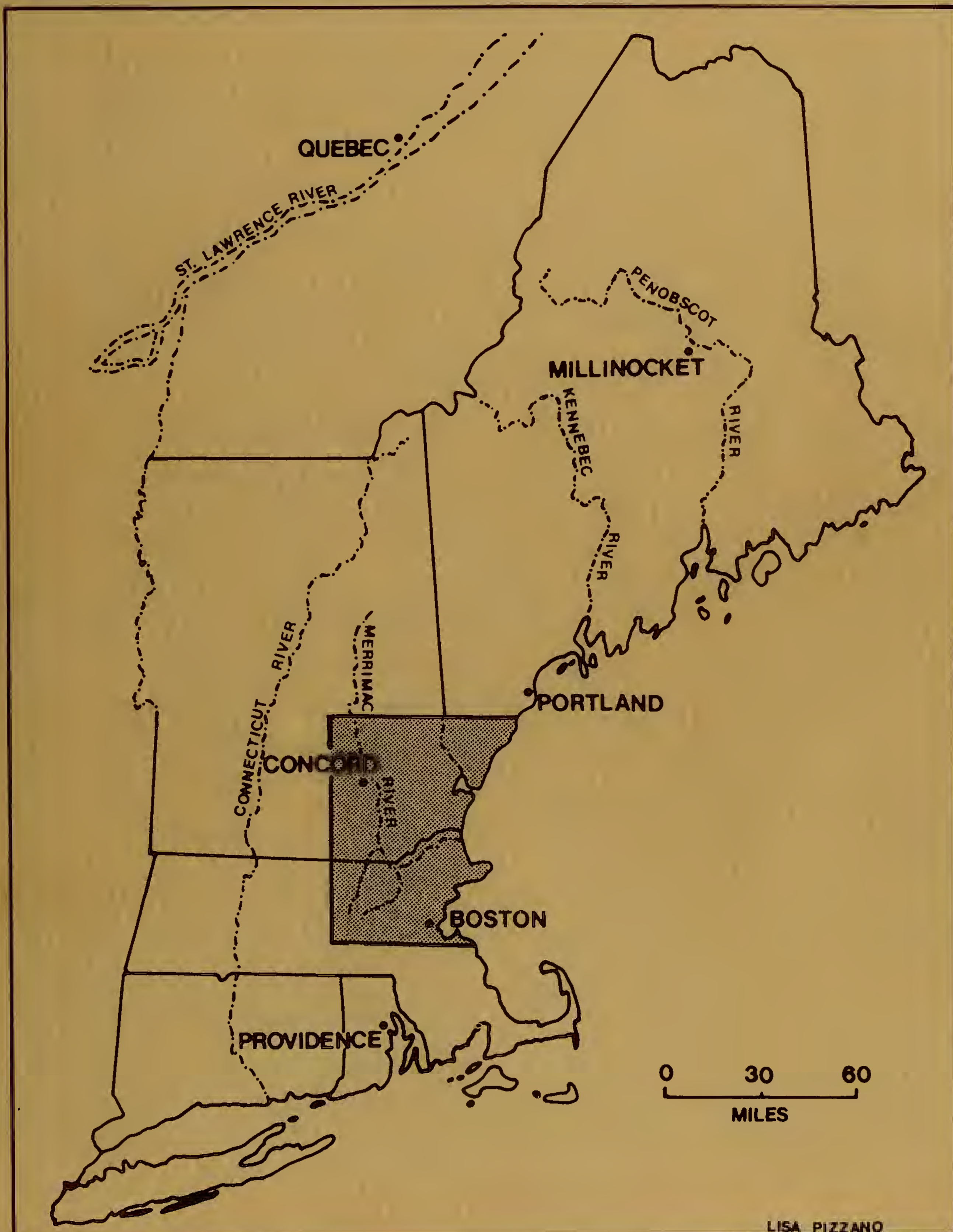
Geology of the Coastal Lowlands Boston, MA to Kennebunk, ME

76th ANNUAL NEW ENGLAND INTERCOLLEGIATE GEOLOGIC CONFERENCE, 1984



Edited by Lindley S. Hanson

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Department of Geological Sciences
Salem State College, Salem, MA

COASTAL LOWLANDS FROM BOSTON TO KENNEBUNK AND INLAND
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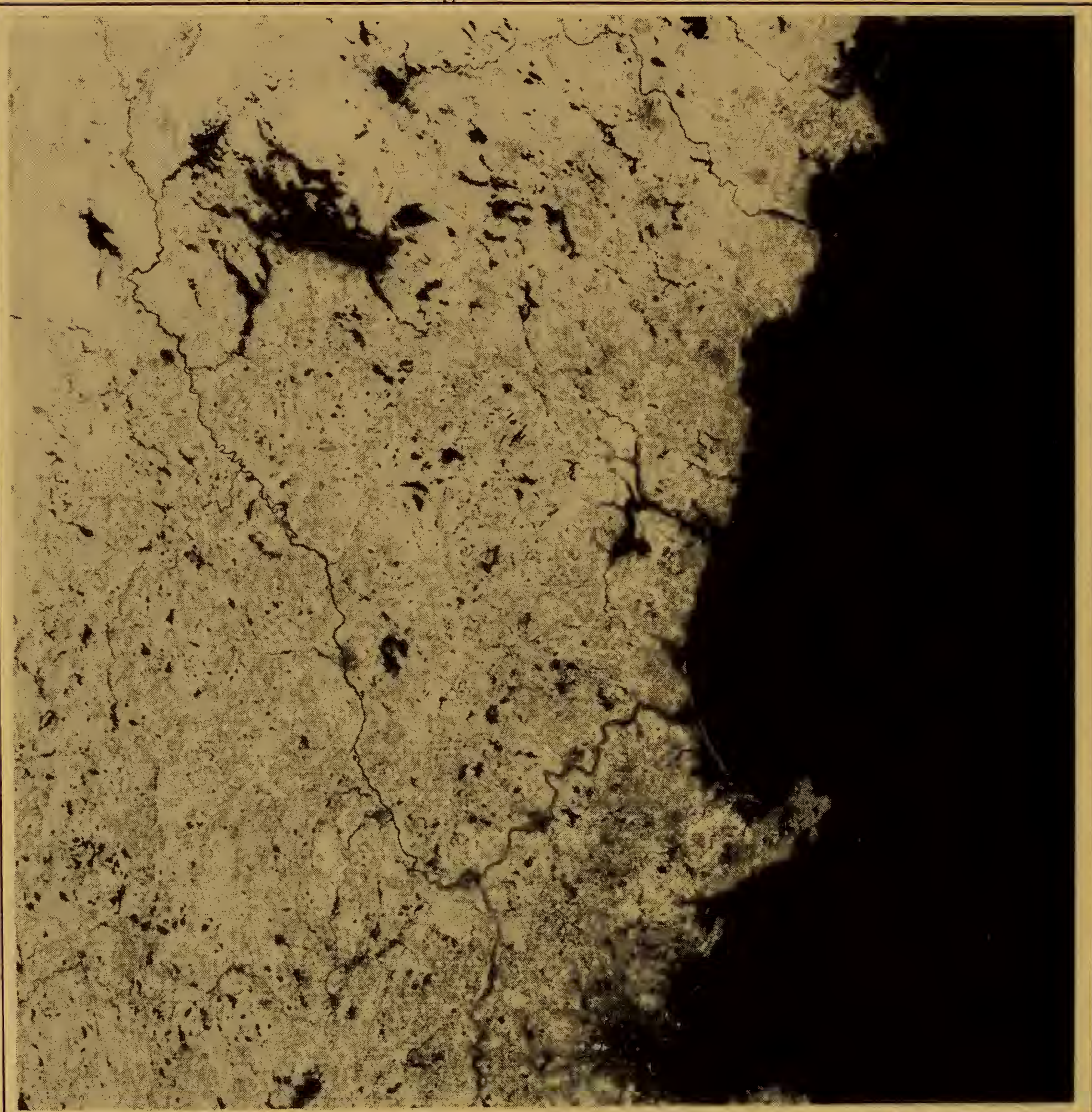

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GEOLOGY OF THE COASTAL LOWLANDS, BOSTON TO KENNEBUNK, MAINE

Edited by

Lindley S. Hanson
Department of Geological Sciences
Salem State College
Salem, Massachusetts 01970

76th ANNUAL MEETING
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

Danvers, Massachusetts

October 12, 13, 14, 1984

Host

Salem State College
Salem, Massachusetts 01970

76TH ANNUAL MEETING
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

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PREFACE

THE NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

The New England Intercollegiate Geological Conference (NEIGC) was organized in the fall of 1901 by William Morris Davis, who led a trip to the Westfield River area in Massachusetts (see Table 1 and Figure 1). Although no record of the first conference exists, it appears that there was a single field trip, perhaps lasting more than one day, and with Davis as the leader. Legend also has it that only a few people attended. With the 1984 meeting, the NEIGC will not have met during 8 years; 6 years during World Wars I and II, and an unexplained 2-year hiatus between 1912 and 1915. The present Secretary understands such a break, however, given the difficulty in finding people to take on the conference organization.

By 1927 or 1928, at least 2 separate field trips (bedrock and surficial geology) were being offered. With only a few exceptions, multiple field have been the rule. So many field trips were being offered by 1959 that only the conference organizers of that and subsequest meetings can be listed in Table 1; the number of trip leaders now exceed the number of people attending the early meetings.

The sole purpose of NEIGC has always been to conduct field trips in areas of recent geologic investigation. The 1984 meeting in northeastern Massachusetts and southwestern Maine certainly is no exception to this rule, with a new map of the bedrock of Massachusetts published in 1983, E-an Zen editor, and a new map of the surficial geology of the Commonwealth in the works, Byron Stone, editor. The State of Maine has two new maps in press; a map of the bedrock edited by Boone, Hussey and Osberg, and a surfical geology map edited by Borns and Thompson.

The early trip leaders and organizers of NEIGC were academics, but for a long time geologists from the United States Geological Survey, the Geological Survey of Canada, from numerous state geological surveys and from industry have all assisted in conducting field trips and organizing the conferences. The field trip leaders of the early conferences presented the geology and road logs in hand outs. These have been replaced by guidebooks that are now important sources of information for the geology of the northeast, and are shelved in some of the major libraries of the country.

The NEIGC has met outside of New England 9 times, indicating that the geology of New England extends far beyond its borders. Five meetings have been in New York, 3 meetings were held in the Province of Quebec, and 1 meeting was held in New Brunswick. In addition, many other conferences held near its borders have had individual field trips outside of New England. With the 1984 meeting in Danvers, the NEIGC has met 26 times in Massachusetts, 11 times each in Maine and Connecticut, 8 times in New Hampshire, 6 times in Rhode Island, and 5 meetings have been held in Vermont.

Future meetings of NEIGC are planned as follows: 1985 in New Haven, Connecticut, Robert Tracy, organizer; 1986 in Lewiston, Maine, Don Newberg and Marc Ratelle, organizers; and the 79th annual meeting in 1987 in Norwich, Vermont, with Fred Larsen and Dave Westerman as the organizers.

D. W. Caldwell, Secretary
NEIGC


LOCATIONS OF PAST NEIGC MEETINGS
QUEBEC
77
FREDRICTON
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15
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21
07,24
32
15
58
82
47
03,06,12,20
63
81
32
17
0
50
Miles
08
37
numbers refer to conference dates in this century
DUANE MATTERS

TABLE 1

LIST OF MEETINGS
OF THE
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

Meeting	Year	Location	Organizer
1st	1901	Westfield River Terrace, Mass.	Davis
2nd	1902	Mount Tom. Massachusetts	Emerson
3rd	1903	West Peak, Meriden, Conn.	Rice
4th	1904	Worcester, Massachusetts	Emerson
5th	1905	Boston Harbor and Nantasket	Johnson, Crosby
6th	1906	Meriden to East Berlin, Conn.	Gregory
7th	1907	Providence, Rhode Island	Brown
8th	1908	Long Island, New York	Barrell
9th	1909	North Berkshires, Mass.	Cleland
10th	1910	Hanover, New Hampshire	Goldthwait
11th	1911	Nahant and Medford, Mass.	Lane, Johnson
12th	1912	Higby-Lamentation Blocks	Rice
13th	1915	Waterbury to Winsted, Conn.	Barrell
14th	1916	Blue Hills, Massachusetts	Crosby, Warren
15th	1917	Gay Head and Martha's Vienhead	Woodworth, Wigglesworth
16th	1920	Lamentation and Hanging Wall	Rice, Foye
17th	1921	Attleboro, Massachusetts	Woodworth
18th	1922	Amherst, Massachussetts	Antevs
19th	1923	Beverly, Massachusetts	Lane
20th	1924	Providence, Rhode Island	Brown
21st	1925	Waterville, Maine	Perkins
22nd	1926	New Haven, Connecticut	Longwell
23rd	1927	Worcester, Massachusetts	Perry, Little, Gordon
24th	1928	Cambridge, Massachusetts	Billings, Bryan, Mather
25th	1929	Littleton, New Hampshire	Crosby
26th	1930	Amherst, Massachusetts	Loomis, Gordon
27th	1931	Montreal, Quebec	O'Neill, Clark, Gill
28th	1932	Providence, Rhode Island	Brown
29th	1933	Williamston, Massachusetts	Cleland, Perry, Knopf
30th	1934	Lewiston, Maine	Fisher, Perkins
31st	1935	Boston, Massachusetts	Morris, Pearshall, Whitehead
32nd	1936	Littleton, New Hampshire	Billings, Hadley, Cleaves
33rd	1937	New York City & Duchess Co.	O'Connell, Kaye, Fluhr, Balk
34th	1938	Rutland, Vermont	Bain
35th	1939	Hartford & Conn. Valley	Troxell, Flint, Longwell
36th	1940	Hanover, New Hampshire	Goldthwait, Denny, Stoiber
37th	1941	Northampton, Massachusetts	Balk, Jahns, Lochman
38th	1946	Mt. Washington, N.H.	Billings
39th	1947	Providence, Rhode Island	Quinn
40th	1948	Burlington, Vermont	Doll
41st	1949	Boston, Massachusetts	Nichols, Billings, Schrock
42nd	1950	Bangor, Maine	Trefethen, Raisz
43rd	1951	Worcester, Massachusetts	Lougee, Little
44th	1952	Williamston, Massachusetts	Perry, Foote, McFadyen
45th	1953	Hartford, Connecticut	Flint, Gates, Peoples
46th	1954	Hanover, New Hampshire	Elston, Washburn, Lyons
47th	1955	Tigonderoga, New York	Rodgers, Walton, Bartolome
48th	1956	Portsmouth, New Hampshire	Novotny, Billings, Chapman
49th	1957	Amherst, Massachusetts	Bain, Johannson, Rice

Meeting	Year	Location	Organizers
50th	1958	Middleton, Connecticut	Rosenfield, Eaton, Sanders
51st	1959	Rutland, Vermont	Zen
52nd	1960	Rumford, Maine	Griscom, Milton, Caldwell
53rd	1961	Montpelier, Vermont	Doll
54th	1962	Montreal, Quebec	Clark
55th	1963	Providence, Rhode Island	Quinn
56th	1964	Chestnut Hill, Massachusetts	Skehan
57th	1965	Brunswick, Maine	Hussey
58th	1966	Katahdin, Maine	Caldwell
59th	1967	Amherst, Massachusetts	Robinson, Drake, Foose
60th	1968	New Haven, Connecticut	Orville
61st	1969	Albany, New York	Bird
62nd	1970	Rangeley Lakes, Maine	Boone
63rd	1971	Concord, New Hampshire	Lyons, Stewart
64th	1972	Burlington, Vermont	Doolan, Stanley
65th	1973	Fredericton, New Brunswick	Greiner
66th	1974	Orono, Maine	Osberg
67th	1975	Great Barrington, Mass	Ratcliffe
68th	1976	Boston, Massachusetts	Cameron
69th	1977	Quebec City, Quebec	Beland, LaSalle
70th	1978	Calais, Maine	Ludman
71st	1979	Troy, New York	Friedman
72nd	1980	Presque Isle, Maine	Roy, Naylor
73rd	1981	Kingston, Rhode Island	Boothroyd, Hermes
74th	1982	Storrs, Connecticut	Joesten, Chestnut, Quarrier
75th	1983	Greenville and Millinocket, Maine	Caldwell and Hanson
76th	1984	Danvers, Massachusetts	Hanson

ACKNOWLEDGEMENTS

I would like to thank the Administration of Salem State College, for their continued support and assistance in the organization of this conference, and the Salem State Department of Geography, for their assistance in the production of illustrations. I would also like to extend my gratitude to all field trip leaders for their participation and efforts to make this a memorable conference.

Lindley Hanson

TABLE OF CONTENTS

QUATERNARY GEOLOGY AND GEOMORPHOLOGY

REGIONAL GEOLOGY AND TECTONIC HISTORY OF SOUTHEASTERN NEW ENGLAND

by

Patrick J. Barosh
Weston Observatory
Boston College

Introduction

Southeastern New England contains some of the most interesting, varied, and complex geology of all of North America. It lies astride the eastern edge of the Paleozoic Appalachian orogenic belt and offers a glimpse of this margin and the structure to the east of this belt that may not be seen elsewhere on the Atlantic coast of the United States. The boundary zone along the margin is the greatest structural zone known in New England and apparently represents a zone of collision between a Paleo-North American and Paleo-African plates. Structure and deposits related to the rifting of the present North Atlantic basin are also displayed in the region, as well as a wide variety of features and materials resulting from Pleistocene glaciation. Earthquake activity indicates it is still a tectonically active region. Numerous kinds of structures are present in the rock, which represents all periods from Precambrian to Quarternary with only the possible exception of the Mississippian.

Repeated glaciation during the Pleistocene stripped off the overbunden from the bedrock and then haphazardly recovered most of it with a wide variety of deposits. Since the retreat of the ice, many hollows left in the topography have filled with soft sediments and peat and the rising sea has formed coastal marshes and beaches. The rising sea level, regional tilt to the south due to post-glacial rebound and local tectonic subsidence have combined to create a very interesting and complex history along the coast.

The extreme variety in rock and surficial deposits and their structure make the region one of the most challenging anywhere for all types of geologists, as nearly every site is different. The growing need to understand more about water movement, both for supply and hazardous waste sites and ground conditions for an ever-expanding variety of construction requires more detailed knowledge of the geology than ever before.

A virtual explosion of new information has become available for the region over the past 15 years that has radically changed the earlier concepts of the region. Most of the data is still unpublished. It is derived mainly from the mapping of the Boston Office of the U. S. Geological Survey under L. R. Page and M. H. Pease, Jr., data from the New England Seismotectonic Study, sponsored by the U. S. Nuclear Regulatory Commission, and the recent work of the office of Marine Geology of the U. S. Geological Survey. This detailed mapping employed modern statigraphic studies, geophysical surveys, and radiometric age dating from well controlled samples, to great advantage. The understanding of the geology of the region is thus, undergoing many changes resulting from these new data and their ramifications. Much of the recent geologic literature on the area, however, reflects this understanding unevenly and is often contradictory. The field trips presented in this volume may not be entirely consistent with one another, but they maintain the valuable tradition of the New England Intercollegiate Geologic Conference of providing an opportunity of seeing the geology in the field. The participants may or may not agree with the interpretation presented, but they will be rewarded with some stimulating discussion and a greater knowledge of the geology of the region.

This report will briefly describe the geology of the region and its tectonic history to provide a background for the field trips. The following description may vary considerable from state geologic maps of Massachusetts (Zen, 1983), Connecticut (Rodgers, 1982), and New Hampshire (Billings, 1955), as considerable more recent work has been drawn upon for this report. This summary draws upon the mapping of a great many people whose work could not be all cited in such a short paper; most of this is listed in Barosh and others (1977). The references cited tend to represent more summary reports and newer findings. Much of the recent work contained herein was supported by the U. S. Nuclear Regulatory Commission under Contract Number AT (49-24) - 0291.

Structural Framework

Southeastern New England is formed of two vastly different geologic terrains separated by a wide zone of steeply west-dipping thrust faults, the Nashoba Thrust Belt (Barosh and others, 1977) (Fig.1). No rocks have been correlated across this belt, nor does the meta-volcaniclastic rock within it occur on either side. The Nashoba Thrust Belt forms the largest structural discontinuity known in the northeastern United States and apparently separates blocks belonging to Paleo-African and North American plates that collided here. This belt passes just to the northwest of Boston. Southeast of the thrust belt is a largely Precambrian granite terrain, the Southeast New England Platform, into which the Boston and other basins have been dropped. To the northwest lies the Sturbridge Geocline of pre-Ordovician meta-sedimentary rock with a block of Late Silurian (?) - Early Devonian (?) rock, caught between thrust fault zones. Remnants of Pennsylvanian and Late Triassic - Early Jurassic rock in grabens and fault slivers are present mainly in the Southeastern New England Platforms. All the terrains are overlapped just off-shore by Cretaceus and Tertiary deposits that form the submerged northward extension of the Atlantic Coastal Plain.

Southeast New England Platform

The Southeast New England Platform (Fig. 1) consists of a late Precambrian batholithic complex and associated metasedimentary and metavolcanic rocks that were intruded by plutons and covered by sediments at various times during the Paleozoic. The covering sediments are preserved in basins that are largely fault bounded (Fig. 1). These include the Boston Basin, which contains latest Precambrian to Middle Cambrian conglomerate, argillite, and volcanic rock (Kaye and Zartman, 1980; Kaye, 1981) and perhaps Orderician volcanic rock, the Narragansett and Norfolk Basins which contain Pennsylvanian stream deposits that locally overlie trilobite-bearing Cambrian limestone and phyllite (Shaler and

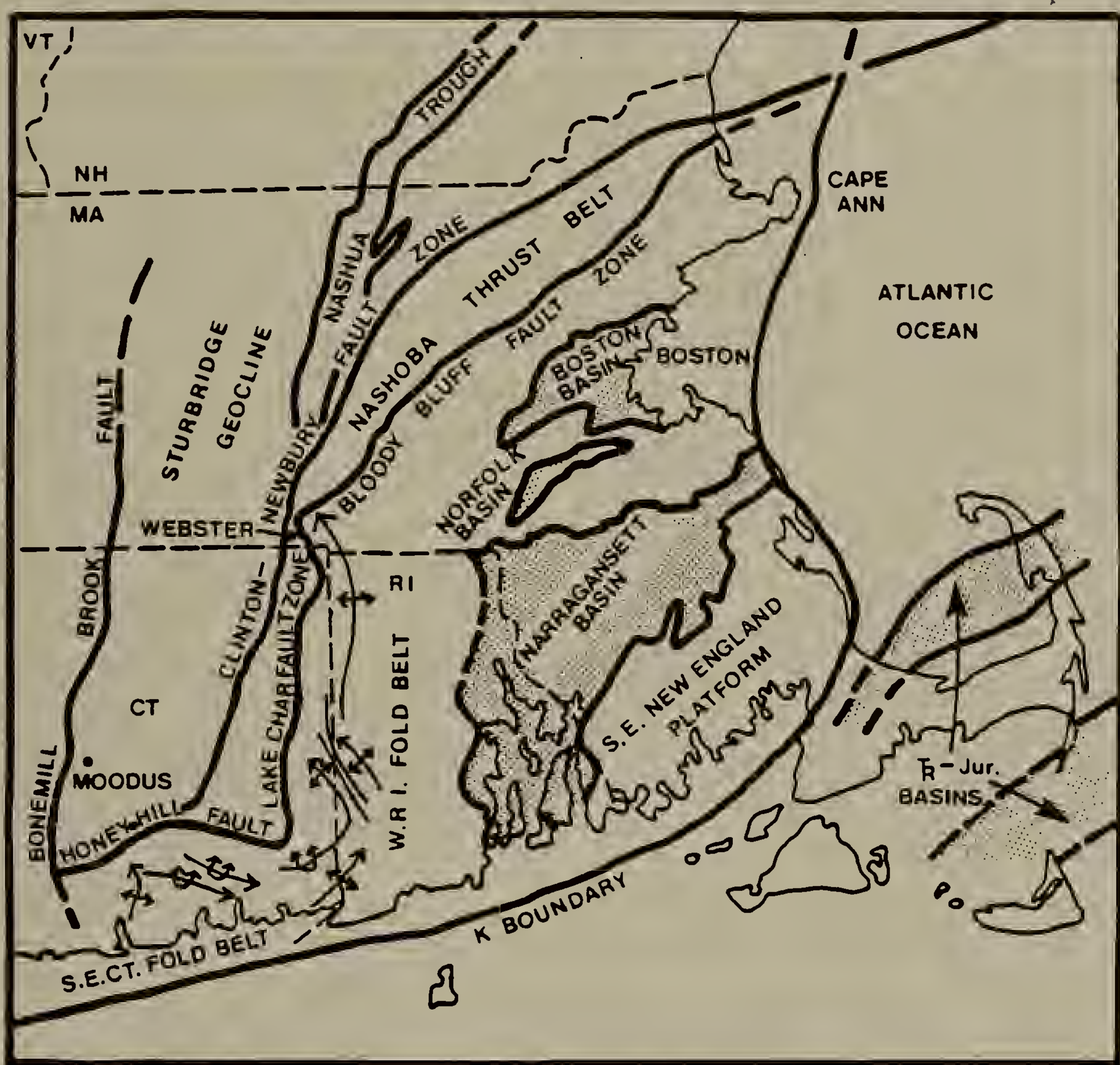


Fig. 1 Map of southeastern New England showing major tectonic provinces and structures.

others, 1899); and many off-shore and rare on-shore basins of Late Triassic to Early Jurassic sandstone, siltstone, and basalt (Ballard and Uchupi, 1975). Locally common are Paleozoic and Mesozoic dikes of diabase and lamprophyre.

The batholithic complex contains xenoliths and large pendants of meta-sedimentary and metavolcanic rock, that appear to be more abundant along its western side. These rocks form a sequence of quartzite and inter-bedded schist grading up into amphibolite and some metabasalt. They are broadly referred to as the Westboro Formation, Plainfield Formation or Blackstone Series in Massachusetts, Connecticut and Rhode Island, respectively (the Westboro Formation is examined northwest of Boston on trip B3 by Bailey). West and northwest of Boston, the quartzite is succeeded by a sequence of fine-grained, thin-bedded tuffaceous meta-sediments, including some very well laminated tuffs, that form the Middlesex Fells Volcanic Complex, the Greenleaf Mountain Formation, Burlington Formation and some unnamed gneiss and quartzite (Bell and Alvord, 1976).

The composition of the late Precambrian complex ranges from quartz-rich alaskite to diorite or gabbro (Hermes, Gromet and Zartman, 1981). Light colored granodiorite and quartz monzonite are common. The Dedham granodiorite is one of the more widespread plutons and at least several others are present. The division between plutons and local variations within them is not everywhere known. Most are nonfoliated and many are remarkably fresh looking, although the Dedham has undergone alteration in most places and the original light gray rock consequently has devel-oped a pinkish cast.

Rocks of the batholithic complex are in the range of 600 to 620 m.y. in age (Zartman and Naylor, in press; Galloway 1973; Smith 1978, Hermes, Gromet and Zartman, 1981), and have not been definitely found elsewhere in New England (the Massabesic Gneiss problem is discussed below). However, this complex is thought to correlate with late Precambrian plutons in the Avalon Peninsula in Newfoundland and together form a distinctive structural belt referred to as the Avalon Zone (Rast and others, 1976) (part of the batholithic complex north of Boston is visited on trip C3 by Smith and Hon).

The complex is more deformed towards the edge of the Nashoba Thrust Belt, marked by the Lake Char and Bloody Bluff fault systems, and may be strongly foliated, sheared, and folded adjacent to it (Figs. 1 and 2). The rock in Rhode Island, west of the Narragansett Basin, forms a complex broad north-trending dome, the Rhode Island Dome (Fig. 2). The general domal structure is defined by the attitude of bedding in the intruded Precambrian metamorphic rock. These strata lie mainly at the edges of the intrusive-cored dome; they dip to the west in eastern Connecticut, to the north in adjacent Massachusetts, and to the northeast in northeast Rhode Island.

This general domal structure is cut by numerous faults and is distorted by smaller folds. A series of north-trending, north-plunging folds, designated the West Rhode Island Fold Belt (Barosh 1976; Barosh and Hermes 1981; and Hermes, Barosh and Smith, 1981), lies along the Con-necticut-Rhode Island border (Figs. 2 and 3). These folds are broad

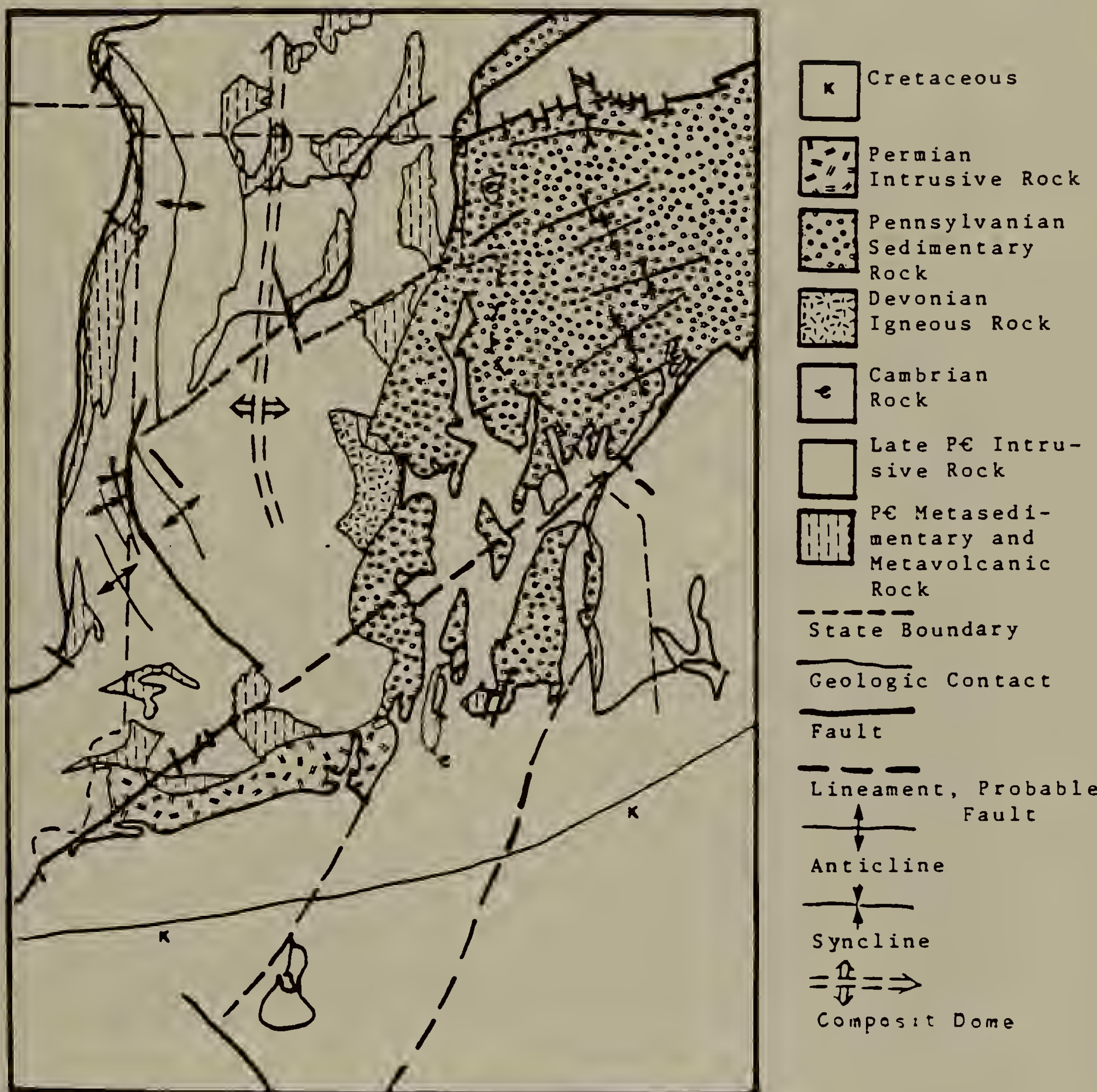


Fig. 2 Sketch map of Rhode Island and vicinity showing major structural features of the southeast New England Platform, (Barosh and Hermes, 1981)

and open in the north, but become progressively more compressed to the south where they are overturned and broken by thrust faults. The faults tend to cut out the synclines. The western part of the fold belt swings southwest and west approximately parallel to the Honey Hill fault zone where the folds are overturned to the northwest. The eastern parts swing southeast across southwestern Rhode Island and is overturned to the northeast. Both the northwest- and northeast-dipping overturned folds and associated thrust faults merge and appear to have formed at the same time.

The West Rhode Island Fold Belt ends to the north just inside Massachusetts, where the Nashoba thrust belt changes direction to the northeast (Fig. 1). Farther north, the batholith is strongly foliated and sheared adjacent to the thrust belt rather than folded (sheared batholithic complex along the Bloody Bluff fault is seen on trip C4 by Barosh). The apparent stratigraphic units formed by the pendants within the batholitic complex trend west-southwest and are cut off obliquely at the Bloody Bluff fault zone that marks the southeast edge of the thrust belt.

Several northeast-trending aeromagnetic and gravity lineaments cross the Rhode Island Dome. The distribution pattern and attitude of the Blackstone Series where crossed by these geophysical lineaments indicates that they are fault zones with a few kilometers of right-lateral offset each. One of these, the Watch Hill lineament, extends at least through southwestern Rhode Island as a fault zone (Hermes, Barosh and Smith, 1981), and projects into the Narragansett Bay where the bay changes shape (Fig. 2). It and several geophysically defined faults in the bay area may form a major northeast-trending zone of an echelon faults that continue northeastward across the bay and through Fall River, Massachusetts. Another northeast-trending aeromagnetic and gravity lineament crosses northern Rhode Island. (Barosh, Pease, and others, 1977). This offset on this lineament, indicating a fault, is supported by ground geophysical studies (Schwab and Frohlich, 1976), and by the approximate alignment of the lineament with a fault in Narragansett Basin (Fig. 2).

The major geophysical features in the poorly exposed areas southeast of the Watch Hill lineament trend north-north-east (R.K. Frohlich, oral comm.) as they also do in the Narragansett Bay and off-shore. This north-east direction appears to represent the major structural trend in this area (Collins and McMaster, 1978; McMaster and others, 1980). The structural grain north of the lineament appears to be north to north-northwest, as expressed by the trend of contracts and a few known faults.

The northern on-shore part of the Southeast New England Platform is cut by numerous faults that were formed and commonly re-activated over a long period of time. Most of the faults trend between northeast and east and relatively younger northerly trending faults are common (Fig. 4).

Paleozoic plutons and both Paleozoic and Mesozoic dikes cut the platform (Mafic dikes north of Boston are visited on trip A6 by Ross). Ordovician plutonic rock has invaded the Southeast New England Platform in

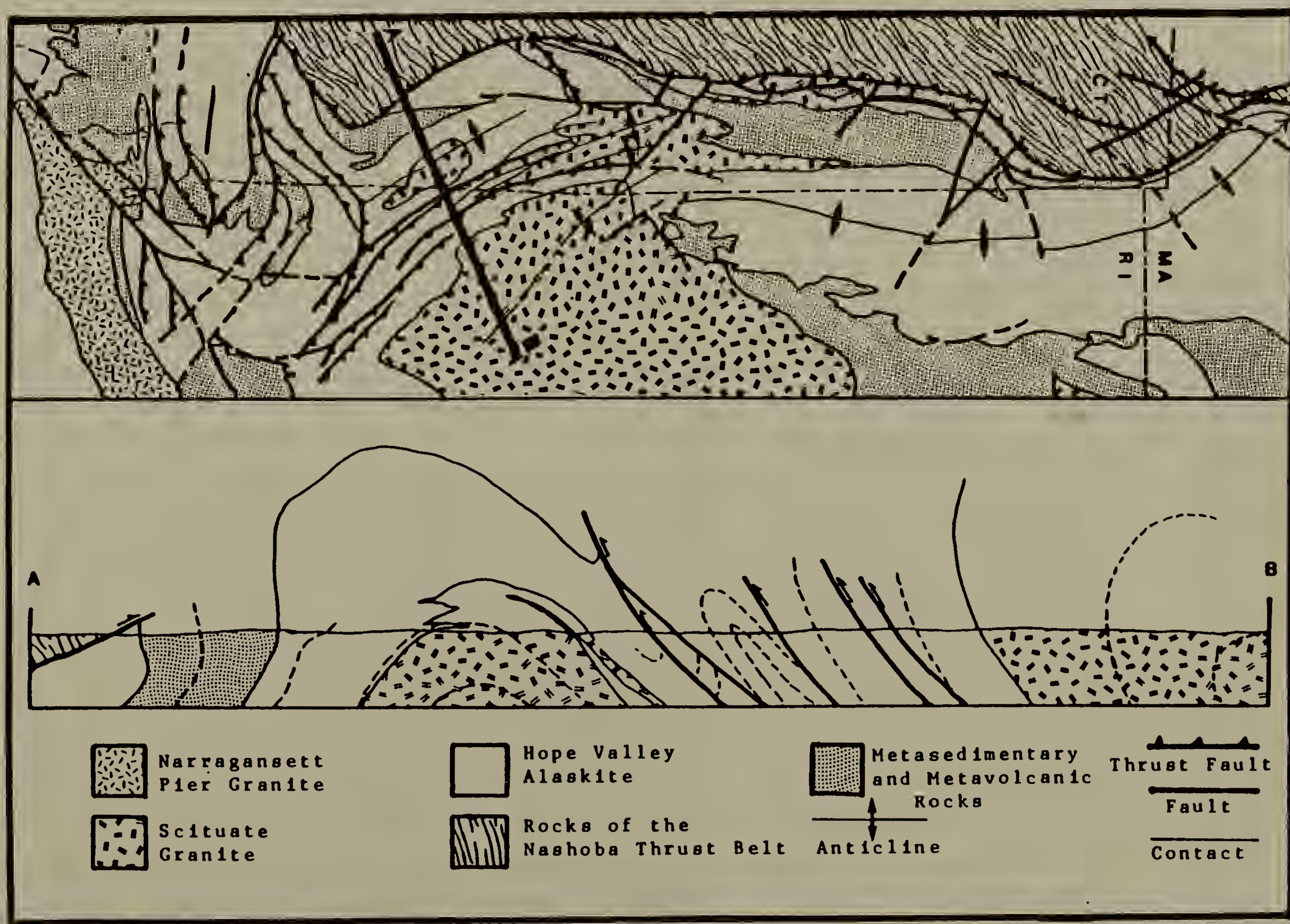


Fig. 3 Sketch map and cross section of the West Rhode Island Fold Belt (Barosh and Hermes, 1981)

places and forms most of the area around Cape Ann, north of the Boston Basin. The Cape Ann Granite, which consists of several phases of quartz monzonite, is approximately coevel with the adjacent highly-variable Salem Gabbro-Diorite (Dennen, 1976, 1981). These rocks are cut off to the west by the Bloody Bluff fault system, but display none of the intense foliation and shearing near the fault that the older batholithic rock does. The Ordovician intrusive rock south of Boston is well represented by the Quincy Granite, which is non-foliated and fresh looking.

Some intrusive rock within the batholithic complex in Rhode Island has recently been found to be Devonian in age; the Cowesett Granite and some within the Scituate Granite Gneiss to the west (Hermes, Gromet and Zartman, 1981) (Fig. 2). However, part of the batholithic complex in southwestern Rhode Island previously dated as Devonian (Moore, 1959) is found to consist of a mixture of late Precambrian and Permian rock.

The southern shore of Rhode Island is underlain by the Narragansett Pier Granite and its aplitic phase the Westerly Granite of Permian age (Quinn, 1971) (Fig. 2). This non-foliated granite and its associated pegmatite are found cutting the Precambrian rock for a considerable distance to the north (Hermes, Barosh and Smith, 1981).

In addition, a large basic intrusion of unknown age with a northward elongation, is interpreted from gravity and magnetic data to underlie the western side of Cape Cod (Barosh, 1976), and magnetic highs off-shore to the south may indicate others (Barosh and others, 1977). The Silurian Preston Gabbro of southeast Connecticut (Zartman in Dixon, 1982) has similar magnetic expression and these may be the same age or they could possibly represent Mesozoic volcanoes.

Several basins lie within and at the edge of the platform. The Boston Basin is a structural basin filled with a wide variety of only very slightly metamorphosed coarse conglomerate, volcanic rock and argillite unconformably overlying the batholithic complex. The age of the rock ranges from latest Precambrian to Middle Cambrian (Kaye and Zartman, 1980) and possibly to Ordovician (Cambrian strata are seen on the north edge of the basin on trip C1 by Bailey). The rocks in the basin are broadly folded and highly faulted, as seen in the several tunnels across the basin (The rocks of the basin and its structure are examined on trip B2 by Kaye) (The southeast side of the basin and the adjacent batholithic complex are visited on trip C10 by Brenninkmeyer and Dillon). The nearly east-trending fault system controlling the Boston Basin continues westward across the batholitic rock complex and ends against the Bloody Bluff fault zone (Figs. 1, 4 and 5).

Fossiliferous Cambrian limestone and meta-shale also occurs outside of the basin as an isolated exposure within the northern Narragansett Basin (Shaler and others, 1899), and near its southern edge, where fossils were found recently by Trem Smith (Skehan and others, 1977) (Fig. 2). Volcanic rock similar to that in the Boston Basin is present on either side of the basin. The Lynn series to the north and that in the Blue Hills to the south (C.A. Kaye, oral comm.). The Lynn volcanic rock is formed of a wide variety of intermediate to acidic volcanic rock on the south side of Cape Ann. Well preserved ash-flow tuffs are prominent in

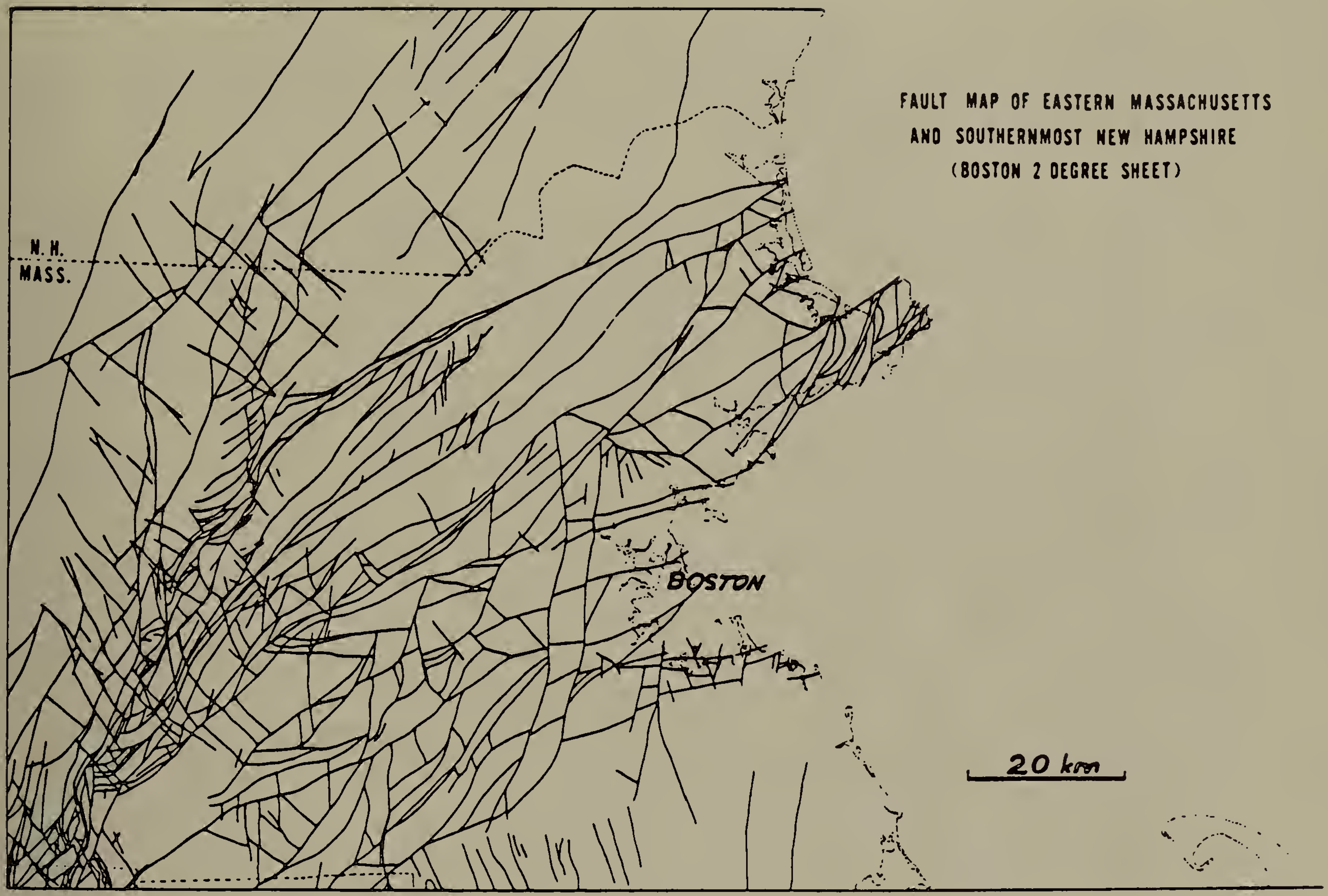


Fig. 4 Map of eastern Massachusetts and southern New Hampshire showing mapped faults (Barosh, Fahey and Pease, 1977)

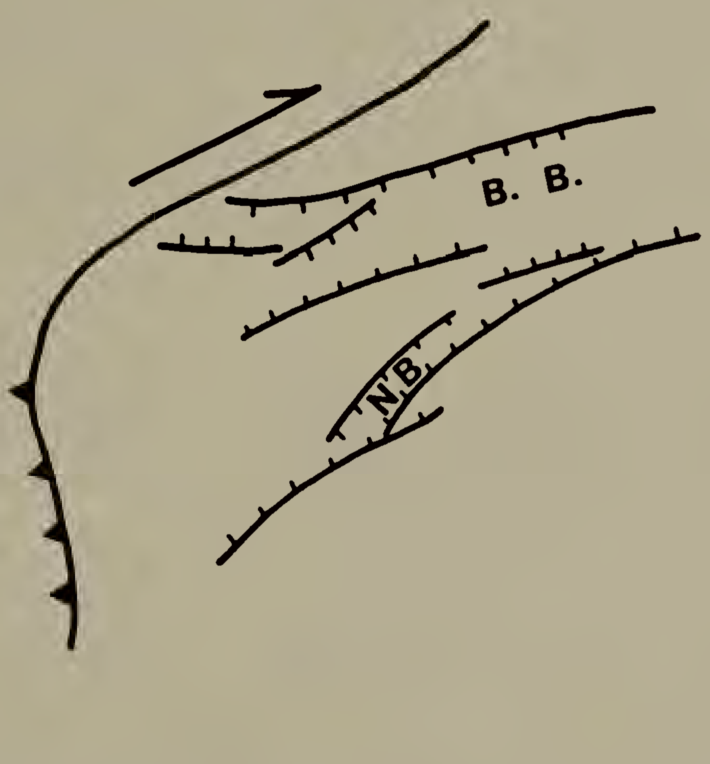
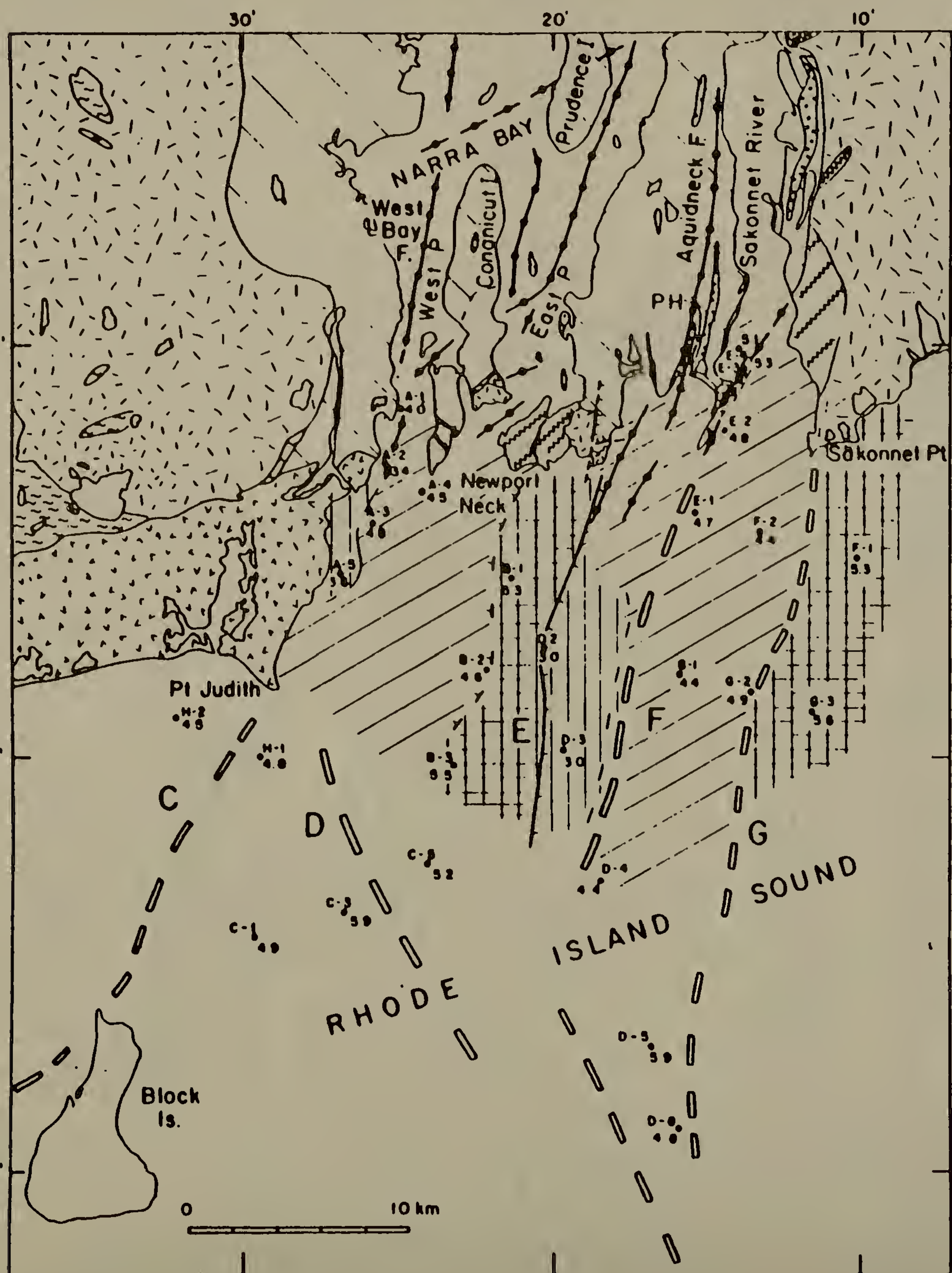


Fig. 5 Sketch map showing structural relations of the Boston (B.B) and Norfolk Basins (N.B.) with the eastern edge of the Nashoba Thrust Belt.

the Blue Hills. These volcanic rocks appear related to Ordovician intrusive rocks that cut the batholith (Dennen, 1981, Warren, 1913, Sayer, 1974), but may possibly be a little older (Kaye and Zartman, 1980; Naylor, 1981).

The largely fault-bounded Norfolk and Narragansett Basins form prominent basins to the south of the Boston Basin (Figs. 1 and 2). They are filled mainly with nonmarine conglomerate, sandstone, shale, and some coal of Pennsylvanian age (Shaler and others, 1899; Lyons, 1978). Much of the conglomeratic rock at the borders are red beds. Rock of the Narragansett Basin is both folded and faulted. The northern part of the basin exhibits a few broad east-northeast-trending bands of rock interpreted as large open folds (Shaler and others, 1899) (Fig. 2), but recent gravity profiles suggest they may be due to fault blocks (Sherman, 1978). In contrast, small isoclinial to recumbent north-northeast-trending folds, some of which are overturned, locally occur in the southern part of the basin. These folds have been interpreted by some workers to represent multiple episodes of deformation (Burks and others, 1981). However, the folding appears to be highly irregular and local and the adjacent older rock does not exhibit these folds. There has been a great deal of movement along the coal seams in this area (John Rabin, oral comm., 1981) and folding above these thrust faults and the soft sediment deformation that is present can account for most features. The structure of the poorly exposed parts of the basin is not well known, but much of the west side is faulted. A northeast-trending fault apparently forms the boundary northeast of Fall River (Fig. 2), and the northern border is cut by numerous small north- to northwest-trending faults (Fig. 1). The basin ends to the south with an irregular border formed mainly by probable north-northeast- and northeast-trending faults near the mouth of Narragansett Bay (McMaster and others, 1980) (Fig. 6). The basin rocks are metamorphozed to upper amphibolite facies in the southwestern part of the basin and adjacent to the Permian Narragansett Pier Granite (Shaler and others, 1899; Quinn, 1971; Hermes, Barosh and Smith, 1981). The grade of metamorphism decreases to the north (Shaler and others, 1899; Quinn, 1971; and Hepburn and Rehmer, 1981). Isograds of this Permian metamorphic episode are truncated by the Narragansett Pier Granite, and the rocks were subjected to local retrograde metamorphism.

Graben basins of Late Triassic to Early Jurassic red clastic rock and basalt lie off-shore and buried beneath the thick glacial outwash sands and gravels of Cape Cod (Ballard and Uchupi, 1975) and have been reached by drill on Nantucket Island (Fogler and others, 1978) (Figs. 1 and 7). Other similar grabens lie farther off-shore to the north and northeast and across central Connecticut and Massachusetts to the west (at the western edge of Fig. 1). In addition, red conglomerate, sandstone and siltstone, that form a fault sliver within the Bloody Bluff fault system northwest of Boston have been found by C.A. Kaye (1983), to be of Late Triassic-Early Jurassic age (Fig. 7). Another similar nearby basin in the Bloody Bluff fault system contains Late Silurian or Early Devonian red congolomerate, sandstone, and siltstone mixed with a large assortment of volcanic rock, termed the Newbury Volcanic Complex (Shride, 1976). This unit of mixed marine and terrestrial rock is not known elsewhere in southeastern New England, but has similarities to rock of the same age in coastal southeastern Maine.



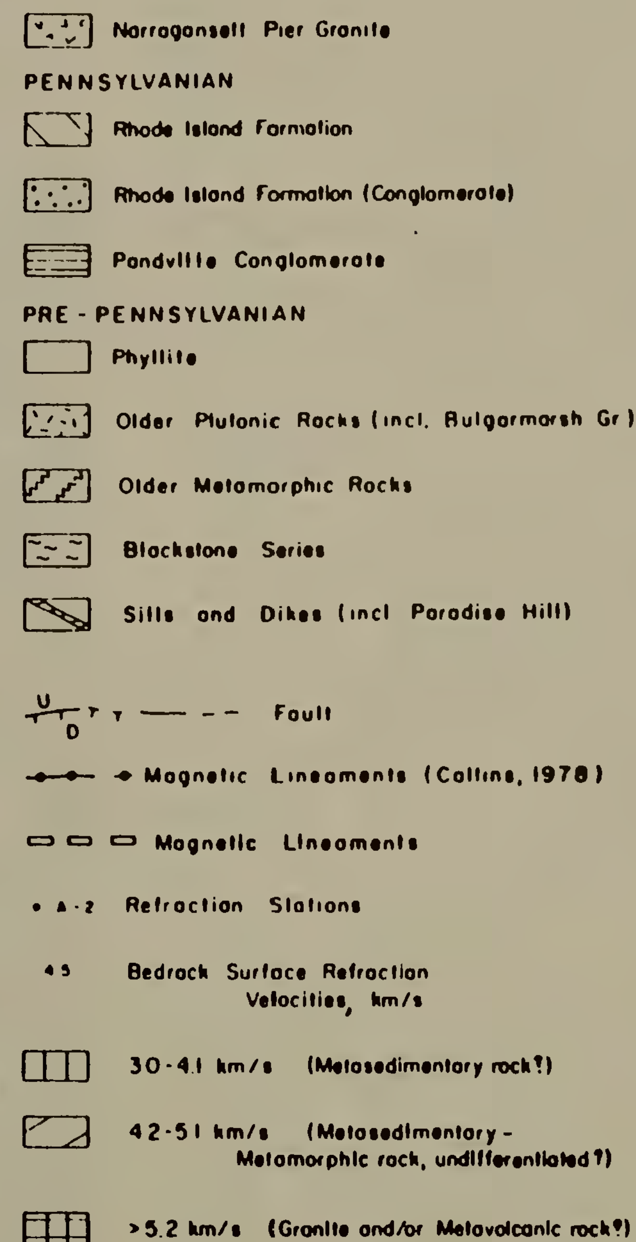


Fig. 6 Geologic map of southern Narraganset Bay and adjacent off-shore area (McMaster and others, 1980)

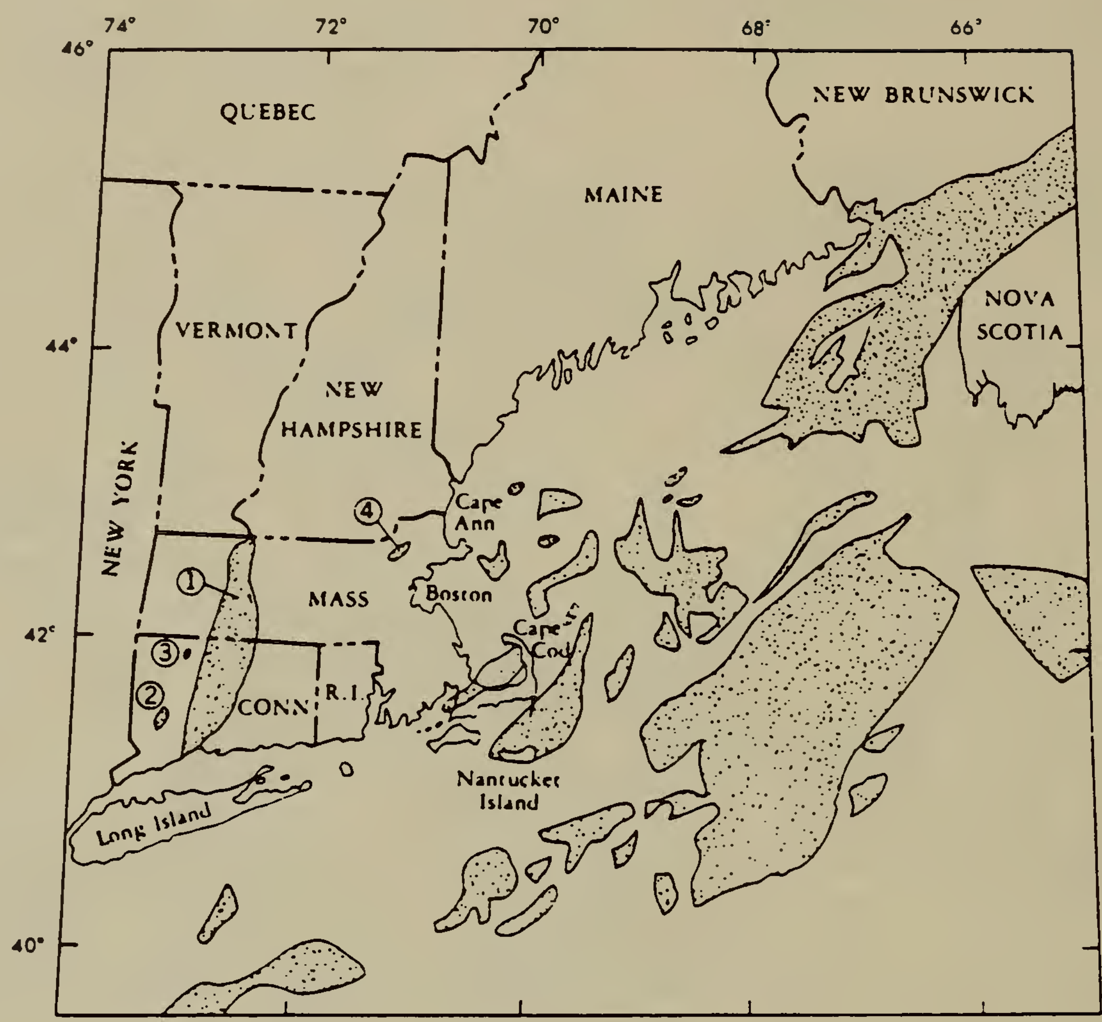


Fig. 7 Map of New England showing locations of basins containing the Late Triassic - Early Jurassic Newark Group (stippled). Off-shore basins from Ballard and Uchupi (1975). On-shore basins are the Hartford Graben (1), Pomperang Valley Graben (2), Canton Center Basin (3) and Middleton Basin (4) (from Kay, 1983)

A seaward thickening wedge of late Mesozoic and Tertiary deposits lies just off-shore and is part of the submerged northeast extension of the Atlantic Coastal Plain (Fig. 8). The Upper Cretaceous deposits offshore to the south form a northward-facing cuesta of sorts at the inner margin (O'Hara, oral comm., 1980). The consolidated Cretaceous clays and sands exposed at a few places on Block Island may be in place, but the Cretaceous and Tertiary deposits exposed at Martha's Vineyard to the east have been thrust up by glacial action (Kaye, 1964 a,b). Farther off-shore to the south and east a nearly continuous sequence from Middle Jurassic to upper Tertiary is present (Gibson and others, 1968; Valentine, 1981; Grow, 1981) (Fig. 8).

The small areas of Eocene sand and silt near the coast in the vicinity of Marshfield, south of Boston is the only outcrop of Tertiary now exposed on the mainland (Kaye, 1983), but others are known to lie just off-shore to the east (Weed, and others, 1974).

The Cretaceous is cut by a north to northwest-trending fault, the New Shoreham fault, off-shore south of Rhode Island (MacMaster, 1971). Some of the high-angle faults that cut the deposits at Marshfield and on Martha's Vineyard may be tectonic in origin and not related to glacial action (C.A. Kaye, pers. comm., 1983).

Sturbridge Geocline

The Sturbridge Geocline that lies in Connecticut, east-central Massachusetts, and southeastern New Hampshire to the west of the Southeast New England Platform and Nashoba Thrust Belt forms another major geologic province (Fig. 1). It is formed mainly of a very thick west-dipping and west-topping sequence of siltstone, graywacke, and shale. The sequence is known to be pre-Ordovician and is probably Precambrian in age (Barosh, 1981). These rocks have undergone high-grade metamorphism and are cut by numerous west-dipping thrust faults (Peper and others, 1975; Pease and Barosh, 1981).

This area has commonly been referred to as the Merrimack Geosyncline. However, it is not a geosyncline and has no structural connection with the structural trough containing the Devonian Littleton Formation in that was originally defined by Billings (1956), as the Merrimack geosyncline in north-central New Hampshire. Therefore, use of the term is confusing and should be abandoned. The Sturbridge Geocline consists of the generally northwest-dipping rock between the Nashoba Thrust Belt and the Bonemill Brook fault and other faults marking the east side of the intrusive cored uplifts of central Connecticut, Massachusetts, and southwest New Hampshire. It is named after Sturbridge, Massachusetts, which lies near the type sections of most of the meta-sedimentary units forming the province. The geocline is divided into western and eastern parts by the Nashoba Trough, a structural zone containing of relatively younger rock caught between steeply west-dipping thrust faults (Figs. 1 and 9).

The western part of the geocline consists of a very thick sequence of moderate to gentle northwest-dipping strata forming from base upwards the Oakdale formation, Paxton Group and Brimfield Group (Fig. 9). These units were defined near the turn of the century (Perry and Emerson,

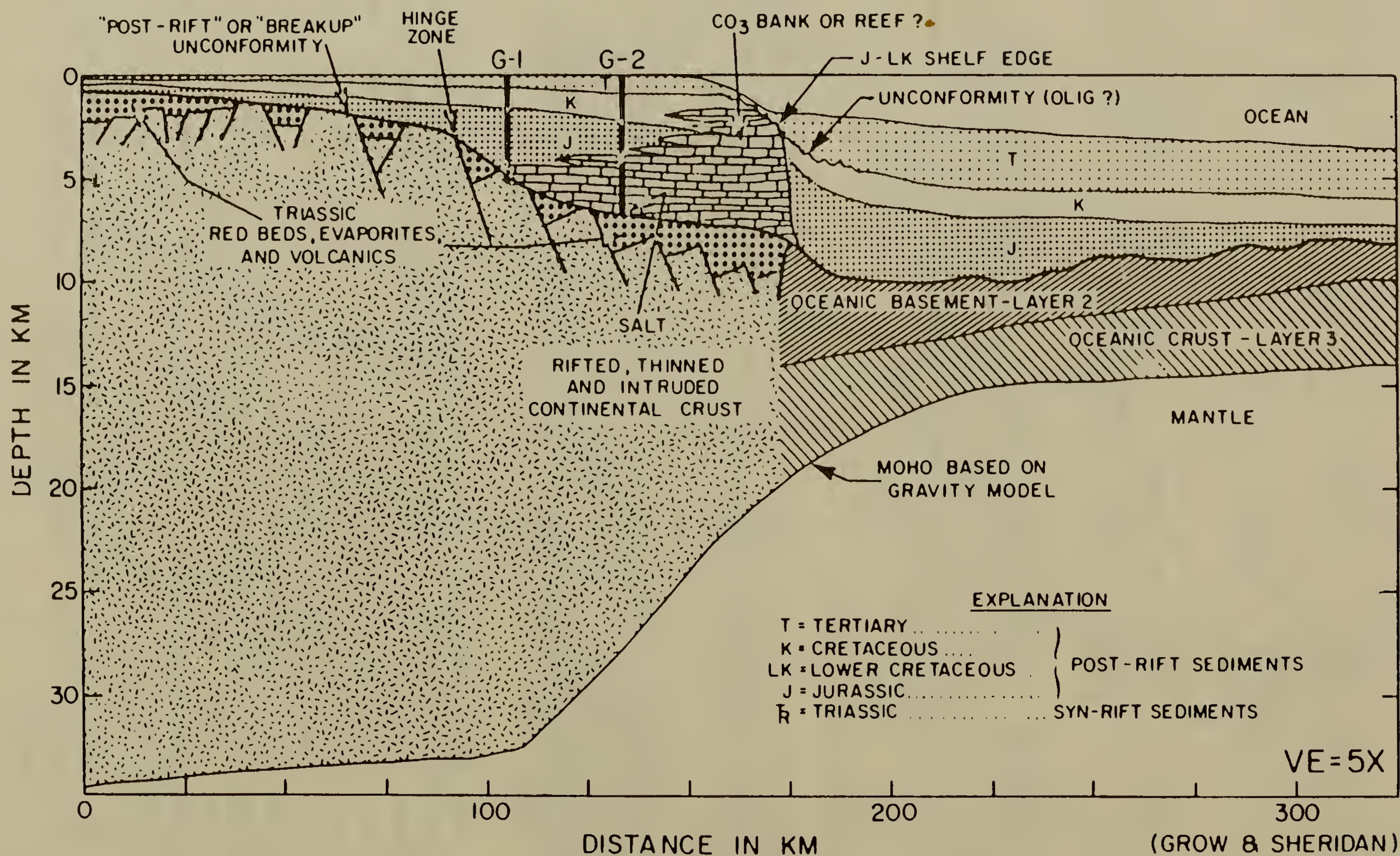


Fig. 8 Composite geologic section across southwest end of Georges Bank off-shore of southeastern New England (Grow, 1981)

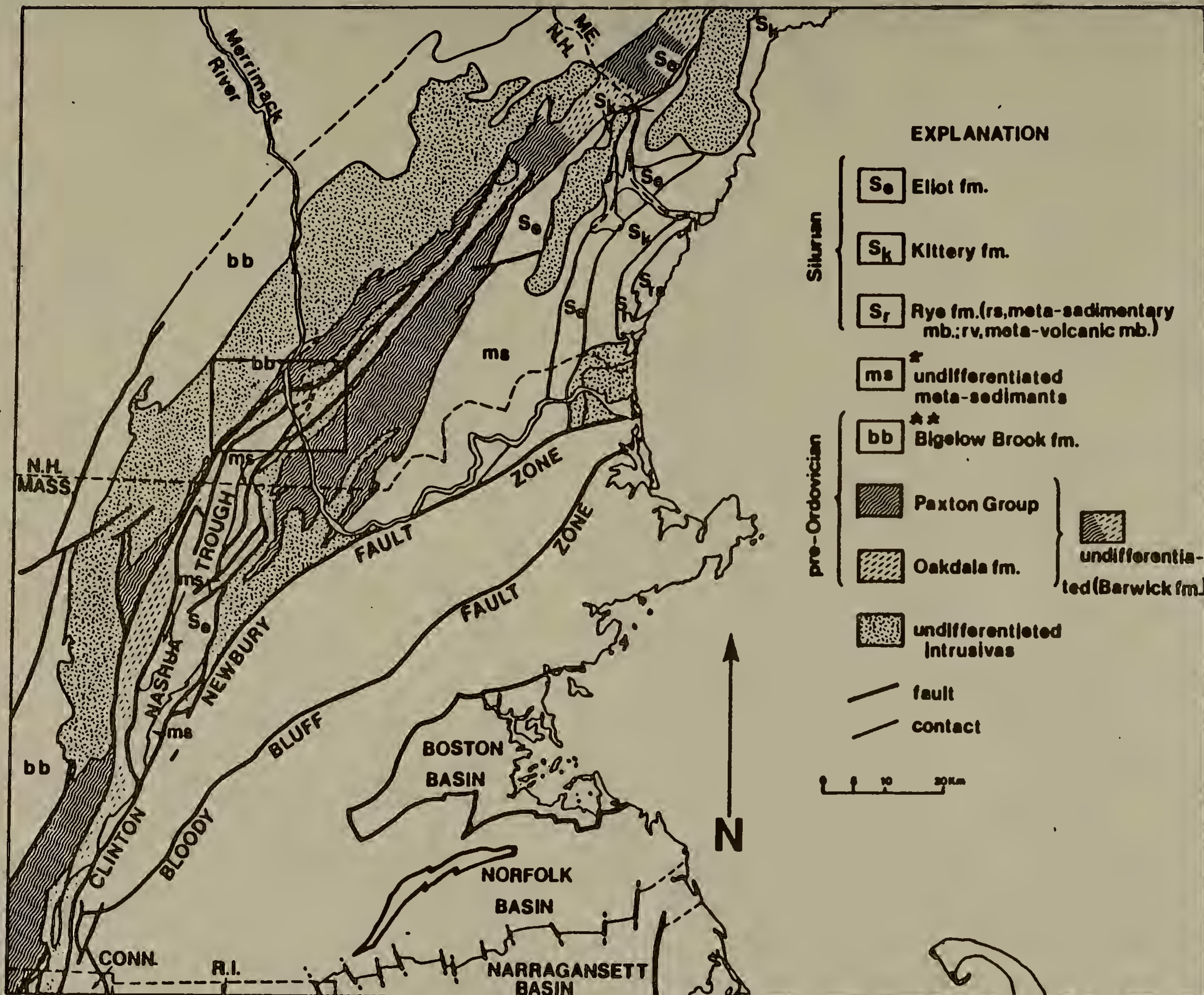


Fig. 9 Map of eastern Massachusetts and southeastern New Hampshire showing the bedrock geology of part of the Sturbridge Geocline (Smith and Barosh, 1982).

* MS is eastern block of Oakdale Formation

** Bigelow Brook Formation forms the base of the Brimfield Group

1903; Emerson, 1917) and have been painstakingly measured and redefined near the Massachusetts-Connecticut border in recent times (Pease, 1972; Peper and others, 1975, G.E. Moore, Jr. written commun. 1974; Barosh and others, 1977). They now have been followed southward to their terminous in east-central Connecticut and northeastward to southern Maine (Pease and Barosh, 1981; Barosh and Pease, 1981; Barosh and others, 1977) (Fig. 10) and provide the basis for understanding the structure of this geologic province. Other stratigraphic terms are still in use locally, such as the Berwick for the combined strata of the Oakdale and Paxton (The Berwick and adjacent rocks are seen on trip C5 by Eusden and others). The Shapleigh Group in Maine and much of the Littleton of southern New Hampshire are part of the Bigelow Brock and higher formations of the Brimfield Group (M.H. Pease, Jr., written commun. 1980) (Fig. 9).

The geocline is cut by numerous near bedding plane thrust faults, and a few high-angle ones in Connecticut. The thrust faults decrease in offset and die out to the north in Massachusetts, where a few very gentle folds are present, as shown in the Wachusetts water diversion tunnel across the region (Callaghan, 1931). The strata in Massachusetts and northward have been hypothesized to represent a series of very complex folded and refolded isoclinial folds (Thompson and others, 1968; Robinson and Hall, 1980), but several detailed stratigraphic studies and the continuous exposures in the tunnel indicate the limbs of the interpreted folds belong to different stratigraphic units (Peper and others, 1975; Pomeroy, 1974; Moore, 1976; Pease 1972; and Callaghan, 1931). No folds are observed in the area other than drag folds along thrust faults and the gentle folds seen in the tunnel.

Intrusive rock has invaded the early thrust faults at many locations, generally as elongated bodies. These mainly are quartz monzonite in composition and non-foliated to strongly foliated, depending largely upon whether or not movement on the intruded fault had stopped or continued. Most of the intrusions are pre-Late Silurian as they are older than the rock in the Nashoba Trough. The oldest, well-dated intrusive is Ordovician in age and cuts the Brimfield Group (Pease and Barosh, 1981). Another intrusion, the Massabesic Gneiss in southeast New Hampshire (Sriramadis, 1966), has been dated as late Precambrian (Besancon and others, 1977), but the sample area is highly contaminated by the invaded country rock and may be more representative of the age of the metasedimentary rocks than the intrusive. This migmatitic border zone forms the "gneissic" part of the Massabesic that otherwise appears less foliated than nearby intrusive rock considered Paleozoic in age. The Massabesic cuts both the Brimfield and Paxton Groups; the age of the meta-sedimentary sequence is therefore pre-Ordovician and probably Precambrian (Barosh and others, 1977, Barosh and Pease, 1981) (Some of these rocks adjacent to the Nashua Trough and important northwest-trending structures are seen on Trip C6 by Whitaker).

The Paxton Group and Oakdale Formation are repeated east of the Nashua trough by the major thrust faults bordering the trough (Figs. 1 and 9). The Oakdale Formation of this eastern block, referred to as "Eliot" by Sundeen (1971), also contains a thick muscovite schist lens equivalent to the thinner Gove member of the western block (Fig. 10). These

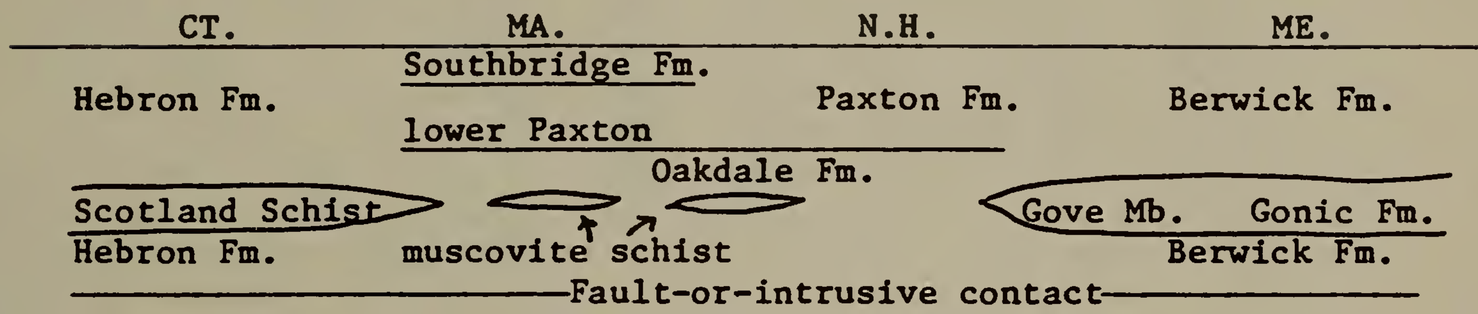


Fig. 10 Correlation chart for the Oakdale and Paxton Formations from Connecticut to southern Maine (Barosh and Pease, 1981). All overlain by the Brimfield Group.

formations are slightly finer grained than northwest of the trough and this suggests a source area to the northwest.

The Nashua Trough contains a sequence of rock consisting of thin-bedded quartzite, siltstone and shales, that are mildly metamorphized, described as units 1 to 4 by Peck (1976) (Smith and Barosh, 1981). These are all deposits formed on a submarine slope, distant from their source, by turbidity currents and, in some cases, slumps. Part of the sequence correlates with the Eliot Formation of southern Maine (Katz, 1918; Hussey, 1962). These are probably Late Silurian to Early Devonian in age. They were soft sediments and are locally isoclinaly folded near thrust faults. These folds are not known to extend into the adjacent older rock. The strata have not been seen to be intruded in this area. The trough impinges to the south on the Nashoba thrust belt and is cut off by the Clinton-Newbury fault zone along its border. Slivers of the rock in the trough are found in the fault zone farther southwest near Worcester and at Webster, near the Connecticut border (Fig. 1), providing some of the evidence for right-lateral movement along the fault zone. These younger strata are missing from the Merrimack River northeastward to the Lamprey River in New Hampshire and only the older more metamorphased Oakdale Formation is present in the trough.

Detailed geologic mapping and geophysical surveys show the Nashua Trough is formed of steeply north-dipping high-angle thrust faults; no major folds are present, although drag folds are common (Fig. 11) (Smith and Barosh, 1981, 1982). The trough forms the southern end of the Lewiston fault zone, that crosses all of Maine (Fig. 12). The zone has been gradually revealed by gravity, LANDSAT and local geologic studies.

Some of the rock in the Nashua Trough, the Eliot Formation, also occurs farther east in another faulted trough (Barosh and others, 1977) (Fig. 9). The units east of the Eliot, the Kittery and Rye Formations, probably underlie it, but their exact stratigraphic position is uncertain (These and nearby rocks are seen on trip A4 by Hussey and others) (The Rye Formation is examined on trip B4 by Swanson and Carrigan).

A volcanic chain developed in the Jurassic and Early Cretaceous along eastern New Hampshire and possibly off-shore to the edge of Cape Ann (One of the volcanoes, Mount Pawtuckaway, is examined on trip B7 by Eby).

High-angle northwest- and north-trending faults cut the northeast-trending thrust faults and are the youngest in the area (Figs. 4 and 11). Some are Late Cretaceous or younger as they cut upper Mesozoic volcanic complexes (Freedman, 1950).

Nashoba Thrust Belt

The Nashoba Thrust Belt is a major structural zone of closely spaced, steeply west-dipping thrust faults that separates the Southeast New England Platform from the Sturbridge Geocline (Figs 1 and 13.). The rock and structure of both these provinces terminate against it. No stratigraphic correlation has been possible between the pre-Ordovician

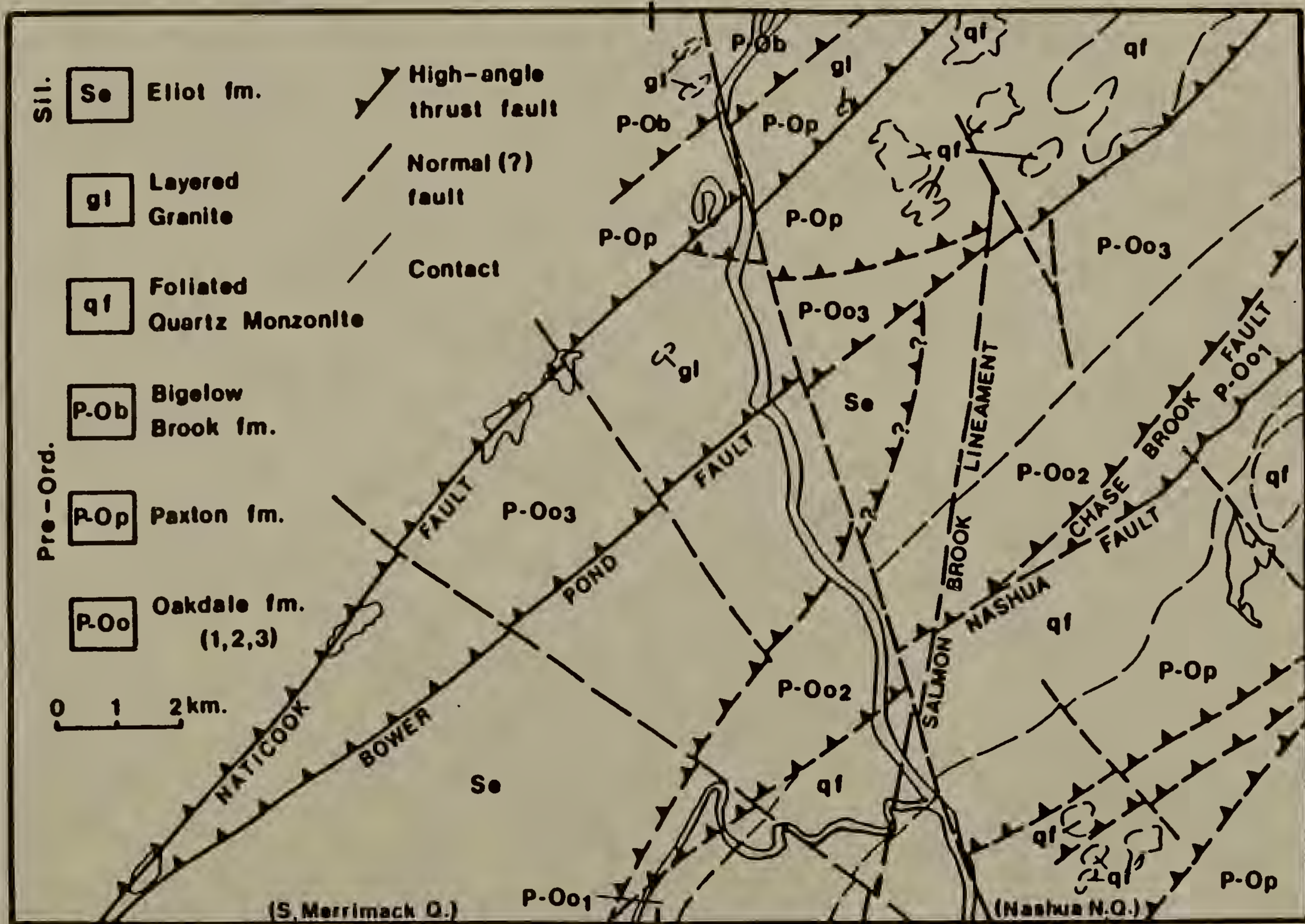


Fig. 11 Geologic sketch map of the Nashua North and South Merrimack 7.5-minute quadrangles, New Hampshire (Smith and Barosh, 1982). Location shown in figure 9.

rocks of these fault-bounded provinces and the rock in them probably formed at a considerable distance from one another.

The Nashoba Thrust Belt is wide in northeastern Massachusetts and contains nearly 18 kilometers of high metamorphic grade volcaniclastic rock measured northwest of Boston (Bell and Alvord, 1976, Alvord and others, 1976; Alvord, Pease and Fahey, 1976) (Fig. 13). This belt narrows drastically to the southwest, due to omission by thrust faults, and only a few hundred meters are present near Webster (Barosh, 1982) (Fig. 1); it widens again in eastern Connecticut. Many of the thrust zones in the belt are invaded by early to middle Paleozoic granitic rock, largely anatectic in origin. The units in the belt form a volcaniclastic sequence that consist mainly of thin, well-bedded amphibolite at the exposed base, a mixture of light-gray gneiss, schist, amphibolite, and marble in the middle and sillimanite-muscovite schist at the top. The units, known respectively as the Marlboro Formation (Quinebaug in CT.), Nashoba Formation of Hanson (Tatnic Hill in CT.), and the Tadmuck Brook Schist (not known in Connecticut), are described by Bell and Alvord (1976) (The Marlboro Formation is seen on trip C2 by DiNitto and others). These strata are invaded by a series of Ordovician (?) and Silurian (?) intrusive rocks and are probably Precambrian in age (The intrusive rocks are examined on trip A5 by Hill and others).

The principal movement along the northeast-trending thrust faults is west over east with a right-lateral component. These faults are cut by several small north- and northwest-trending high-angle faults (Fig. 4 and Fig. 4, Barosh, Trip C4)). Fault slivers of younger rock also occur along the borders of the Nashoba thrust belt in Massachusetts and attest to repeated movements along the thrust belt. Late Silurian (?) -Early Devonian (?) distal turbidites and Middle Pennsylvanian argillites occur in the Clinton-Newbury zone on its west side (Barosh, 1977) and Late Silurian-Early Devonian volcanic and sedimentary rock (Shride, 1976) and Late Triassic-Early Jurassic sedimentary rock in the Bloody Bluff fault system on its east side (the Bloody Bluff fault system and associated rock is examined on trip C4 by Barosh).

Tectonic Setting and Geologic History

The principal tectonic development of the region is believed to have taken place along the colliding border of two continental plates. The general concept of plate movements in New England was first mentioned by Wilson (1966) and elaborated on by Bird and Dewey (1970). Subsequent work has provided information on the structures involved and the timing of events. The movements appear to have occurred much earlier than first conceived. The Nashoba Thrust Belt apparently represents the west-dipping subduction zone between the Southeast New England Platform, a fragment of a former Paleo-African plate on the east, and the Sturbridge Geocline, a foreland basin of the North American plate (Barosh, 1979). These two plates probably moved towards one another obliquely along a path oriented east-northeast - west-southwest, as suggested by strain analysis of fault patterns (Barosh, 1976) and Paleozoic dikes and joints (Dennen, 1981). If the thick probable Precambrian strata of all three provinces in the region are broadly contemporaneous, then during the latter part of the Precambrian the following may have occurred. Muddy sand and sulfitic mud were carried southeastward and deposited in

Fig. 12 Map of the northeastern U.S. and adjacent Canada showing recent earthquakes and the Lewiston gravity lineament and other selected interpreted structures found or greatly extended on the basis of magnetic and gravity data (Barosh, 1982). 1. Northern Fall Line; 2. Connecticut River lineament; 3. Watch Hill lineament; 4. North Scituate-Blackstone lineament zone; 5. Higganum dike system; 6. Clinton-Newbury fault zone; 7. Winnipesaukee-Winooski lineament zone; 8. Penobscot lineament; 9. Norumbega fault zone; and 10. Upsalquitch lineament zone. Some geographical locations are: S, Sebago Lake; L, Lewiston; M, Medford; SL, South Lincoln; P, Passamaquoddy Bay; C, Chiputneticook Lakes; MB, Miramichi Bay.

a marine foreland basin now forming the Sturbridge Geocline. These sediments may have been derived from an inner volcanic arc, perhaps represented by some of the intrusive-cored domes of central Connecticut, Massachusetts, and western New Hampshire; an arc that stood off-shore east of the North American craton as reshaped by the 1,000 m.y. Grenville orgogony. Farther to the east andesitic and basaltic debris, along with some lime and mud accumulated around an outer volcanic arc. Across the sea to the east lay the edge of the Paleo-African craton being buried by quartz-rich sand, succeeded by mud, basaltic debris, basaltic and rhyolitic tuff and tuffaceous mud.

As the Paleo-African plate moved against and under the North American one, the outer arc became torn apart and carried into the subduction zone leaving only remnants of volcaniclastic rock found in the Nashoba Thrust Belt.

The main movement appears to have been in the late Precambrian with successive and gradually diminishing pulses during the Paleozoic. The greatest deformation on the Southeast New England platform is in the Late Precambrian rocks along its western edge against the Nashoba Thrust Belt. Late Precambrian intrusive cored (600-620 m.y.) folds are present and the intrusive rock along the Lake Char and Bloody Bluff faults is intensely sheared, whereas the Ordovician Cape Ann Granite and Salem Gabbro-Diorite along the Bloody Bluff fault zone, although faulted, are not intensely sheared. The Nashoba Thrust Belt parallels the general configuration of the folds, suggesting that thrusting was contemporaneous with the folding, a syntectonic late Precambrian event. It is also possible that the configuration of the late Precambrian folds somehow acted as a surface upon which later thrusting occurred, but this requires a number of coincidences. The major regional metamorphism occurred at this time on the platform and probably to the west as well.

Closely following the emplacement of the late Precambrian plutons and the accompanying tectonic activity, there was considerable uplift and erosion, perhaps due to isostatic rebound following the thickening of the crust in the subduction zone. The Boston basin was initiated in the latest Precambrian possibly due to extension caused by continued movement along the adjacent thrust belt (Fig. 5). The basin filled with a sequence of latest Precambrian near-shore volcanic rock and conglomerate interbedded with argillite, that grades upward to Middle Cambrian marine argillite (Kaye and Zartman, 1980). The rhyolitic and andesitic volcanic rock may reflect continuing activity along the western edge of the Paleo-African plate followed by a general transgressive sequence of off-shore muds and turbidites. Much of the rest of the platform was also covered by Cambrian sediments as indicated by scattered remnants. The coarse conglomerate and volcanic debris suggests high relief adjacent to the Boston Basin and probably basin and range structure existed then (C.A. Kaye, oral comm., 1982).

The Taconic orogeny affected the entire region and appears to have lasted over a longer period than in the western Appalachian Mountains, as suggested by the scattering of radio-metric age dates of associated plutons from mid Ordovician to mid Silurian. Thrusting took place in the Nashoba and Sturbridge terrains accompanied by formation of largely anatectic granite and pegmatite along the thrust faults. To the east,

Fig. 13 Geologic map of the central part of the Nashoba Thrust Belt in Massachusetts (Alvord and others, 1976). Intensive rock shown by stippling.

on the platform, the Cape Ann and Quincy Granite and Salem Gabbro-Diorite were intruded and accompanied by volcanism. The ash-flow tuffs in the Blue Hills probably came from a volcano represented by the Quincy Granite, although it is possible that they are slightly older (Naylor, 1981). The metamorphism accompanying the orogeny, although locally intense, is apparently less than previously occurred and the Boston Basin sediments were only very slightly affected.

Volcanic activity still affected the region in the Late Silurian or Early Devonian as shown by the variety of volcanic rock and red clastic rock mixed with marine sediments in the Newbury Volcanic Complex in north-eastern Massachusetts (Shride, 1976). Possibly, a volcanic chain connected them with the contemporaneous coastal volcanic sequence of eastern Maine. The two sites may have been much closer at that time and subsequently shifted farther apart by right-lateral movement along the Clinton-Newbury fault zone and others in the Nashoba Thrust Belt.

A deeper marine basin, into which sediments moved downslope southeastward in turbidity currents and occasional slumps, apparently lay farther northwest. The ridge from which similar southeastward moving sediments were derived passes through northwest Maine and northern New Hampshire (R. Moench, written comm., 1980) and probably continued through central Massachusetts and into Connecticut, perhaps farther away than the one that suppled the pre-Ordovician sediments.

Granitic rock intruded the region during the Acadian orogeny again over perhaps a wider age range than the traditional Middle Devonian date. Some cut the Sturbridge Geocline, especially in central New Hampshire and on the platform in Rhode Island. These generally alkalic rocks range in age from Early to Middle Devonian (Hermes, Gromet and Zartman, 1981). In Massachusetts, they include the Wenham Monzonite and the Rattlesnake Hill pluton (Lyons and Kruger, 1976). Southward in Rhode Island, alkalic rocks of the East Greenwich Group and parts of the Scituate Granite Gneiss yield Devonian ages (Hermes, Gromet and Zartman, 1981). These mid-Paleozoic plutons of the Southeastern New England platform, although generally contemporaneous with Acadian plutons in tectonic blocks to the west generally maintain a distinct petrologic character (Hermes, Gromet and Zartman, 1981). Moreover, the Spencer Hill volcanics of central Rhode Island have been interpreted by Quinn (1971) to be comagmatic with the nearby Devonian-aged Cowesett Granite. The Nashoba Trough does not appear to have been intruded, but the slight to moderate regional metamorphism that effected the rock in the trough (Peck, 1976) probably occurred at this time along with the development of local contact metamorphic aureoles.

Re-activation of the thrust faults in the Nashoba and Sturbridge provinces probably in the mid-Devonian caught up sedimentary rock in the Nashoba trough and other fault slivers and folded it locally during thrusting (Peck, 1976; Smith and Barosh, 1981). The thrusting continued to be west over east with a right-lateral component of movement.

The region may or may not have experienced the uplift and extensional faulting that led to the shedding of post-orogenic Late Devonian red clastic deposits in the coastal volcanic zone of eastern Maine (Schluger, 1973) and the Mississippian red beds in New Brunswick.

However, uplift, perhaps with associated extensional faulting, probably occurred on the Southeast New England Platform during the Pennsylvania to produce the non-marine conglomerate, sandstone, shale, and coal of the Narragansett and Norfolk Basins (Shaler and others, 1899). These deposits overlapped the Nashoba Thrust Belt as shown by the presence of a fault sliver of Pennsylvania rock on its west side in Worcester (Kemp, 1887; Grew, 1973). Fault scarps probably developed locally along the border of these basins during deposition, as indicated by the pebble to some very coarse conglomerates at the borders and the associated red sandstones and shales. These sediments are similar to those of the Late Devonian and Late Triassic-Early Jurassic grabens. The Norfolk Basin, in particular, may have formed similarly to that suggested for the Boston Basin by moment on the Bloody Bluff fault system (Fig. 5).

Metamorphism, intrusion and some deformation effected the southern edge of New England during the Permian Alleghenian orogeny. The sedimentary rock in the Narragansett Basin may have been deformed and metamorphosed mostly before the intrusion of the Narragansett Pier and Westerly Granites in the Permian (Burks and others, 1981; Hermes, Barosh and Smith, 1981), but these events are all part of the same tectonic episode. The highest grade of metamorphism roughly borders the southwestern margin of the basin adjacent to the granite and drops off to the north away from the granite. Illite crystallinity studies hint that two Alleghenian thermal events may have occurred (Hepburn and Rehmer, 1981). A number of recent studies indicate that the metamorphism may be more widespread than formerly realized (Zartman and others, 1970; Day and others, 1980; Dallmeyer, 1981). The metamorphism and intrusion are undoubtedly part of the same event with the high temperature just preceding the granite. The granite intruded in a relatively passive manner not disturbing the earlier structure of the rock (Hermes, Barosh and Smith, 1981).

Prior to the Late Triassic, the region was tilted to the north as shown by northward plunging structures and northward decrease in effects of different metamorphic events. Some of this tilt may have started early and contributed to the folding of the soft sediments at the south end of the Narragansett Basin.

Southern New England underwent uplift and extensional faulting during the Late Triassic and Early Jurassic as major rifting was initiated across the North Atlantic Basin. Deposition of continental clastic sediment and basalt occurred in local basins like that of the Hartford graben, that forms the Connecticut Valley (Hubert and others, 1978) and the Middleton basin (Kaye, 1983), but most basins now lie off-shore to the east (Ballard and Uchupi, 1975) (Fig. 7). Numerous diabase dikes were injected into the older rocks. Lamprophyric dikes also occur locally and may be of generally similar age (Ross, 1981). Continued extension into the Early Cretaceous probably helped create a volcanic chain that extended northerly, roughly parallel to the Hartford graben, along eastern New Hampshire. Normal movement appears to have occurred along many of the older thrust faults. Reactivation of the northeast-trending faults, that had a right-lateral component of movement during the Paleozoic, probably produced left-lateral movement (Ballard and Uchupi, 1975). However, there is no indication of any large-scale

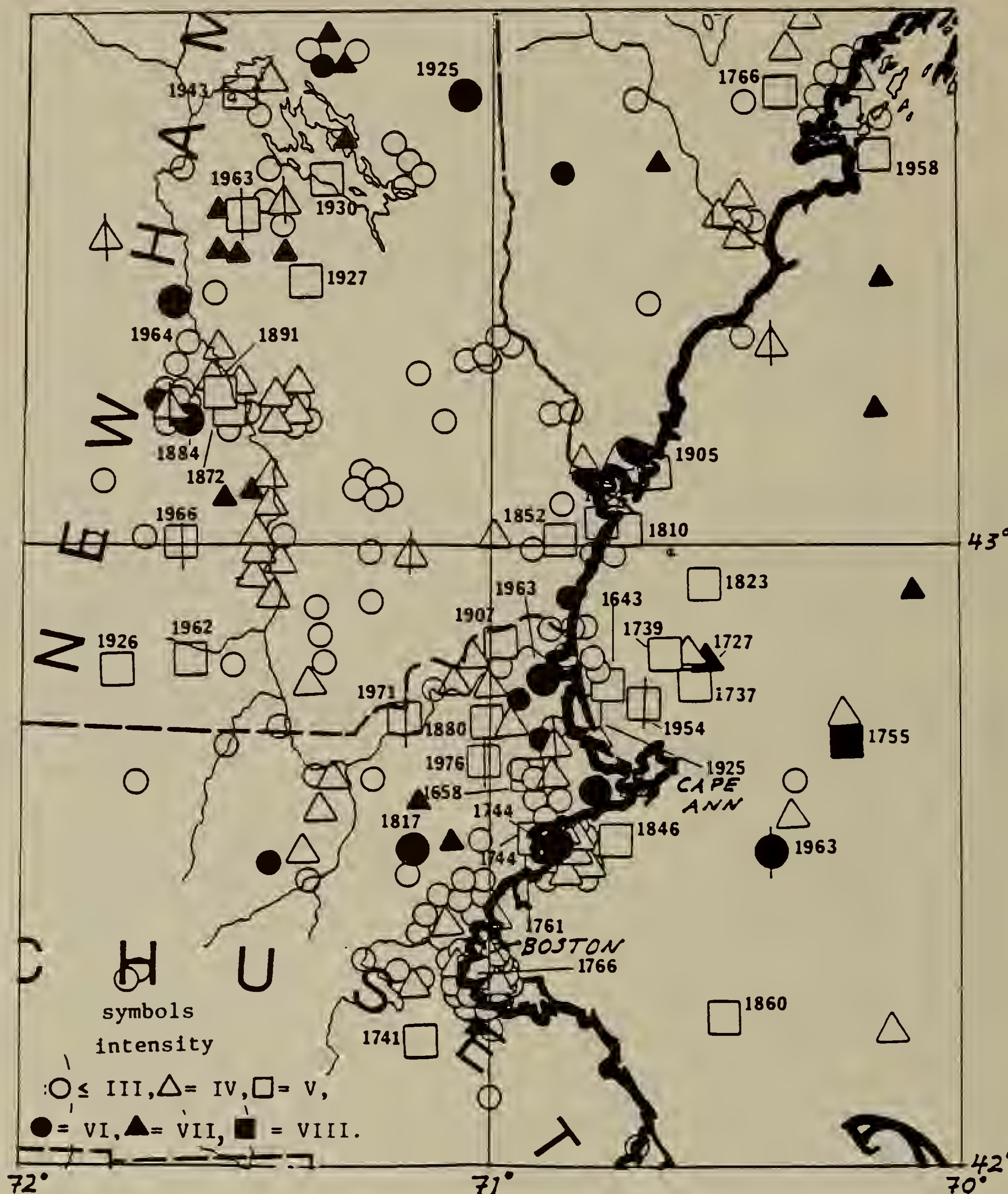


Fig. 14 Epicentral map of northeastern Massachusetts, southern New Hampshire and southern Maine through 1980 (from Nottis, 1983).

left-lateral movement such as that hypothesized by Kent and Opdyke (1978) from paleomagnetic data.

The southeastern edge of New England sagged to the south and east during the Middle Jurassic and Cretaceous as the North Atlantic Basin continued to open, and an apron of clastic sediments of Cretaceous age was deposited on it from the rising and eroding Appalachians to the west (Grow, 1981). Deposition continued into the Late Tertiary with an erosional hiatus during a low sea stand in the Oligocene (Weed and others, 1974; Valentine, 1981; and Kaye, 1983). Post-Cretaceous movements formed the north- to northwest-trending New Shoreham fault just west of Block Island (McMaster, 1971), and may have caused small movements, along many northwest-and some north-trending faults on shore.

Several periods of glaciation effected the region during the Pleistocene, but the only clear record is of the retreat of the late Wisconsin glacial cover. The entire region was covered by ice, including the near-shore area to the east and the region was depressed by the weight of the glacial ice. The rebound of the crust that began soon after the ice started its retreat 13,500 years ago has resulted in a regional tilt to the south of about 1m/km. This tilt and the post-glacial rise in sea level have caused the Late Pleistocene shore line to be deeply submerged off-shore to the south (USGS, 1976; O'Hara and Oldale, 1980), whereas it rises to the north above the present sea level just south of Boston.

Tectonic activity continues in the region as shown by the earthquakes around the Merrimack River Valley (Fig. 14), Cape Ann and Narragansett Bay. In the 1700's the Cape Ann area was much more active and two moderately large earthquakes occurred. These earthquakes appear to be due to local subsidence in the bays and river valley, that is related to continued opening of the North Atlantic basin (Barosh, 1981, in press). Short segments of northwest- and north-trending faults appear to be involved in this movement, particularly at fault intersections.

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QUATERNARY GEOLOGY AND GEOMORPHOLOGY

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The Quaternary geology of the coastal lowlands from Boston to Kennebunk, Maine is dominated by glaciation. Nearly all preglacial regolith was stripped off as glaciers scoured the landscape. Basins, such as the Boston Basin (Kaye, 1976), were excavated in regions underlain by less resistant rock, while preglacial valleys were widened and more deeply incised. Bedrock bosses were streamlined or rounded and an irregular blanket of till was deposited over the region. The drumlinoid topography characteristic of the Boston and Danvers areas (fig. 1) appears to be the product of at least two glaciations (Kaye, 1961, 1976, 1981; Oldale, 1964; Schafer and Hartshorn, 1965). In lowlands and valleys inland from the coast, ice retreat was accompanied by the deposition of glaciofluvial and glaciolaucustrine sediments (Mayewski and Birch, Trip C7; Koteff et al., Trip B9). Deglaciation of coastal regions was accompanied by a marine incursion which deposited a blanket of flocculated rock flour, or marine "clay". The thin deposits of marine clay, found as far south as Quincy, Massachusetts (Kaye, 1976), thicken and become more widespread northward into Maine. Figure 2 illustrates the areas of marine submergence, as determined from the occurrence of glaciomarine sediments. The thickness of glacial drift both onshore and offshore suggests that the average amount of glacial erosion may be 65 feet (approximately 20 meters) or more (Schafer and Hartshorn 1965).

Glacial History

Glacial events preceeding the early-Wisconsinan (Altonian) glaciation are poorly recorded in New England. Till deposits identified as pre-Wisconsinan (Kaye 1964a, 1964b; Oldale, 1982) are exceedingly rare and difficult to decipher. The scarcity of older drift probably attests to the highly erosive, cannibalistic power of the younger glaciers rather than to an absence of pre-Wisconsinan glaciation. If the weathered mantle formed through millions of years of exposure can be stripped off with barely a trace, the same may hold true for previous drift sheets.

Stratigraphic evidence on Long Island and the islands south of Cape Cod indicates that the southeast margin of the early-Wisconsinan ice sheet terminated on the continental shelf (Kaye 1964a, 1964b; Oldale, 1982; Sirkin, 1982). By 75,000 years B.P. the ice had retreated from New England and the St. Lawrence Lowlands. This date corresponds to the St. Pierre beds (Stuiver, et al., 1978, McDonald and Shilts, 1971) which separate the early-Wisconsinan Becancour and middle-thru late- Wisconsinan Gentilly tills in the St. Lawrence Lowlands (fig. 3). The non-glacial, freshwater St. Pierre deposits signify that the region at that time was ice free.

The early-Wisconsinan ice sheet deposited a thick blanket of till which was subsequently eroded during the late-Wisconsinan (Woodfordian) substage. Many drumlins in Boston and Danvers contain a mantle of oxidized drift, 20 to 50 feet thick, and are believed to have been formed by differential erosion and remolding of older Altonian, or possibly pre-Wisconsinan, drift during the late-

Wisconsinan glaciation (Kaye, 1976, 1981; Oldale, 1964; Schafer and Hartshorn, 1965). Oldale (1964) suggests that the early-Wisconsinan drumlin till may be as thick as 200 feet in the Salem-Danvers area.

Figure 1. Map of the northeast coast of Massachusetts showing the drumlin fields of the Boston and Danvers areas. In part after Kaye (1976). Many of these drumlins are composed of late-Wisconsinan, or pre-Wisconinan, till which has been more deeply weathered than surrounding late-Wisconsinan drift. The gravel beaches of Wintrop and Nantasket are derived from erosion of adjacent drumlin headlands. Note that majority of islands in Boston Harbor, including Thompson Island (TI), are composed of drumlins.

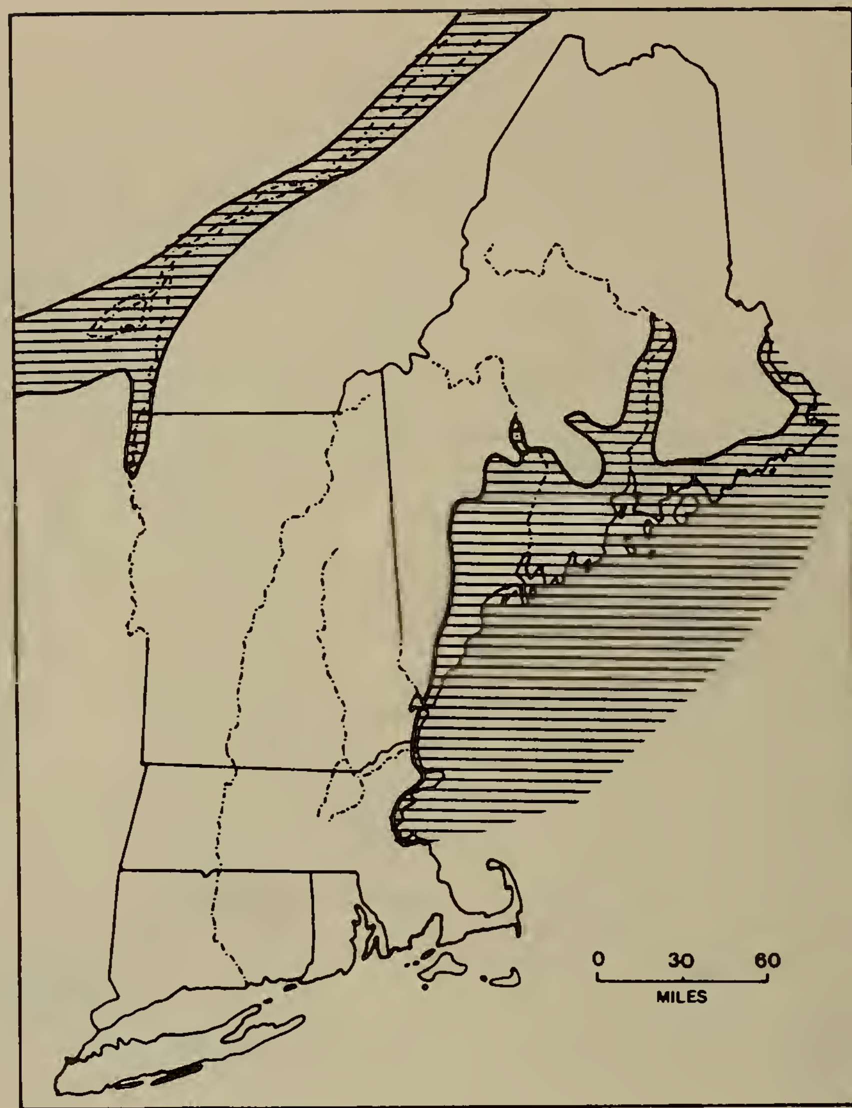

Figure 2. Map of New England and Southern Quebec illustrating regions inundated by the late- and post-glacial marine submergence (Compiled from numerous sources). Note the marine submergence is much more extensive in Maine. The marine transgression into the St. Laurence Lowlands separated the Laurentide ice sheet, stranding ice southeastern Quebec and in Maine.

The southeastern ice margin of the Altonian ice sheet consisted of numerous lobes which thrust sediment and dumped debris along their termini, initiating the construction of Long Island and the islands south of Cape Cod (Sirkin, 1982; Oldale, 1982).

Retreat of the early-Wisconsinan glacier from southeastern New England was followed by a lengthy interstade during which Altonian drift deposits were deeply weathered. Middle-Wisconsinan pollen stratigraphy of sediments on Long Island outlines a fairly detailed account of climatic trends preceeding the late-Wisconsinan glacial advance (Sirkin and Stuckenrath, 1980, Sirkin, 1982). Many climatic events recorded in the pollen zones of Long Island and Block Island correlate remarkably well with glacial events to the northeast in southeastern Quebec and the St. Lawrence Lowlands (fig. 3). The dissipation of the early-Wisconsinan ice sheet was followed by a period of warming, recorded by temperate forest pollen dated older than 42,000 B.P. (Sirkin and Stuckenrath, 1980). The Nassauan Spruce Pollen Zone (Sirkin and Stuckenrath, 1980) records a subsequent cooling trend between 42,000 and 33,000 B.P. and although no evidence indicates a readvance into southeastern New England at this time a glacial advance into southeastern Quebec is recorded by the deposition of the middle-Wisconsinan Chaudiere Till (McDonald and Shilts, 1971). Following the Nassauan substage a warming trend occurred between 33,000 and 28,000 B.P. as evidenced by the Portwashington Oak Pollen Zone of Long Island (Sirkin and Stuckenrath, 1980). In southeastern Quebec this warming trend is reflected by a glacial retreat to the St. Lawrence Lowlands and deposition of the glaciolacustrine Gayhurst Formation in the Chaudiere and St. Francis river basins. These glacial-lake clays are overlain by the late-Wisconsinan Lennoxvile Till (McDonald and Shilts, 1971). Cooling between 28,000-21,750 B.P. preceeded the arrival of the late-Wisconsinan Laurentide ice sheet into southeastern New England and is recorded by the Farmdalian Spruce Pollen Zone (Sirkin and Stuckenrath, 1980).

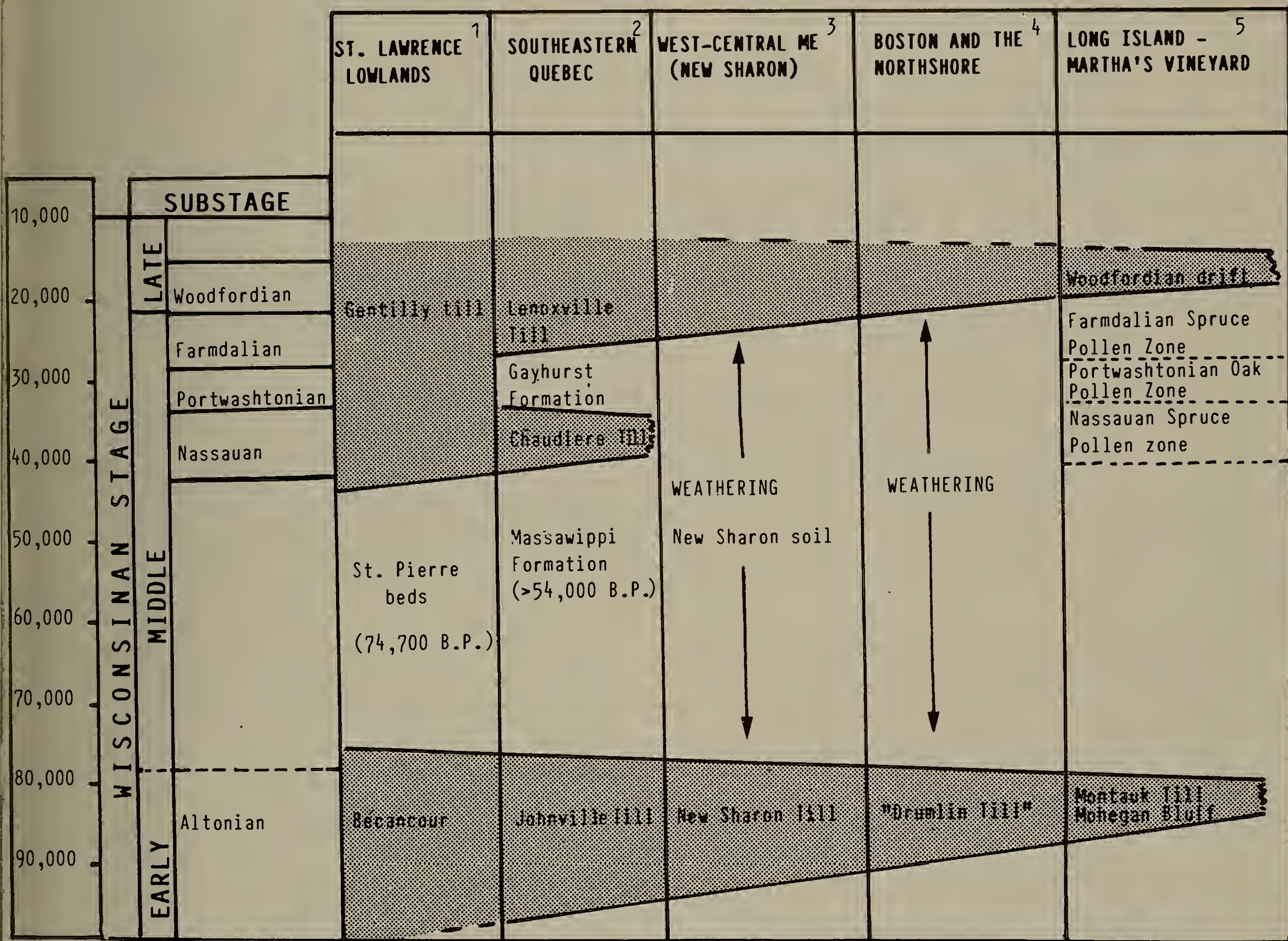

Figure 3. Time-distance diagram and stratigraphic chart correlating Wisconsinan substages, as defined by pollen stratigraphy on Long Island and Block Island (Sirken and Stukenrath, 1980, Sirkin, 1982), with glacial advances and retreats through southern Quebec and New England. Numbers refer to the following references: 1) McDonald and Shilts (1971); 2) Gadd (1971); 3) Caldwell (1959, 1960); 4) Odale (1964) and Kaye (1981); and 5) Sirkin and Stuckenrath (1980) and Sirkin (1982).

The Laurentide ice sheet which covered the region during the late Wisconsinan advanced southeastward through New England and reached its terminal position soon after 21,750 B.P. (Sirkin and Struckenrath, 1980). The ice margin was multilobate and occupied nearly the same position as the early-Wisconsinan glacier Long Island, Block Island, Martha's Vineyard and Nantucket are composed of the terminal moraines of both the early- and late-Wisconsinan glaciers (Sirkin, 1982; Oldale, 1982). In addition to Wisconsinan moraines, Kaye (1961b, 1961c) has found evidence of pre-Wisconsinan morainal deposits on Martha's Vineyard. Retreat of the Laurentide ice sheet from its terminal position on Martha's Vineyard occurred soon after 15,300 B.P. (Kaye, 1964b). The large recessional moraines and outwash plains of Cape Cod were deposited as oscillating ice lobes along southeastern ice margin followed a path of nonsynchronous retreat from the continental shelf into southeastern Massachusetts (Woodworth and Wigglesworth, 1934, Larson, 1982, Odale, 1982). Until the ice reached the Boston area its southeastern margin was predominantly land-based. Isostatic depression of coastal regions south of Quincy, Massachusetts was not sufficient to cause submergence beneath the prevailing sea level. Two dates from barnacle plates taken from marine clay in West Lynn show that the ice had retreated to the Northshore area by 14,000 B.P. (Kaye and Barghoorn, 1964, Kaye 1961, Stone and Peper, 1982).

In the Boston area Kaye (1982) has identified two late-Wisconsinan tills (Till III and IV). The older till (Till III) is the principal till of the late-Wisconsinan (Woodfordian) glaciation and is overlain by marine clay (Outwash III), commonly referred to as the "Boston blue clay". Locally, in the Back Bay and in Cambridge, a younger, enigmatic till, as thick as 35 feet (10m), overlies the clay and is thought to have been deposited by the Fresh Pond (Cambridge) readvance describe by Chute (1959). At least six dates, ranging in age from 12,200 to 12,700 B.P., on wood and peat collected from the underlying clay would place the readvance around 12,000 B.P. (Kaye, personal communication), at a time when most of New England is considered ice free. In explanation, Kaye suggests the possible existence of an ice mass in the uplands northwest of the Boston Basin. The structure, stratigraphic relationships, and wide-spread occurrence of the younger till in the Boston area, precludes the possibility that it may be a debris flow (Kaye, personal communication).

The sequence of events outlined by Oldale (1964) for the Salem area is quite different from that suggested by Kaye for the Boston area. Only one late-Wisconsinan till has been recognized and a lower relative sea level, concurrent with deglaciation of the coast, is inferred by the presence of low-level glaciofluvial sediments in coastal valleys. According to Oldale, further retreat was accompanied by a late-glacial submergence of low-lying coastal regions as the rise in sea level exceeded isostatic recovery. The submergence is recorded by marine clay found 50 feet or more above present sea level and high-level glaciofluvial deposits graded to the higher sea level. In contrast, glaciotectonic deformation of marine clays in Lynn attest to the proximity of the ice during submergence (Kaye, 1961; Stone and Peper, 1982). The deformed clays could be explained by a readvance, following a rise in sea level of over 50 feet, or by a re-interpretation of the low-level glaciofluvial sediments as discussed later in this introduction.

Soon after the onset of deglaciation, rapid calving of ice in the Gulf of Maine brought the glacial margin close to the Maine coast (Fastook and Hughes, 1982; Thompson, 1982). In southern Maine the ice had withdrawn from its terminal postion on the continental shelf to the coast by 13,500 B.P. (Stuiver and Borns, 1975), or possibly as early as 14,000 B.P. (Smith, Trip A7), where it remained until approximately 13,200 B.P..

Stratigraphic evidence recorded by numerous workers in Maine indicates that ice withdrawal from the coast was contemperaneous with marine submergence. However, Bloom (1960 and 1963) reported well sorted stratified drift, described in engineering testhole logs, 62-68 feet beneath present sea level and overlain by marine clay. He assumed these sediments to be subaerial-fluvial deposits based on their well-sorted character and postulated an emergence of approximately 70 feet prior to the marine transgression. Understanding of glaciomarine lithofacies, as outlined by Smith (Trip A7) can be invaluable to the interpretation of such sediments. The fluvial sediments described by Bloom are better interpreted as proximal subaqueous outwash formed by the retreat of a marine-based glacier. Such a re-evaluation of the low-level glaciofluvial deposits along the Northshore in Massachusetts may also be in order.

While ice still occupied coastal Maine, between 14,000 and 13,200 B.P., deglaciation of Massachusetts and southern New Hampshire was nearly complete (Caldwell et al., 1978; Koteff et al., Trip B9). Stratigraphic evidence throughout New England indicates that deglaciation was accomplished by active-ice, stagnation-zone retreat and not by regional stagnation (Koteff, 1974; Koteff and Pessl, 1981). Active-ice retreat produced sucessive sequences, termed morphosequences, of ice-contact and proglacial sediments which can be used to determine consecutive ice-marginal positions (Koteff, 1974; Koteff and Pessl 1981, and Koteff et al., Trip B9). Koteff et al (Trip B9) discusses morphosequences formed in a glaciolacustrine environment while tracing the retreat of ice up the Merrimac Valley. Mayewski et al. (Trip C7) illustrate evidence for active- ice retreat from coastal New Hampshire. Morphosequences and lithofacies formed in a glaciomarine environment are outlined by Smith (Trip A7).

As the marine transgression accompanied withdrawal of ice from coastal Maine and New Hampshire a variety of glaciomarine sediment were deposited, including a thick blanket of marine clay known in Maine as the Presumscott Formation (Bloom, 1960). Glaciomarine deltas graded to the higher relative sea level form the broad elevated plains commonly seen in the coastal lowlands (Trips A7 and B7). The blueberry industry of Maine owes its existence to the abundant, well drained, stratified glaciomarine sediments formed during this stage of deglaciation.

By 12,400 B.P. the ice had retreated inland from limit of marine submergence (Stuiver and Borns, 1975; Smith, Trip A7). Withdrawal of ice from coastal Maine was contemporaneous with the formation of a calving bay in the Gulf of St. Lawrence (McDonald, 1968; Borns 1963, 1966, 1967; Stuiver and Borns 1975). A later transgression of the Champlain Sea into the St. Lawrence Lowlands cut into the Laurentide ice sheet and stranded ice in southeastern Quebec and Maine (Shilts, 1976; Hanson 1977). Rapid dissipation of the ice soon followed. By 11,500 B.P isostatic rebound and consequent marine regression along the coast was complete (Stuiver et. al, 1971; Smith, Trip A7). The retreat of marine waters from submerged portions of the coast is recorded by numerous sequential beach ridges and wave-cut escarpments (Smith, Trip A7).

COASTAL GEOLOGY

Bedrock geology, glacial history, sediment supply, hydrographic regime, and post-glacial sea-level fluctuations all play an important role in the construction and modification of the present shoreline. The ragged outline of the coast from Boston northward, seen in figure 1, is largely the result of differential erosion and scouring of bedrock during glaciation and subsequent submergence. Major headlands composed of resistant rock surround embayments underlain by weaker rock. The headlands of Cape Ann and Salem are underlain by the more resistant Cape Ann Granite and Salem Gabbro-Diorite. Directly to the south, lies the Boston embayment underlain by the generally weaker, slightly metamorphosed, sedimentary rocks of the Boston Basin. Highly faulted, metamorphosed Precambrian and Paleozoic rocks form the embayment north of Cape Ann.

The preglacial shoreline, exposed to weathering and coastal processes for millions of years, was undoubtably smoother than the present shore and fringed with well developed beaches. These more mature beach sediments were carried offshore onto the continental shelf while glaciers scoured the underlying bedrock and deposited irregular sheets of till, sand and gravel.

Figure 4. Nasa Landsat image of of the coast from Boston Massachusetts to York Harbor, Maine.

Drumlins, minor glaciofluvial deposits, and bedrock supply most of the beach sediments from Scituate to Cape Ann. Drumlins dominate the landscape in the Boston area and along the north and south shores. In Boston Harbor, eroding drumlins form small islands and shoals. A till drumlin and gravel outwash plain are combined to form Thompson Island, one of the larger islands in the harbor (Caldwell, Trip C8). Erosion and redeposition of these glacial deposits by waves has produced a variety of gravel spits on the island (Rosen, Trip A1). Johnson (1910, 1919, 1925) illustrated how erosion and redistribution of drumlin-derived sediment into gravel tombolos and spits lead to the construction of the Winthrop and Nantasket barrrier beaches (Brenninkmeyer and Dillon, Trip C10; FitzGerald, Trip A1).

The large sandy barrier islands of Plum Island and Castle Neck are nourished by the Merrimack River, as it erodes through the thick blanket of glaciofluvial sediments left in it's valley. Whether these barrier beaches were created from spits, high dune ridges, or offshore bars which migrated shoreward during the Holocene transgression has been the subject of much controversy. For further discussion see McIntire and Morgan (1963) or Jones and Cameron (1976).

Hydrographic regime governs the patterns of sediment transport and deposition that contol the shape and dimensions of the barrier beaches, their intervening inlets, and tidal deltas. The coast from Boston to Kennebunk is predominantly a mixed energy coast (Hayes, 1979) whereby tides and waves maintain nearly equally important roles in sediment transport and deposition (see FitzGerald, Trip A1). Barrier beaches are short, backed by marsh and tidal creek systems, and are associated with well developed flood and ebb tidal deltas.

The rise in relative sea level has resulted in the landward migration of barrier beaches for the past several thousand years. The shoreward migration of barrier beaches is a cannibalistic processes. Overwash, tidal, and eolian processes transport sediment from the foreshore to the back-barrier environment, thereby enabling landward migration and preservation with rising sea level. According to Kaye and Barghoorn (1964), sea level rose from a minimum of -70 feet, at 10,000 B.P., to approximately its present position by 2,000 B.P.. Since this time sea level has been fluctuating to within 1.5 feet of present sea level. Using barnacles as an indicator of recent sea level fluctuations, Kaye (1964) observed that between 1922 and 1931, following a fall in sea level of .17 feet since 1856, sea level was -.07 feet. By 1961 sea level has risen .72 feet and is continuing to rise. The rapid rise in sea level in recent years has lead to accelerated erosion and destruction in coastal communities. Many barrier beaches, such as those in Revere, Lynn, and Swampscott, have been over developed. The construction of seawalls and other coastal structures, which decrease foreshore erosion and back-barrier sediment transport. may ultimately lead the destruction of these barrier beaches if sea level continues to rise. FitzGerald (Trip A1) discusses the effectiveness and consequences of various coastal structures along the Winthrop shore.

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COASTAL GEOLOGY OF WINTHROP, MA

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INTRODUCTION

The Winthrop shoreline, which is comprised of open ocean beaches and protected harbor environments, has been studied by several members of the Coastal Environmental Research Group, Boston University. The results of these investigations are reported in two Masters theses (Levin, 1981; Sullivan, 1982) and in a number of publications (FitzGerald, 1980, 1981; FitzGerald et al., 1981; Levin and FitzGerald, 1981). This paper is a compilation of these works.

Winthrop's 7.5km eastward facing coast encompasses Winthrop Beach to the north and Yirrell Beach to the south (Fig. 1). This stretch of shoreline contains numerous engineering structures and thus presents an excellent opportunity to determine the influence of different types of coastal structures on beach processes. The sand and gravel composition of the Winthrop beaches also provides a means of assessing sediment transport under varying energy conditions. Many of the sediment transport patterns that are reported in this paper were a result of the February Blizzard of 1978, the largest storm to affect this region in 50 years.

PHYSICAL SETTING

The Town of Winthrop is a peninsula forming the northeast boundary of Boston Harbor (Fig. 2). Like much of the surrounding region its topography is dominated by drumlins that have been connected by sand and gravel spits (Johnson, 1919) (Fig. 3). Beaches consist of moderately-sorted sand with local concentrations of gravel. The gravel content of the beaches increases toward the drumlin headlands and in an offshore direction.

The entire shoreline in this area is backed by seawalls of various construction except for a small section of beach south of Point Shirley (Fig. 1). Other coastal structures along the Winthrop shore include five groins, two pedestrian ramps and five closely spaced offshore breakwaters.

The region has a mean tidal range of 2.8m increasing to 3.3m during spring tides. Seasonal wave energy fluxes for this part of the coast are shown in Figure 4. Maximum wave energy occurs during the fall and is smallest during the summer. The dominant wave approach is from the east-northeast, a condition prevalent during northeast storms. The highly irregular nearshore and offshore

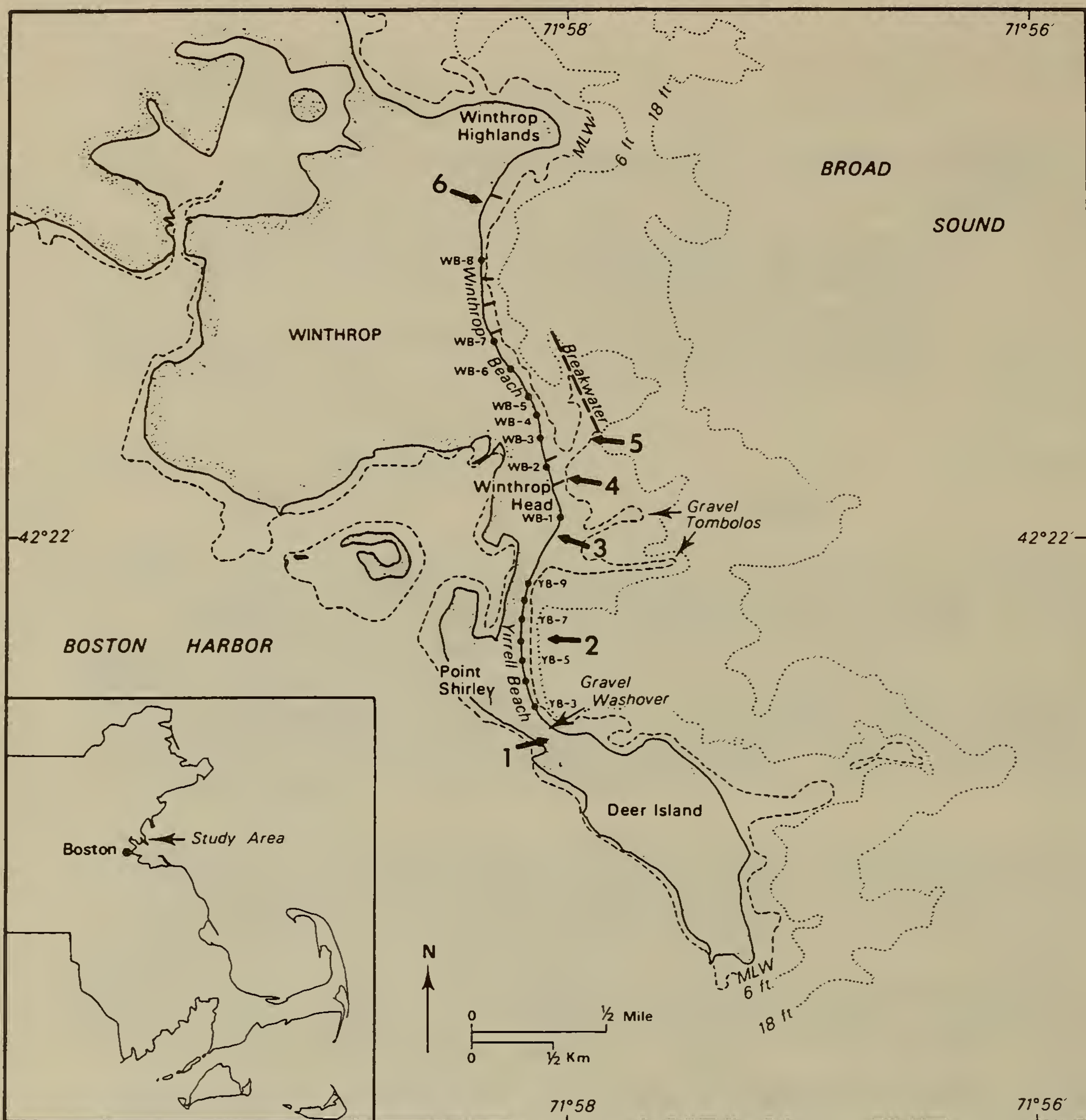


Figure 1. Location Map and fieldtrip stops.

Figure 2. 1978 Vertical aerial photograph.

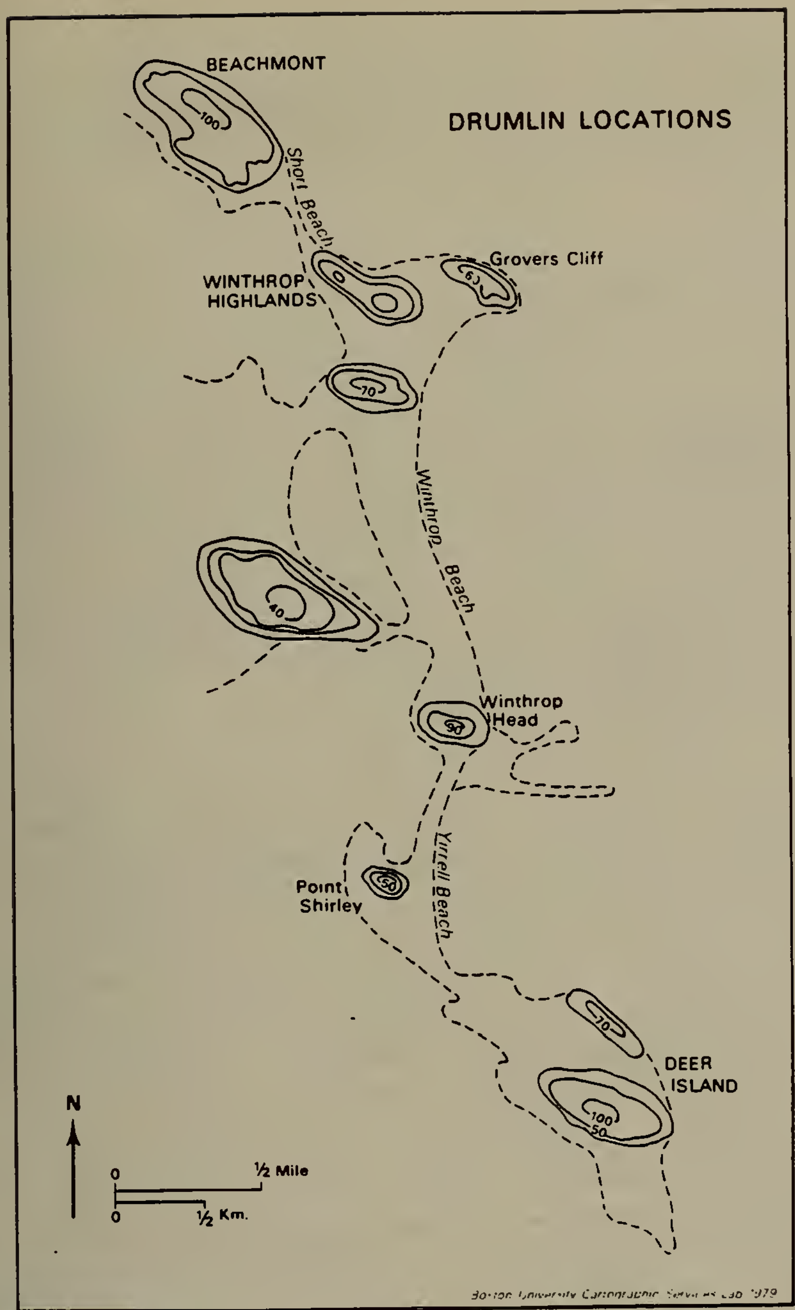


Figure 3. Location and height of Winthrop area drumlins.

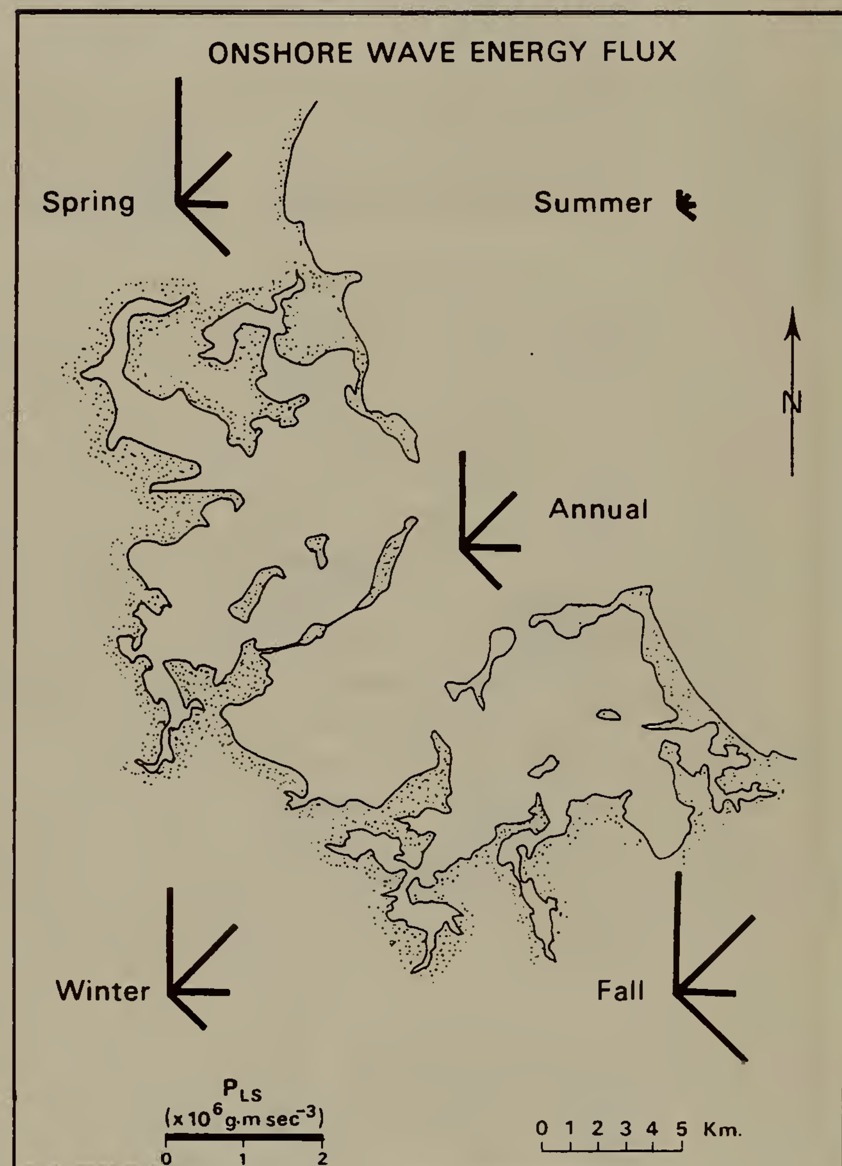


Figure 4. Wave energy fluxes determined from SSMO data.

Figure 5. 1739 Historical sketch of the Winthrop and Revere shorelines.

bathymetry coupled with the presence of numerous coastal structures produce complicated and variable patterns of littoral transport.

SHIRLEY GUT (Stop #1)

As seen in the historical map in Figure 5 the region between Point Shirley and Deer Island was once the site of a tidal inlet known as Shirley Gut. Evidence of the former inlet is seen in the fathometer profiles taken offshore of Yirrell Beach (Fig. 6). Note that the two seaward profiles contain a 2 to 3m scarp adjacent to Deer Island. This bank was likely cut by ebb tidal currents issuing out of the inlet. Evidence of the inlet channel along the inner profiles is probably masked by a sediment cover.

Shirley Gut closed in 1934. Leading to that time the inlet had gradually shoaled and narrowed due to spit growth on both sides of the inlet. Cross sectional changes of the channel and historical changes of the inlet shoreline are depicted in Figures 7 and 8. It is likely that the inlet closed as a result of storm processes (Fig. 9). Under normal conditions the inlet was probably stable because tidal scour would have been sufficient to erode any sediment dumped into the channel by wave-generated currents. However, during storms larger waves would have dramatically increased the transport of sediment toward the inlet along Point Shirley and Deer Island. Although some of this sediment would have been removed by the strong tidal currents accompaning the storm surge, much of the sediment would have remained. The reason for this is that during storms most of the increased flow into and out of Boston Harbor was accomodated through the much larger Presidents Roads channel. Thus, Shirley Gut was gradually filled in during storms due to a greater amount of sediment being delivered to the inlet than the volume of sediment that could be removed by the tidal currents.

Another factor that contributed to the closure of the inlet was its location at the center of the embayment. At this site, sediment was transported toward the inlet from both longshore directions. Note that the orthogonals in the wave refraction diagram in Figure 10 show that northeast storm waves would move sediment toward the former Shirley Gut location. This pattern of sedimentation has persisted to the present time as evidenced by the large gravel washover that was deposited in this region during the February Blizzard of 1978 (Fig. 11, Location A). The gravel consisted of well rounded cobble-sized material and was deposited over a previous gravel ridge and washover complex. The source of the gravel is believed to have come from the intertidal gravel terrace next to Deer Island (Fig. 11, Location B). Although offshore gravel may have been a possible source, a grain size distribution map of the region (Fig. 12) indicates that the sediment seaward of washover area is almost entirely sand.

The washover was mapped by a theodolite survey and seven beach profiles (Fig. 13). Its topography was ridge like, which is typical morphology of post storm gravel beaches (Lewis, 1931). The deposit measured 200m long, 50-60m wide and 70-80cm thick. Its depth was determined from three trenches dug along profile SP-6 (Fig. 14). The bottom of washover was estimated by the presence of dead weeds in growth position, a layer of oxidized gravel and concentrations of wood and debris.

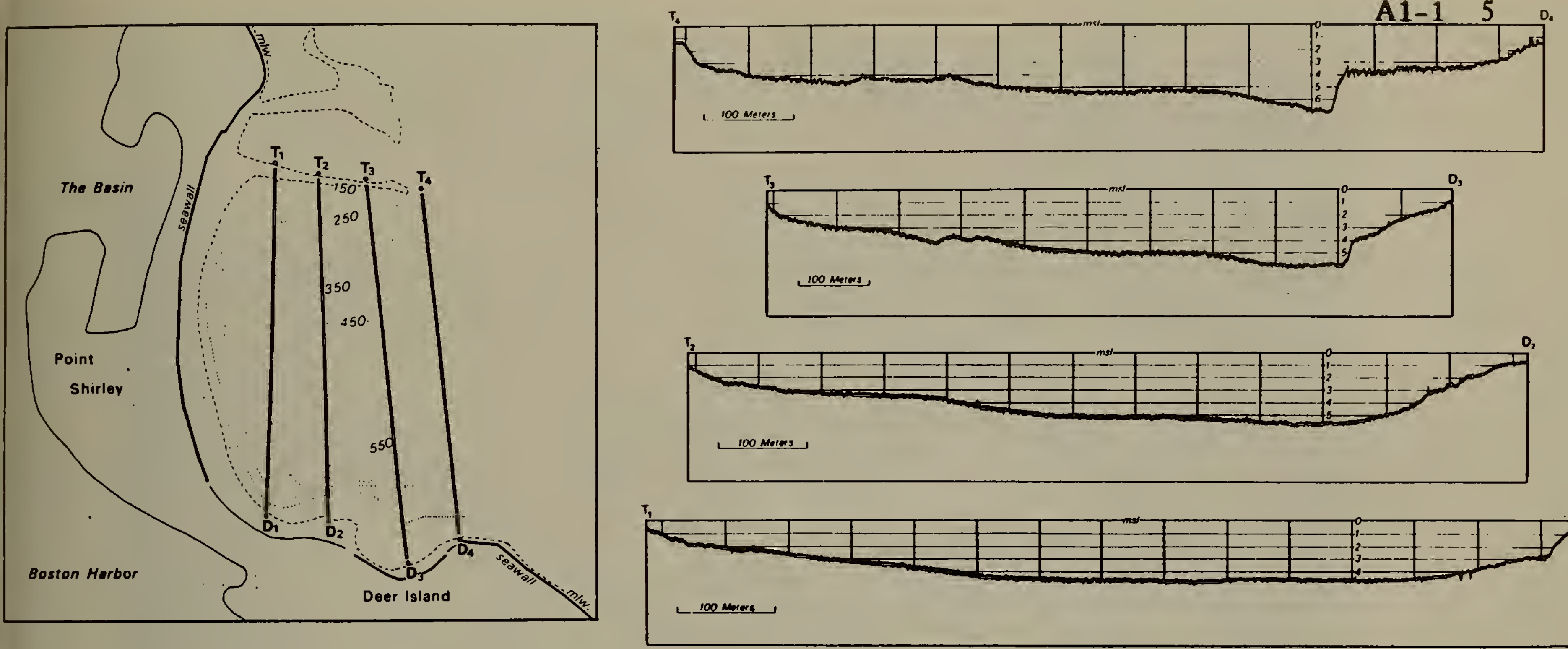


Figure 6. Fathometer profiles off Yirrell Beach.

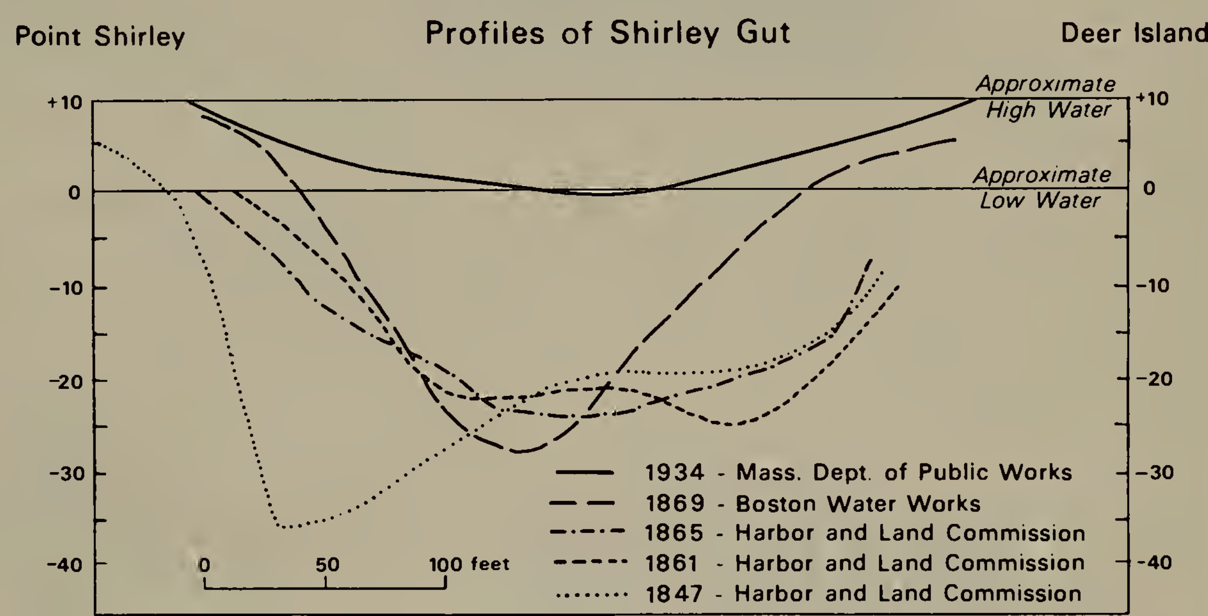


Figure 7. Historical cross sectional profiles of Shirley Gut.

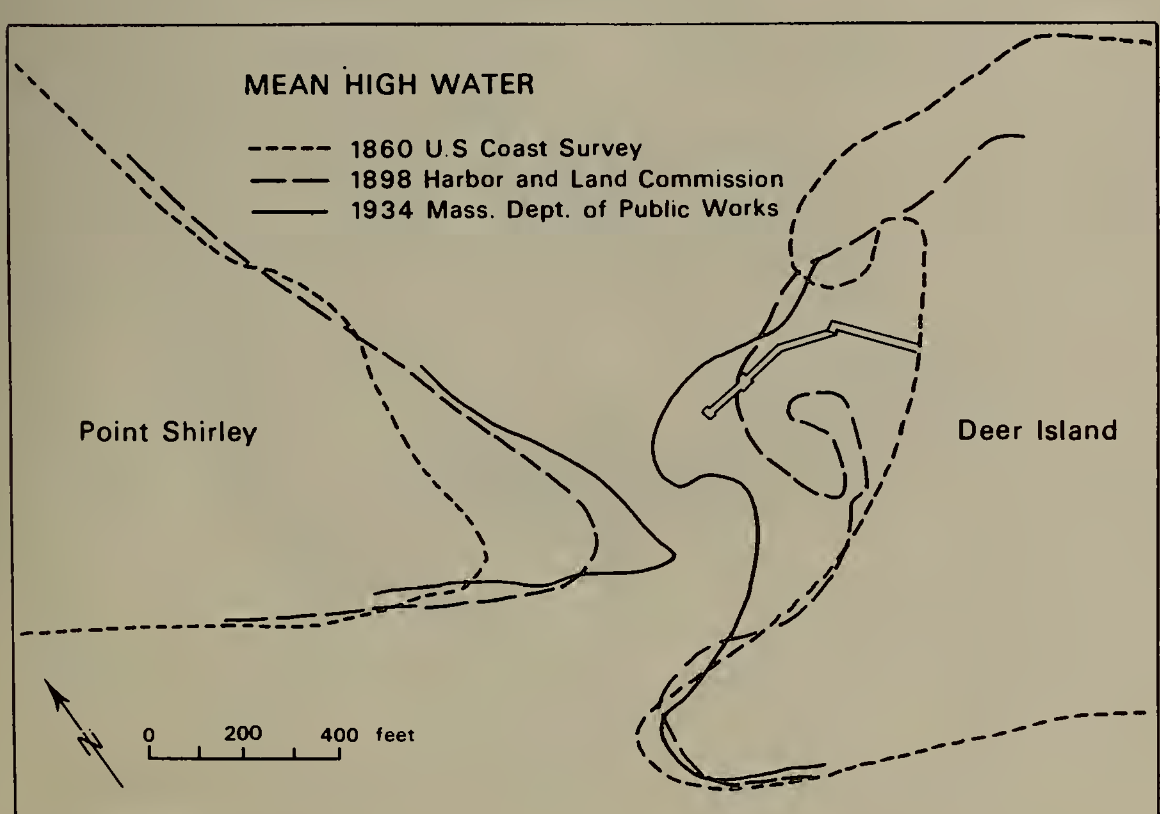


Figure 8. Historical shoreline changes of Shirley Gut.

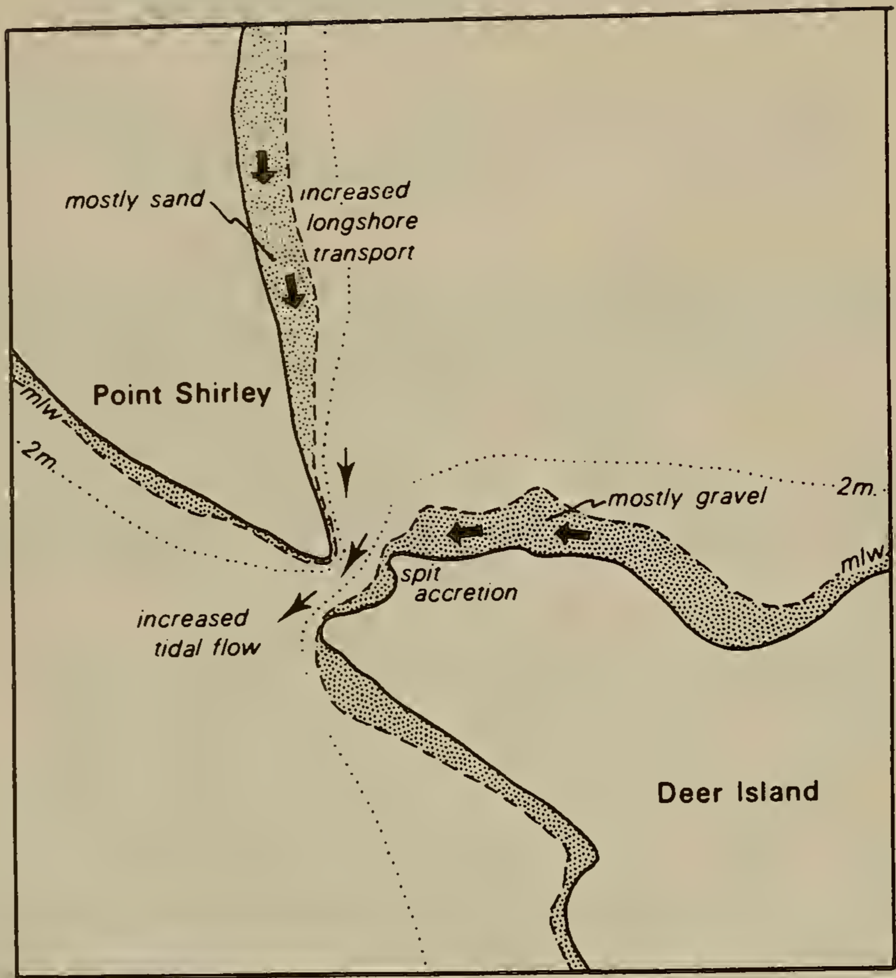


Figure 9. Storm-generated sediment transport patterns and processes at previously opened Shirley Gut.

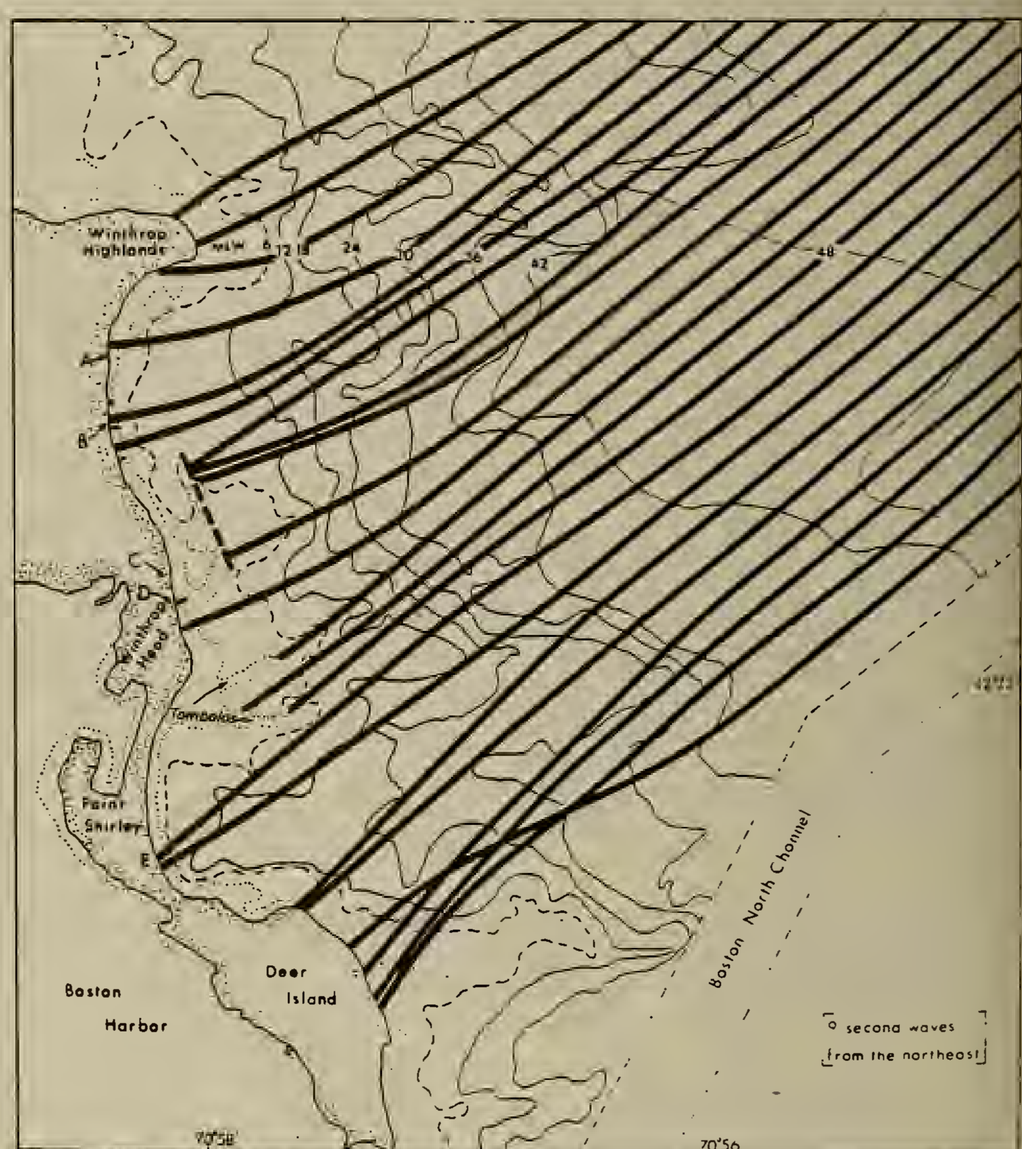


Figure 10. Storm wave refraction diagram for the Winthrop shoreline using 9 sec. waves from the Northeast.

Figure 11. Oblique aerial photograph of sourthern Point Shirley region.
Location A - gravel washover area
Location B - intertidal gravel platform

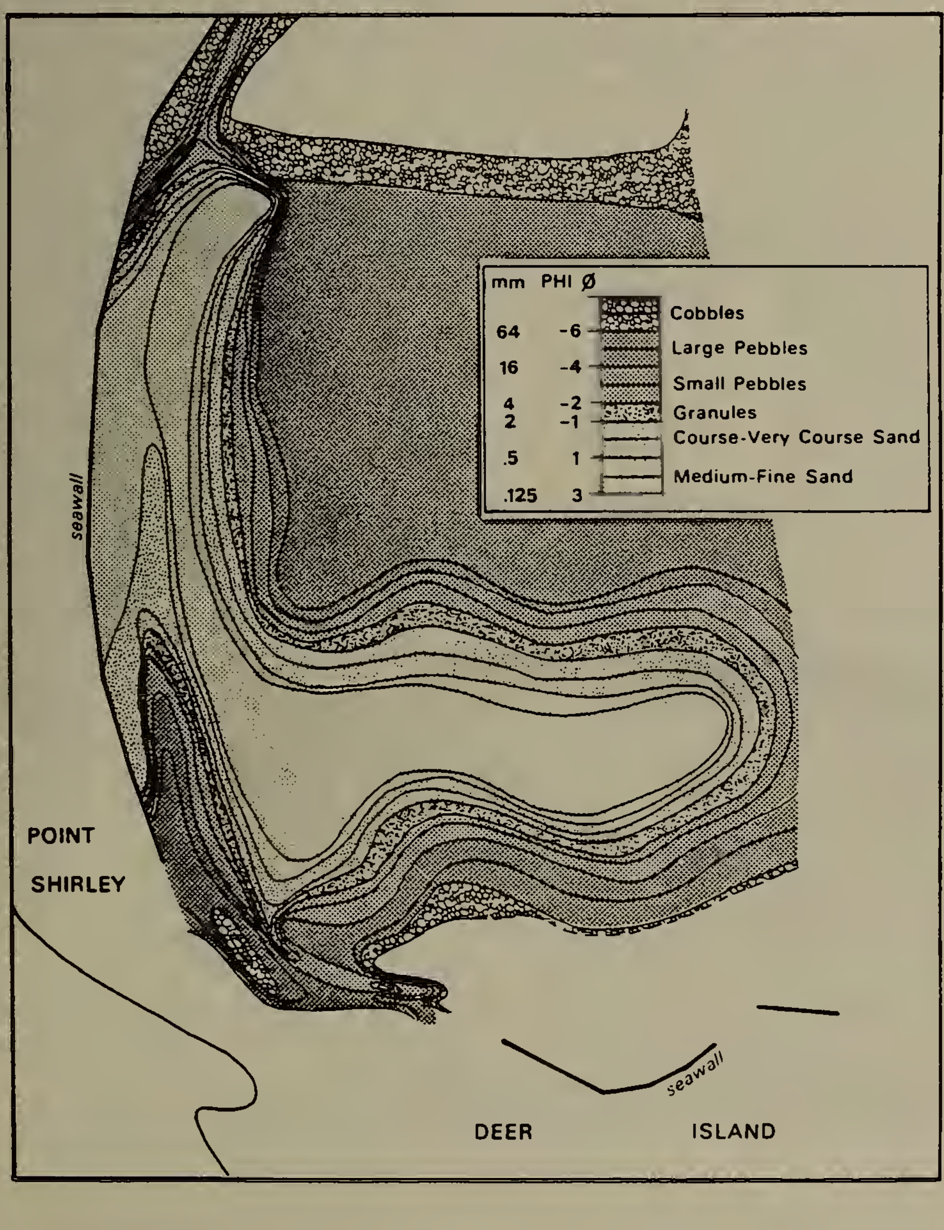


Figure 12. Grain size distribution map of Yirrell Beach and the offshore region. The san offshore of Point Shirley was likely deposited by the ebb tidal currents associated with the previously active Shirley Gut.

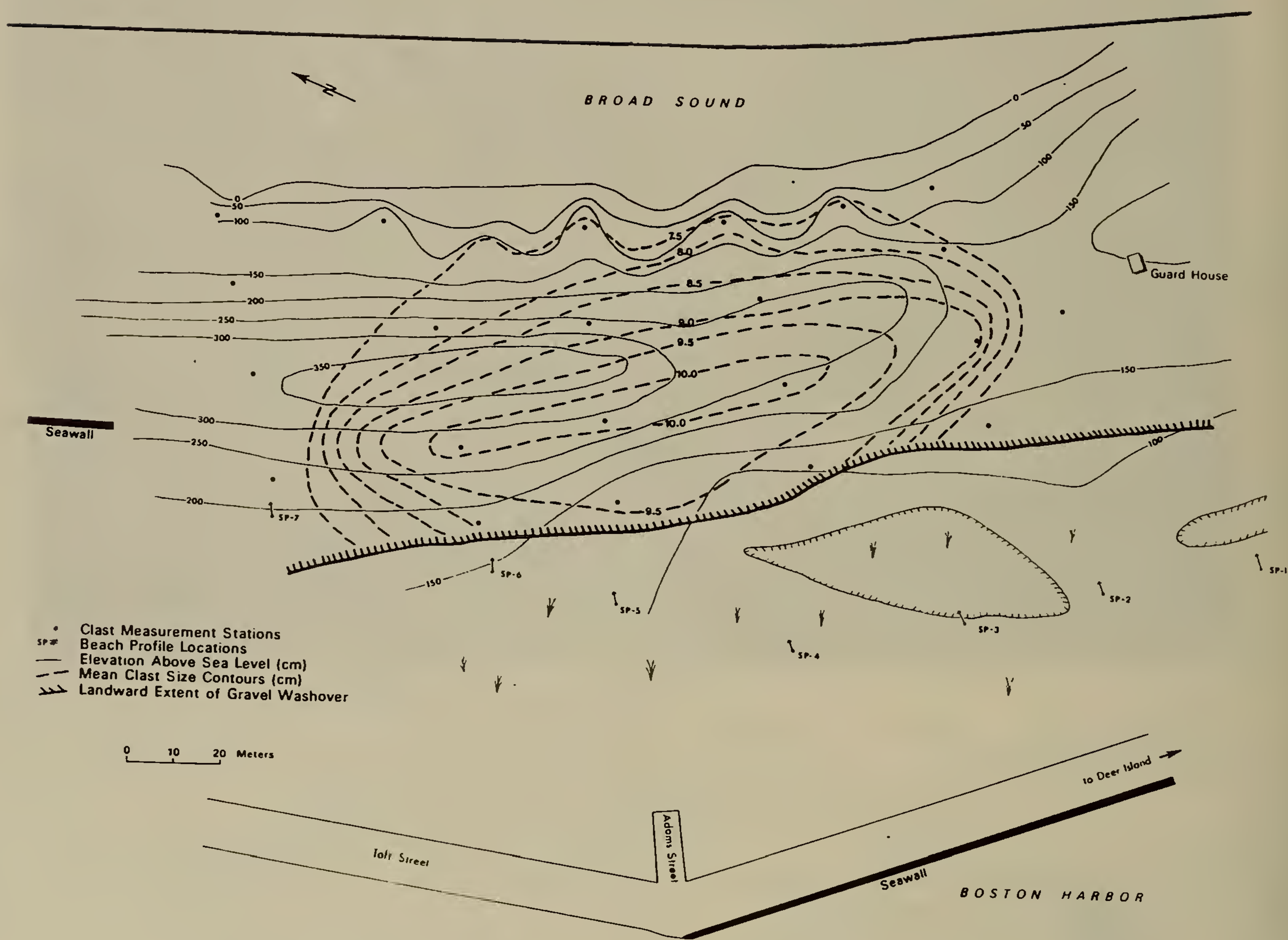


Figure 13. Topographic map and grain size distribution of gravel washover. Note that clast size increases in a landward direction.

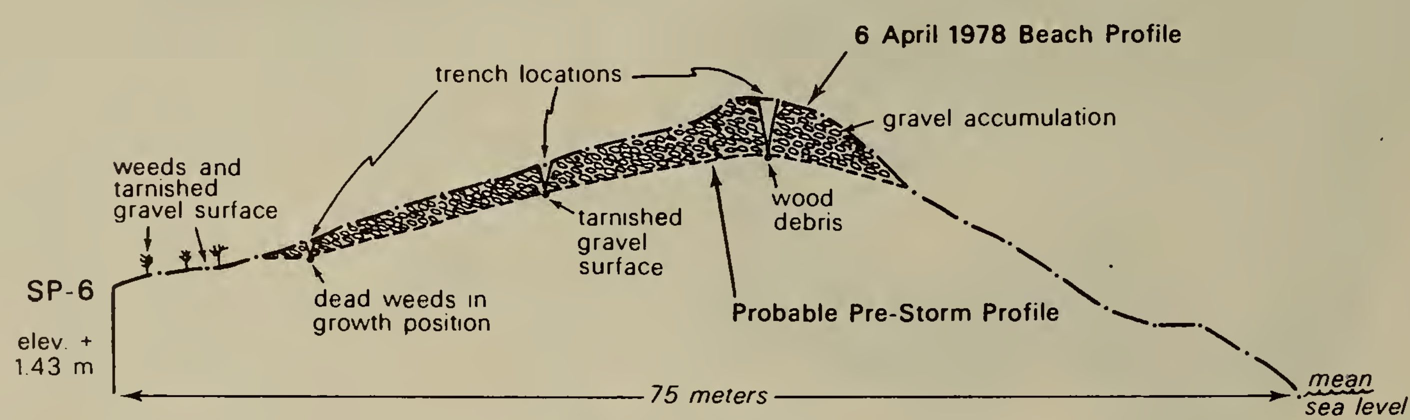


Figure 14. Trench through gravel washover.

The gravel was well imbricated. Long axis clast measurements at 22 stations showed a range in mean clast size of 3.2 to 10.2cm. Gravel shapes consisted mostly of disks and blades, with a few rods, but almost no spheres. This distribution is predicted by Bluck (1967) and also reported for a nearby beach by Brenninkmeyer (1976). Generally, the size of gravel clasts increased in landward direction (Fig. 13). This supprising trend suggests that during intense storm conditions when wave energy is sufficient to move all size fractions on the beach, sediment transport, to a degree, is more a function of clast cross sectional area than weight or fall velocity.

YIRRELL BEACH (Stop #2)

Yirrell Beach is part of a low narrow barrier that connects Winthrop Head to Point Shirley. Prior to the building of numerous summer cottages in the area, Yirrell Beach was backed by a vegetated dune system (Fig. 15). With the development of this region came the need for protection from winter storm damage (Fig. 16). In the 1950's a seawall was constructed along the entire length of the beach. The seawall is normally 1.0 to 1.5m above the level of the sand.

During the Blizzard of 1978 25,000m^3 of coarse sand was transported over the seawall, inundating houses and filling streets to a depth of a meter or more (Fig. 17). U.S. Army bulldozers and front-end loaders were brought in to remove the sand. The sand was placed in front of the seawall and a 600m long artificial dune ridge was constructed (Fig. 18). The source of the sand appears to be a cannibalization of the beach face sediments. A longshore source is unlikely, in that the entire shoreline experienced similar changes. An offshore source is also improbable. As illustrated in Figure 12 the beach sands extend only a short distance seaward of the depositional region.

Although no survey data exist to verify this explanation, beach profiles monitored during a much less severe northeast storm (25 January 1979) suggest a similar trend (Fig. 19). The profiles, which were taken before and immediately after the storm, showed that while the beach face retreated 1 to 6m, the upper berm accreted 10 to 60cm. Unlike sandy beaches that are backed by a foredune ridge and which develop a flat to concave upward profile during storms (Hayes and Boothroyd, 1969), the post storm profile at Yirrell Beach was consistently steeper than the pre-storm profile (Fig. 19). These changes suggest that during storms the beach face is eroded and most of the sediment is moved offshore. However, during the same period the upper beach face sands are transported onshore by wave swash and deposited next to the seawall (Fig. 20). If deposition next to the seawall continues for a long duration then sediment and water will be transported over the wall.

WINTHROP HEAD (Stop #3)

Winthrop Head is one of the many drumlins comprising the Boston Harbor shoreline (Fig. 3). It appears to be the remains of a larger drumlin that was partially destroyed by wave action (Kaye, 1967). It is likely that the greatest rate of drumlin recession occurred during storms when wave erosion caused undercutting and slumping of the adjacent scarp (Fig. 21). During the late 1800's there was a

Figure 15. A view of Yirrell Beach and Deer Island from Winthrop Head (from White, 1893).

Figure 16. Ground photograph taken during the 25 January 1979 northeast storm. Waves are eroding the beach face and depositing some of the sediment next to the seawall.

Figure 17. February Blizzard of 1978 caused 25000m^3 of sand to be transported over the seawall along Yirrell Beach.

Figure 18. An artificial dune ridge was constructed from the sand removed from behind the seawall along Yirrell Beach after the February Blizzard.

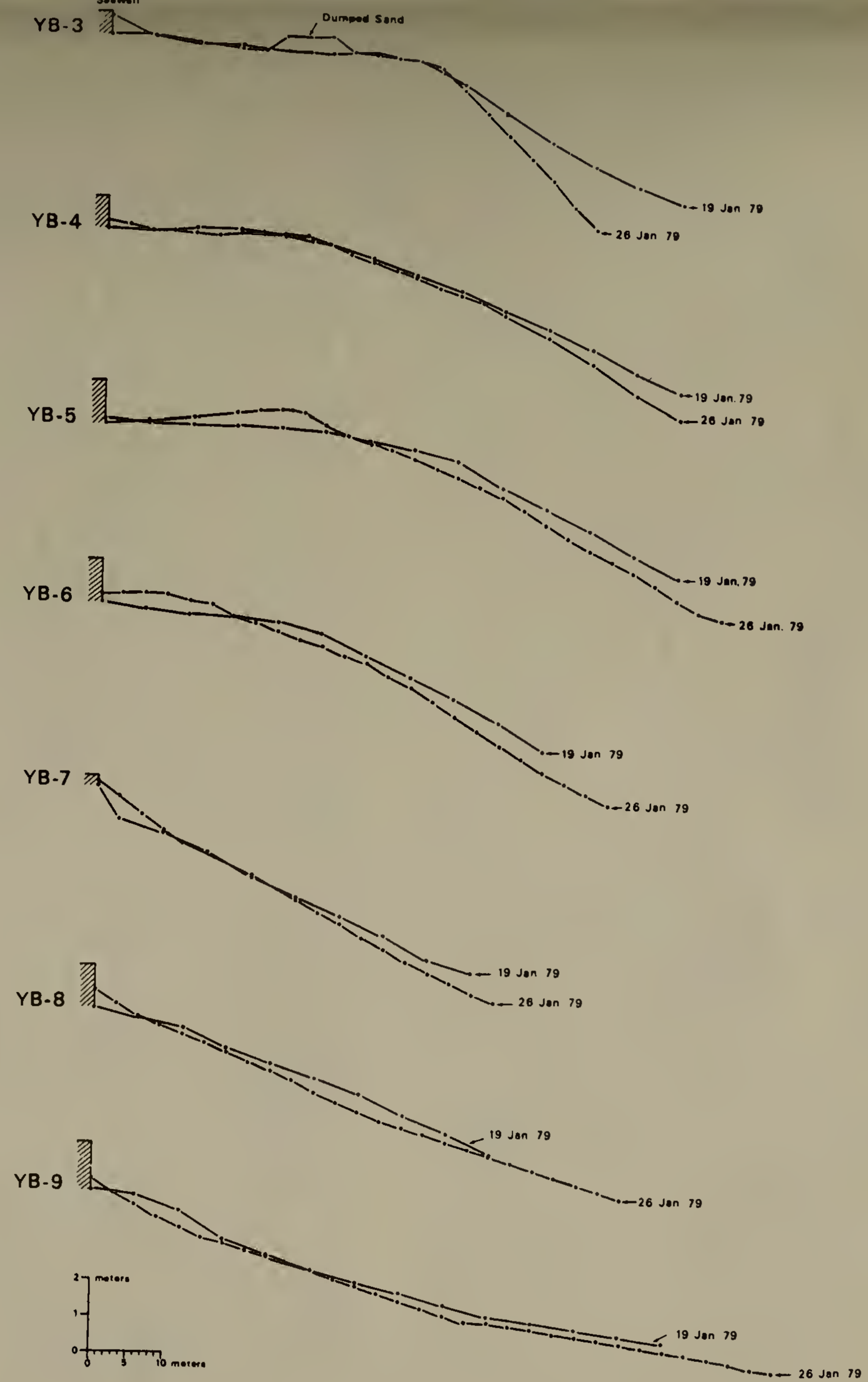


Figure 19. Beach profiles taken along Yirrell Beach before and after a northeast storm. Note the erosion of the beach face and deposition next to the seawall.

Figure 20. Sketch of beach processes active during a storm along Yirrell Beach (Bill Thoen).

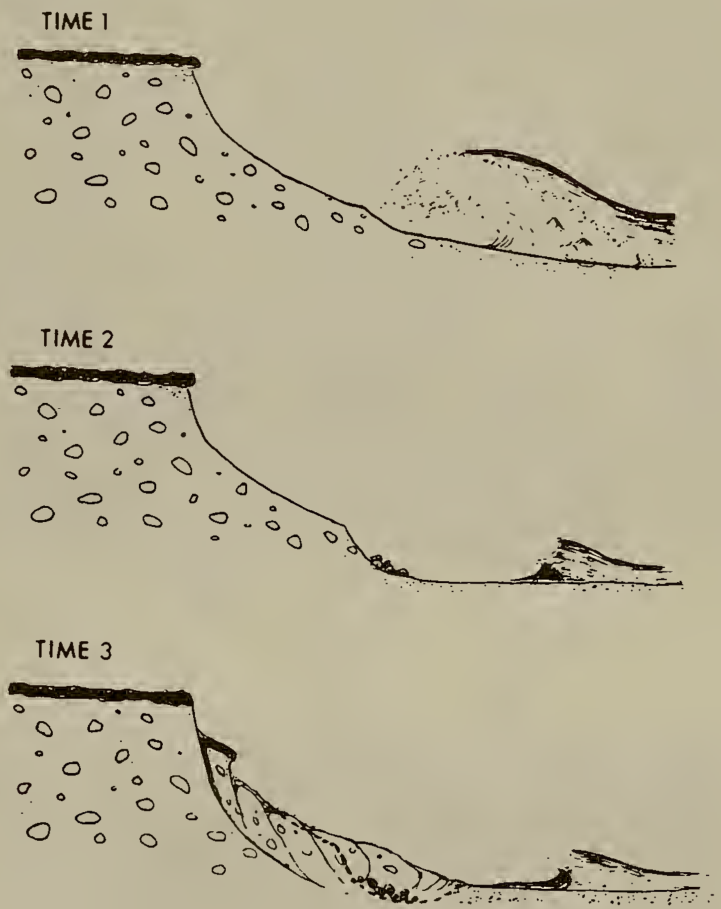
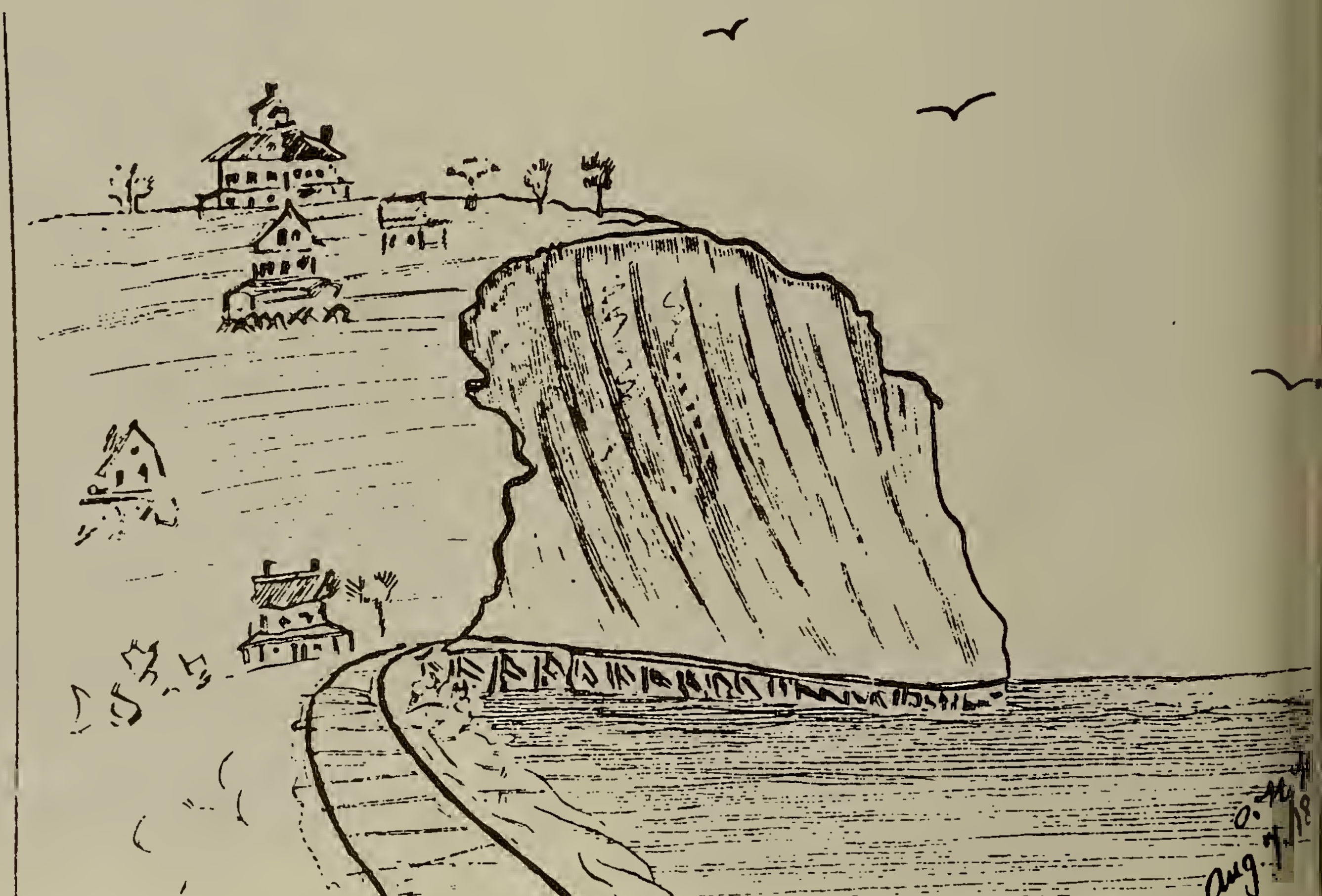


Figure 21. Wave erosion of a drumlin (sketch by Bill Thoen).

Figure 22. Sketch of railroad trestle in front of Winthrop Head, 1883.



railroad trestle in front of the drumlin (Fig. 22), however, it was demolished during a northeast storm in 1885. In 1927 a 3m high seawall was constructed to prevent further erosion (Fig. 23). However, these attempts have not been altogether successful.

The drumlin is the highest point in the town (31m) and thus is the site of a large standpipe. When the standpipe was built in 1909, it was located about 26m from the edge of the scarp. During the past 75 years the scarp has retreated more than 9m. At the present rate, if no measures are taken to protect the slope, the standpipe will be at the edge of the scarp in the year 2117. Measurements taken over the past four years indicate that this erosion rate is much too conserative. Also, it should be emphasized that as the scarp moves closer to the standpipe, the overburden of 5000 tons can be expected to cause a failure of the slope much sooner than orginally predicted.

Seaward of Winthrop Head is a double tombolo system that extends approximately 500m offshore (Fig. 24). The tombolos consist primarily of cobble-sized gravel which is believed to have been derived from an offshore eroded drumlin. The southern tombolo is present in the oldest U.S. Coast and Geodetic Survey charts dating to 1857. However, the northern tombolo was first documented in a 1945 vertical aerial photograph. A map of the morphological changes of the bars between 1945 and 1978 is shown in Figure 25. Note that during this 33 year period of time the ends of the tombolos were progressively flattened and moved onshore. Today the bars are joined at seaward ends and the northern tombolo has a spur that extends far to the northwest.

SOUTHERN WINTHROP BEACH (Stop #4)

The southern end of Winthrop Beach is comprised of sand and gravel and is characterized by a narrow or nonexistent high tide beach (Fig. 2). The region is backed by seawalls of various constructions (Fig. 26) and contains a number of groins. It is interesting to observe the riprap on both sides of the southern most groin. The 1 to 2 ton stone on the southern side of the groin has been there for about 30 years and is moderately rounded. The riprap on the northern side was placed there in 1980 and still retains many highly angular edges.

Riprap was put along the base of the seawall at Winthrop Beach to protect the wall during storms and to prevent wave reflection. Storm waves entrain large amounts of gravel which are propelled against the wall during each breaking wave. The riprap therefore abrades instead of the wall.

When waves break directly on the seawall much of the wave energy is reflected and propagated back offshore to meet the next incoming wave. This results in increased turbulence and offshore sediment transport (Fig. 27) (Herbick and Ko, 1968). Riprap in front of seawalls causes waves to break thereby reducing the reflection process. Many of the Boston area beaches have had seawalls along their shorelines for over 50 years. In these areas the shoreward erosion has been halted but vertical erosion has resulted in little or no high tide beach.

Beach profiles monitored in the southern Winthrop Beach region (Fig. 1, Location WB-1-3) during the winter of 1980 have revealed

Figure 23. Late 1920's photograph of Winthrop Head. Note the seawall that has been constructed of the eroding drumlin.

Figure 24. Oblique aerial photograph of the double tombolo system off Winthrop Head.

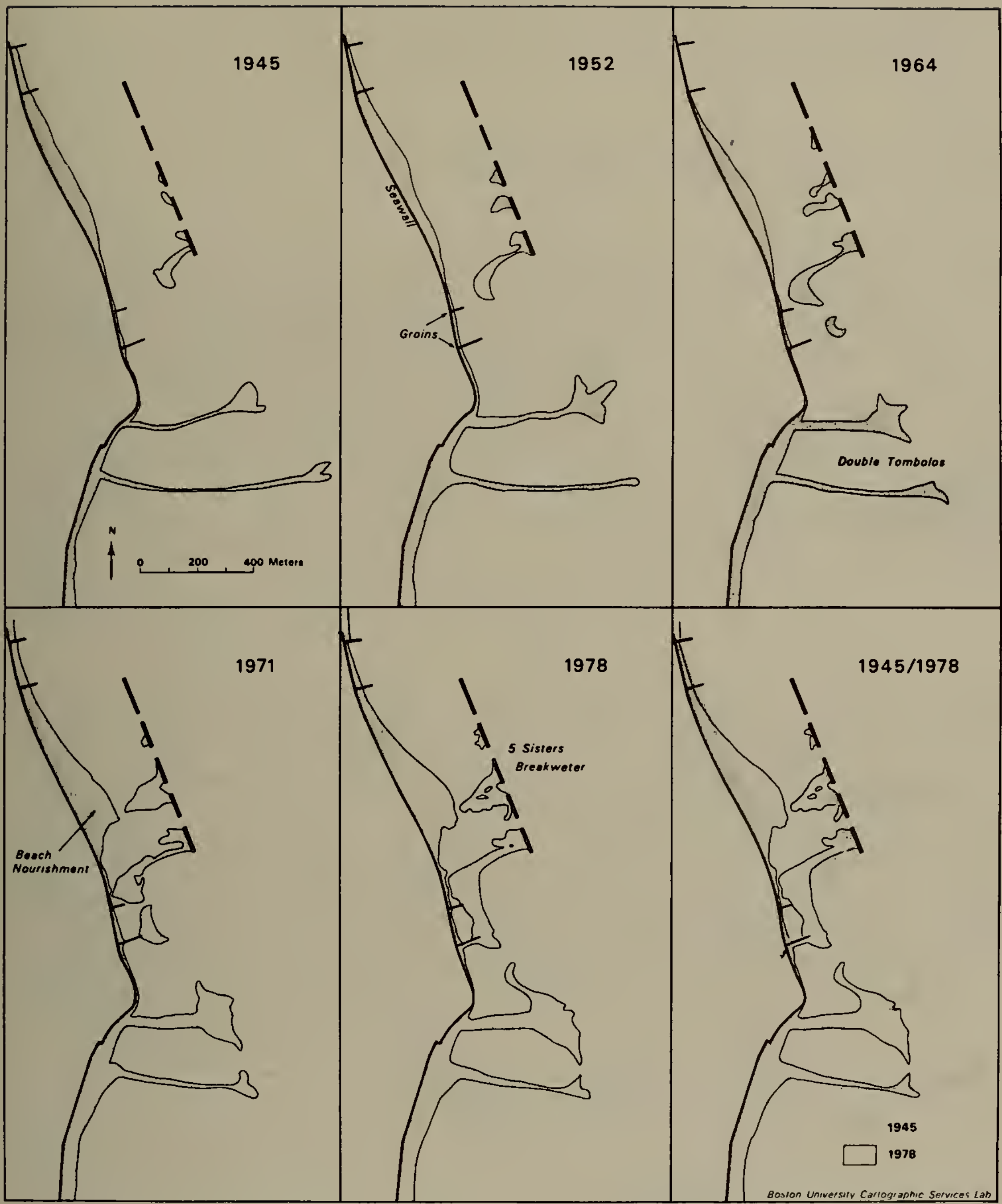


Figure 25. Historical morphological changes of the double tombolo system and gravel bars at the Five Sisters Breakwater.

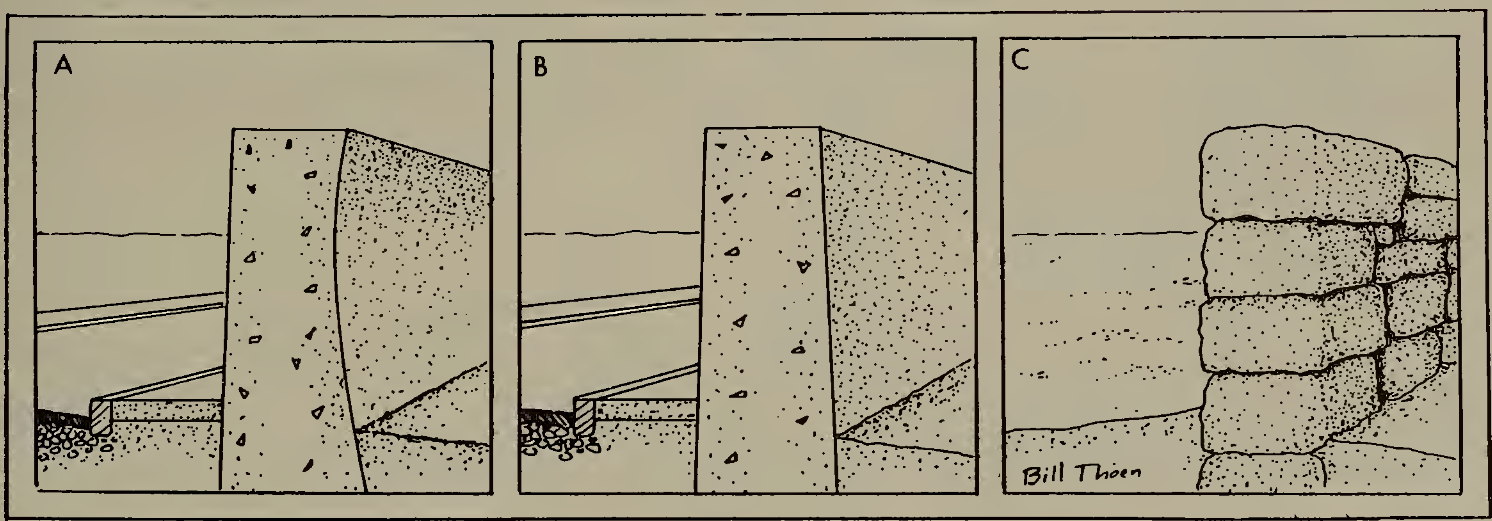


Figure 26. Sketch of various seawall constructions along Winthrop and Yirrell Beaches (by Bill Thoen).

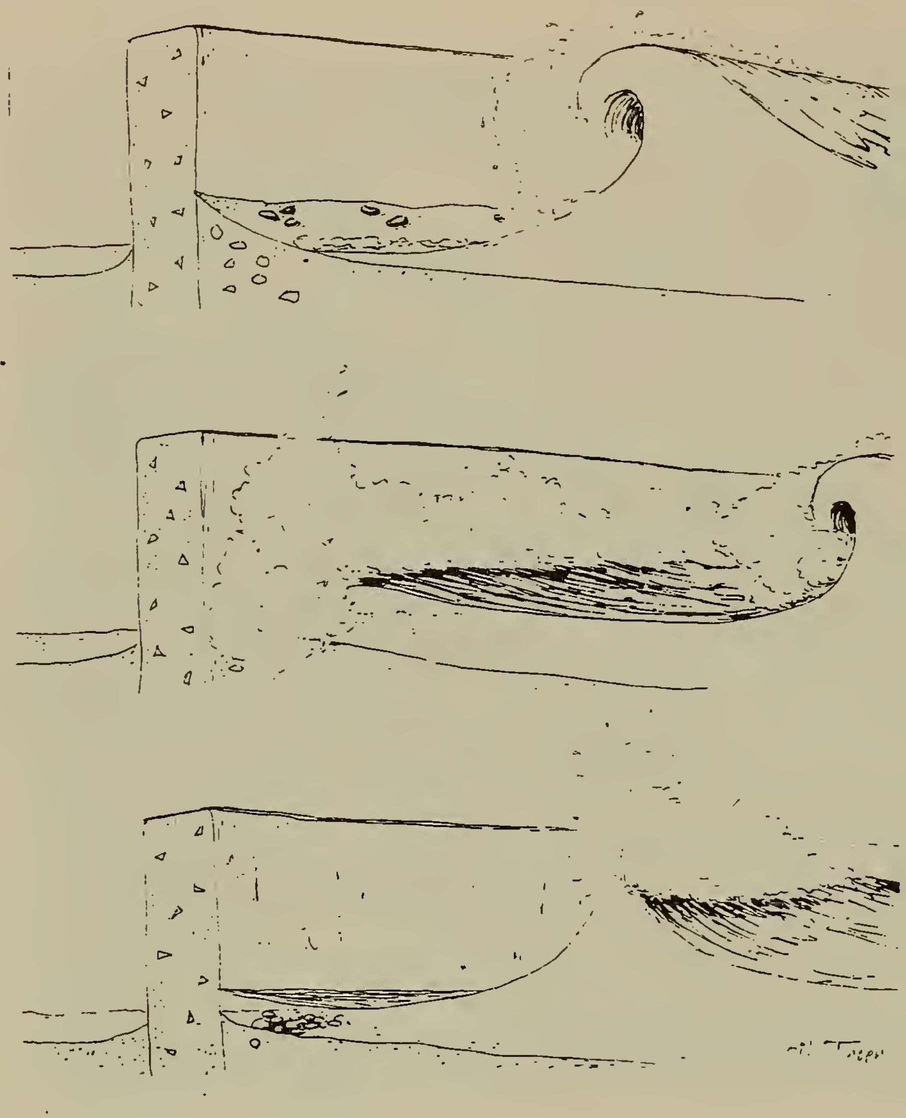
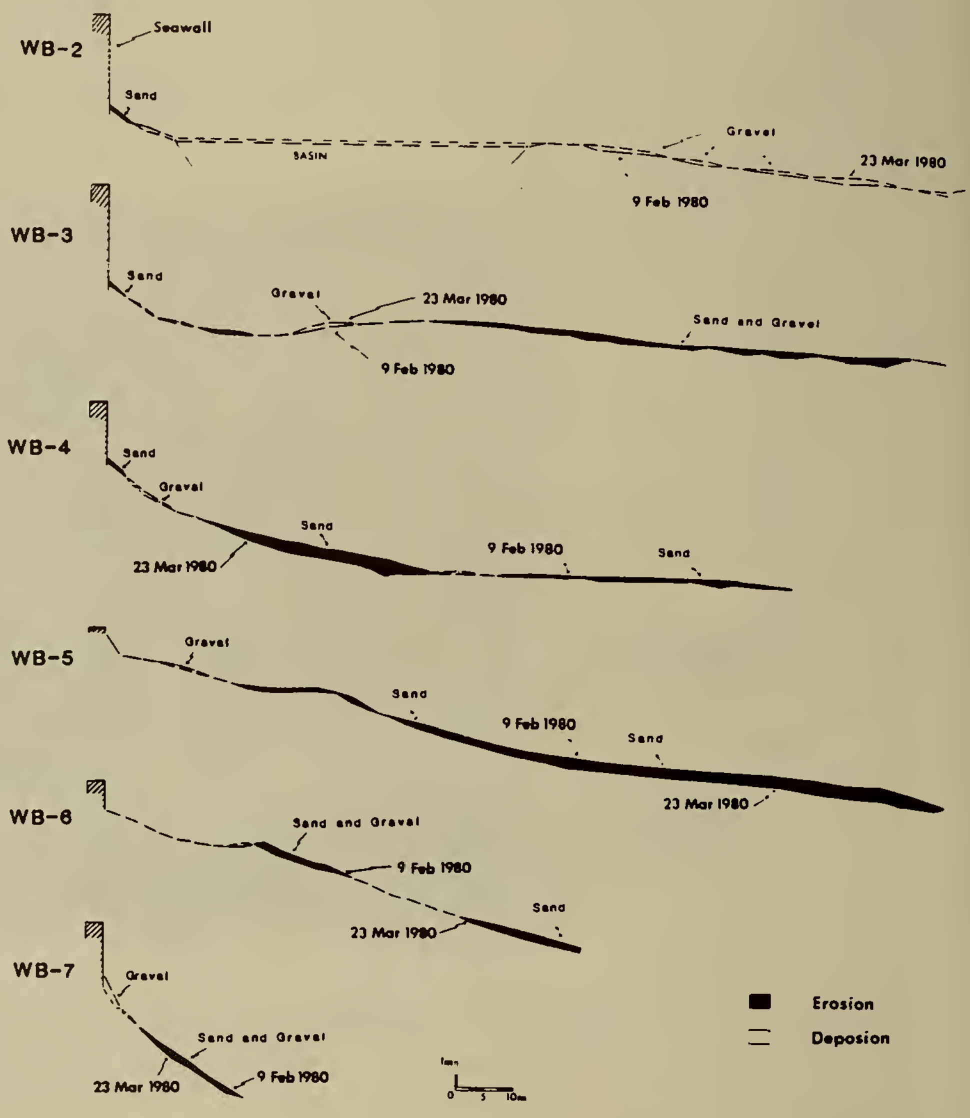


Figure 27. Process of wave reflection results in seaward transport of sand and vertical erosion next to seawalls (sketch by Bill Thoen).

Figure 28. Profile changes along Winthrop Beach due to a northeast storm. Note that sand is eroded while gravel is deposited.



two basic sedimentation trends (Fig. 28). During storm conditions sand is moved offshore while gravel is transported onshore. The only gravel locations that experienced erosion were areas in which wave reflection was an active process.

FIVE SISTERS BREAKWATER (Stop #5)

The Five Sisters Breakwater is located 300m offshore of the middle of Winthrop Beach (Fig. 2). It extends along a 700m stretch of shoreline and is built of piled granite blocks rising to a height of 3-4m above mean high water. Prior to its construction the shoreline behind the breakwater was eroding at a rate of $1250m^3$/yr. After completion of the project the beach accreted at a rate of $3000m^3$/yr. The deposition of sand in this area is attributed to wave refraction causing the longshore transport of sand toward the sheltered area behind the structures. This location has also been the site of sand nouishment.

Seaward of the breakwater is an eroded drumlin that rises to within 1.0m of mean low water. The gravel component of the drumlin has been transported onshore and deposited between sections of the breakwater. As illustrated in Figure 25 there have been substantial gravel bars that have formed behind the structures since 1945. Wave surge measurements made in the openings of the breakwater during low wave energy conditions ($H < 1.0m$) recorded velocities up to 30cm/sec (Fig. 29) (Sullivan, 1982). During storms, when waves are much larger, surge velocities would be much greater and more than adequate to move the gravel onshore.

NORTH WINTHROP BEACH (Stop #6)

The northern end of Winthrop Beach is highly structured with three groins and a massive seawall (Fig. 30). Riprap was placed two-thirds of the way up the height of the seawall in 1980. This amount of protection is neccessary due to the direct exposure of this region to wave energy. A beach is absent in this area and the nearshore slopes steeply to >4.0m within 300m of the shoreline. Under normal conditions, waves break on the seawall from mid to high tide, but during storms waves assault the wall continuously.

During the February Blizzard of 1978 a 10m section of the seawall was dismantled, leading to an undermining and collapse of the adjacent sidewalk and roadway (Fig. 31). The wall that failed was built of large stacked granite blocks that originally, had been held together with cemented (Fig. 26). At the time of the storm much of the mortar had been remove through abrasion, frost wedging and wear. This allowed the storm waves to more easily pluck granite blocks from the wall.

In other areas the seawall is constructed of reinforced cement and has a curved face (Figs. 26 and 32). During intense storms, prior to the placement of the riprap, storm waves that broke higher up the wall were only partially reflected. Most the water within one of the breaking waves was propelled upward to heights as great as 15m above the top of the wall (Fig. 33).The occurrence of these water fountains was dependent on an elevated sea level, and they increased in height with increasing wave height. The vertical uprush of water entrained sand and gravel at the base of the seawall and

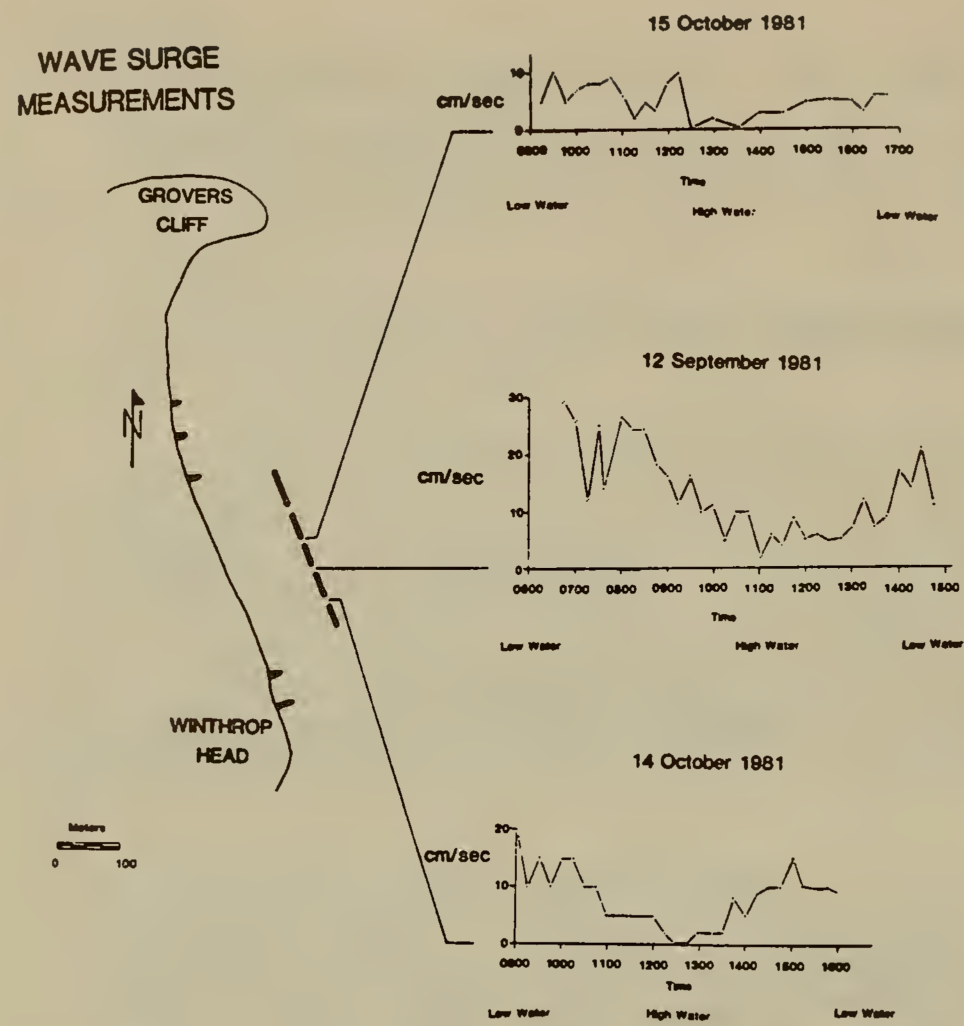


Figure 29. Wave surge current velocities measure between the sections of the Five Sisters Breakwater (from Sullivan, 1981).

Figure 30. Oblique aerial photograph of the northern portion of Winthrop Beach.

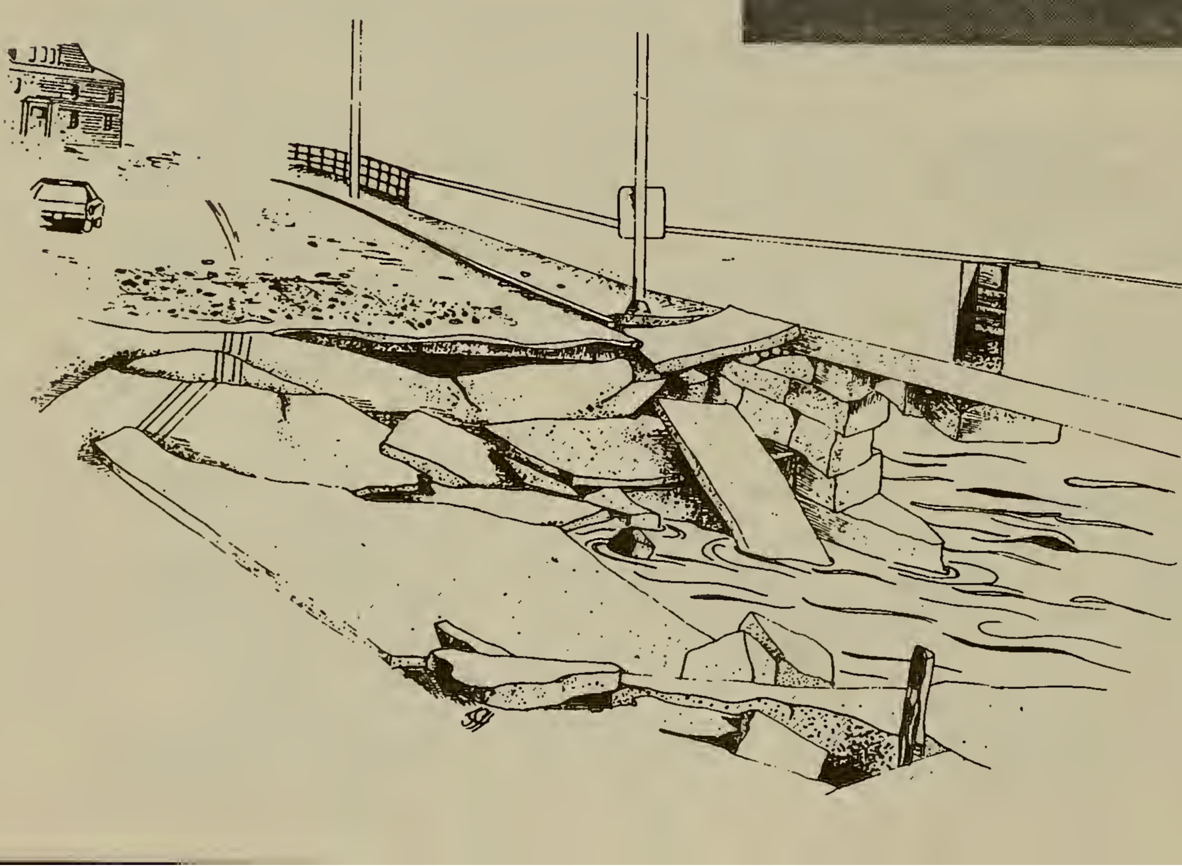


Figure 31. Sketch of the collapse of the sidewalk and roadway that occurred during the February Bizzard of 1978 at Northern Winthrop Beach (by Lindley Hanson).

propelled it skyward. Once above the top of the seawall the strong onshore winds, that accompany northeast storms, blew the water and sediment mixture onto the the adjacent street (Fig. 33). During the Blizzard of 1978 gravel projectiles that were transported over the seawall by this mechanism filled streets and ripped apart nearby houses.

To retard the water fountains and the ensuing gravel projectiles, riprap was placed against the wall in 1980 (Fig. 34). Prior to this time, deposition next to the seawall was prevented due to wave reflection processes. With reduced wave energy the onshore movement of gravel was allowed to accumulate around the riprap. The gravel slope that was formed by this process provided a ramp for water and gravel to be moved over the seawall during even moderate storm wave conditions (Figs. 35 and 36). Beach profiles shown in Figure 37 illustrate the changes to the beach. Thus, in this case, the riprap while protecting the seawall exacerbated the problem of transport over the wall.

CONCLUSIONS

A summary of the sedimentation processes along the Winthrop coast is illustrated in Figure 38. A consistent trend of storm generated sediment transport has been documented for this region. Gravel is moved onshore while sand is transported offshore. This pattern is corroborated by the gravel washover that occurred at Point Shirley, the historical flattening of the double tombolo system, the build up of gravel between sections of the breakwater and the accumulation of gravel around the riprap. It also has been documented with beach profile data that during storms sand is eroded from Winthrop and Yirrell Beaches. The large volume of sand that was transported over the seawall at Yirrell Beach during the February Blizzard of 1978 was likely attributable to beach face erosion and wave swash processes.

The riprapping of the seawall along northern Winthrop Beach that was done to prevent water fountains and gravel projectiles has had mixed results. Although the riprap has protected the wall from abrasion, a gravel ramp has been formed in the absence of wave reflection processes. This has led to more frequent occurrences of water and gravel being transported over the seawall.

ACKNOWLEDGEMENTS

The research along the Winthrop coast was funded by the Town of Winthrop and Massachusetts Coastal Zone Management. The author would like to acknowledge the individual studies and field assistance of the following Boston University graduate students: David Sullivan, Andrew Magee, Larry Oates, Doug Levin and Andrew Bakinowski. Dr. Denis D'Amore is thanked for his detailed investigation of the Winthrop Head drumlin. The sketches in this paper are the art work of Bill Thoen and Lindley Hanson. The figures were done by Eliza McClennan of the Boston University Cartographic Services.

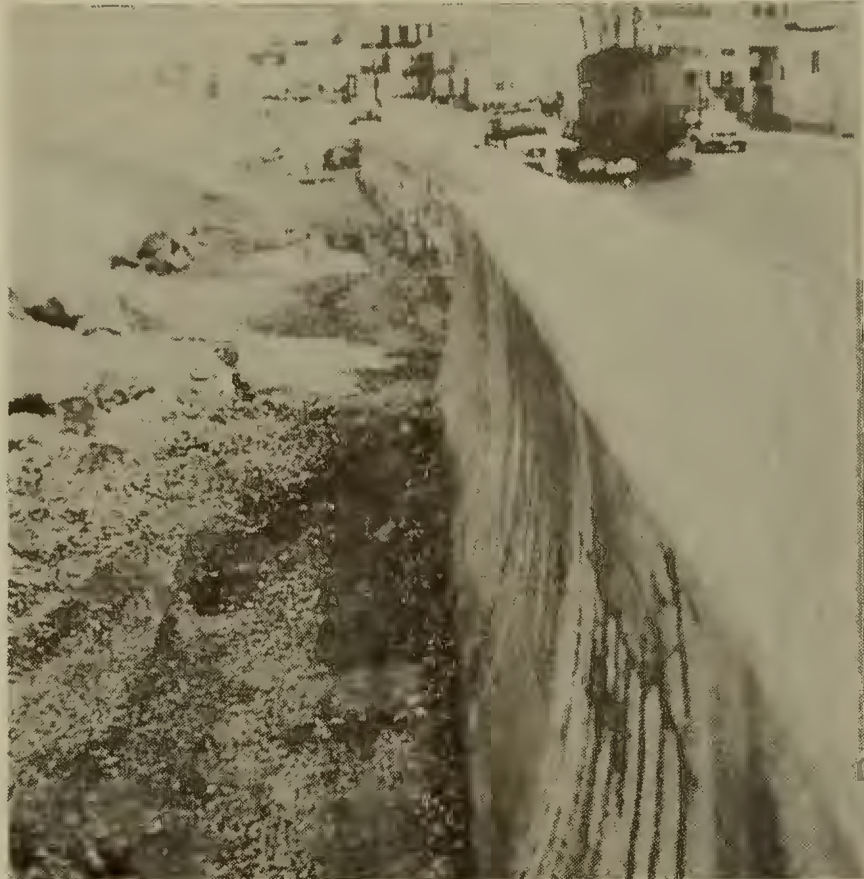


Figure 32. Cured-face seawall.

Figure 34. Riprap along northern Winthrop Beach. Photograph taken right after completion of project.

Figure 33. Water fountains along northern Winthrop Beach.

Figure 35. Gravel that had accumulated around riprap.

Figure 36. Photograph of water and gravel being transported over the seawall after the seawall had been riprapped.

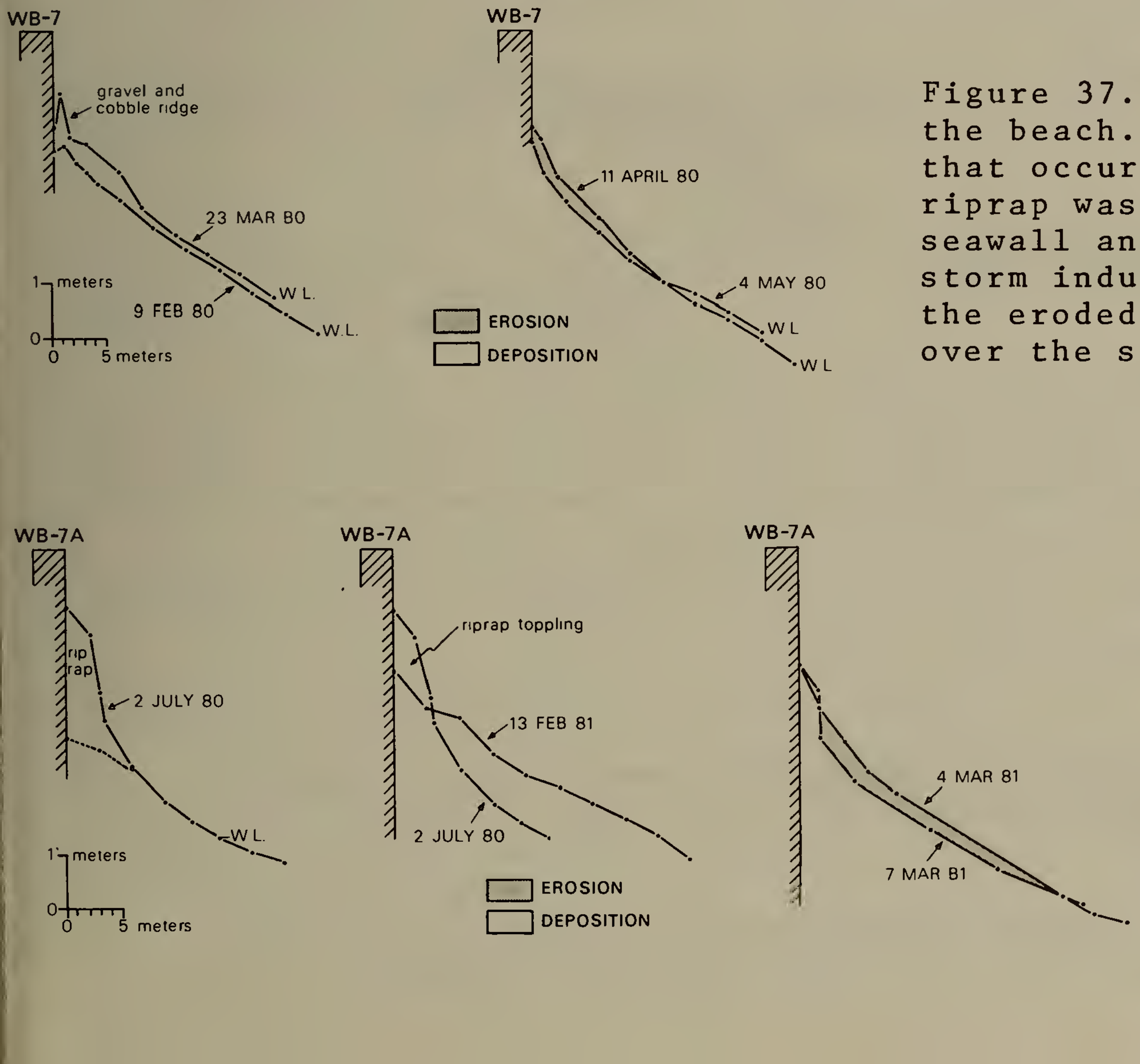


Figure 37. Profile changes to the beach. Note deposition that occurred after the initial riprap was placed along the seawall and the subsequent storm induced erosion. Some of the eroded gravel transported over the seawall.

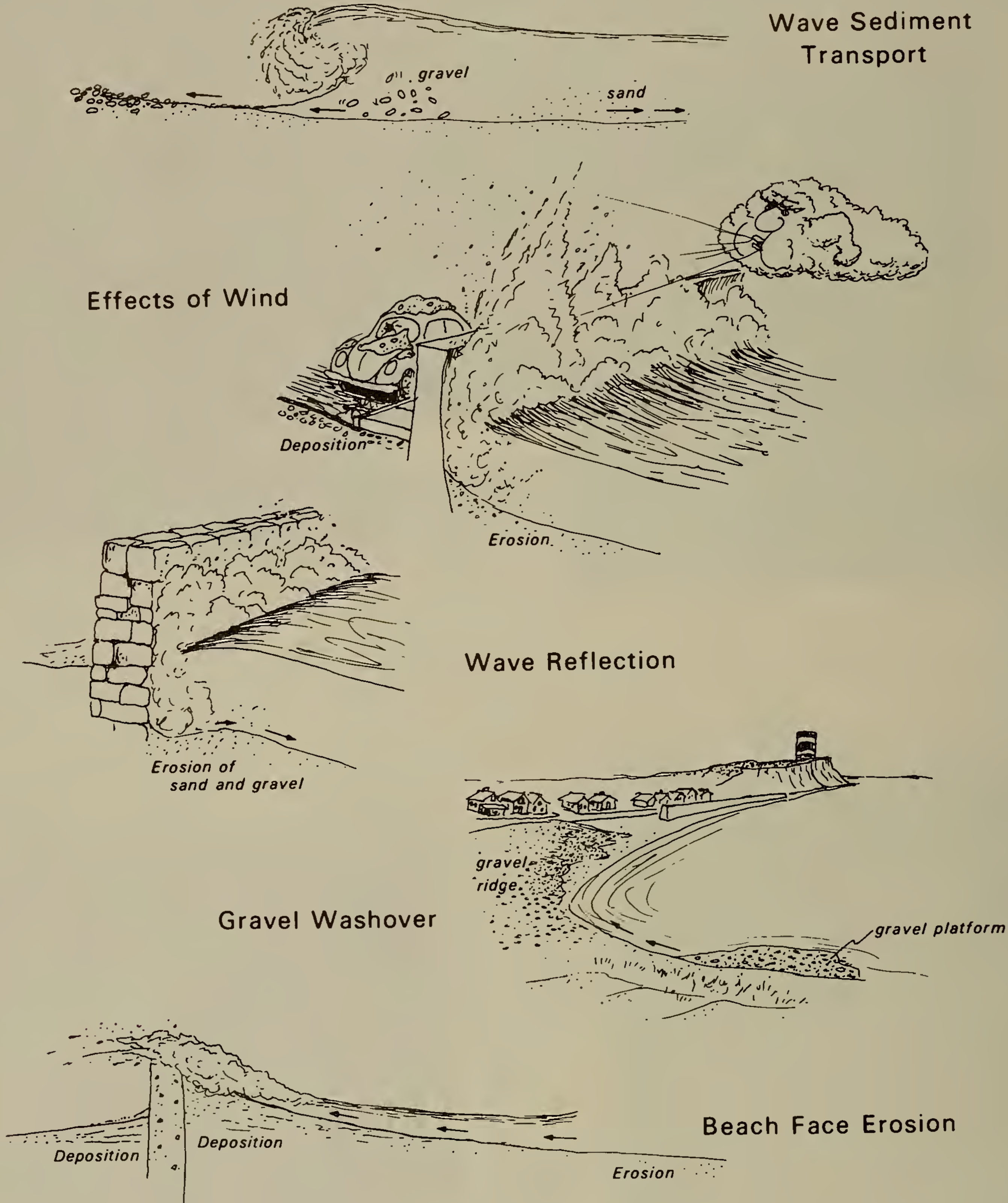


Figure 38. Summary of physical processes along Winthrop and Yirrell Beaches.

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BEACH LOG

The field trip will begin at the gate to the Deer Island Prison. Please don't check in. There are parking lots on the right and left. Please use the left parking lot (on the ocean side).

Kilometers

0.0 STOP #1 Shirley Gut.
Inlet filling, Gravel washover

0.7 STOP #2 Yirrell Beach
Storm processes, Beach management

1.5 STOP #3 Winthrop Head
Drumlin erosion, Tombolos

1.9 STOP #4 Southern Winthrop Beach
Coastal structures, gravel transport

2.2 STOP #5 Five Sisters Breakwater
Gravel bars, Wave surge processes

3.0 STOP #6 Northern Winthrop Beach
Riprap, Transport over the seawall

Ride by bus back to the Deer Island parking lot. Proceed to South Boston to take the ferry to Thompson Island.

GRAVEL SPIT PROCESSES, THOMPSON ISLAND, BOSTON HARBOR, MASSACHUSETTS

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Introduction

Thompson Island is one of nearly 30 islands in Boston Harobr created by submergence of Pleistocene drumlinoid topography (Figure 1). Modern wave processes in this low energy setting have reworked the eroding unsorted gravel to create a variety of accumulative landforms, including longshore spits, cuspate spits, and tombolos. Thompson Island has two cuspate spits and one longshore spit.

The longshore drift directions on the island are controlled by wave refraction patterns around the island. Oncoming waves refract around the island, so a drift convergence can form at the leeward side resulting in the formation of cuspate spits.

Boston Harbor mouth opens into Massachusetts Bay with a direct northeast exposure. However, Thompson Island is sheltered from the harbor mouth. Therefore, locally-generated seas can play a role in affecting net drift directions. Along the eastern shore of the island, net drift is toward both the north and south directions, representing influence from both the prevalent southwest and dominant northeast winds.

There is no onshore-offshore cycling of sediments on the beaches. The nearshore zone is composed of sandy silts derived from the fine fraction of eroding Pleistocene bluffs.

South Cuspate Spit

The South Cuspate Spit is one of the few accumulative forms in Boston Harbor with a high percentage of sand. Therefore, the emergence and preservation of the landform results from dune growth above the gravel beach. Figure 2 shows longshore currents converging at the spit in response to a northern wave approach. This figure also demonstrates that the cuspate form represents a reorientation of the shore into two opposing wave approach directions. A dearth of source material to the south (due to blocking by the South Longshore Spit), and abundance of source sediments to the north (due to an eroding, sandy bluff) results in gradual accumulation of the spit northward. This trend is recorded in a succession of preserved dune ridges parallel to the northern shore (Figure 3). There has been no discernable change in the shore position since 1847.

Boston Harbor

FIGURE 1

FIGURE 2

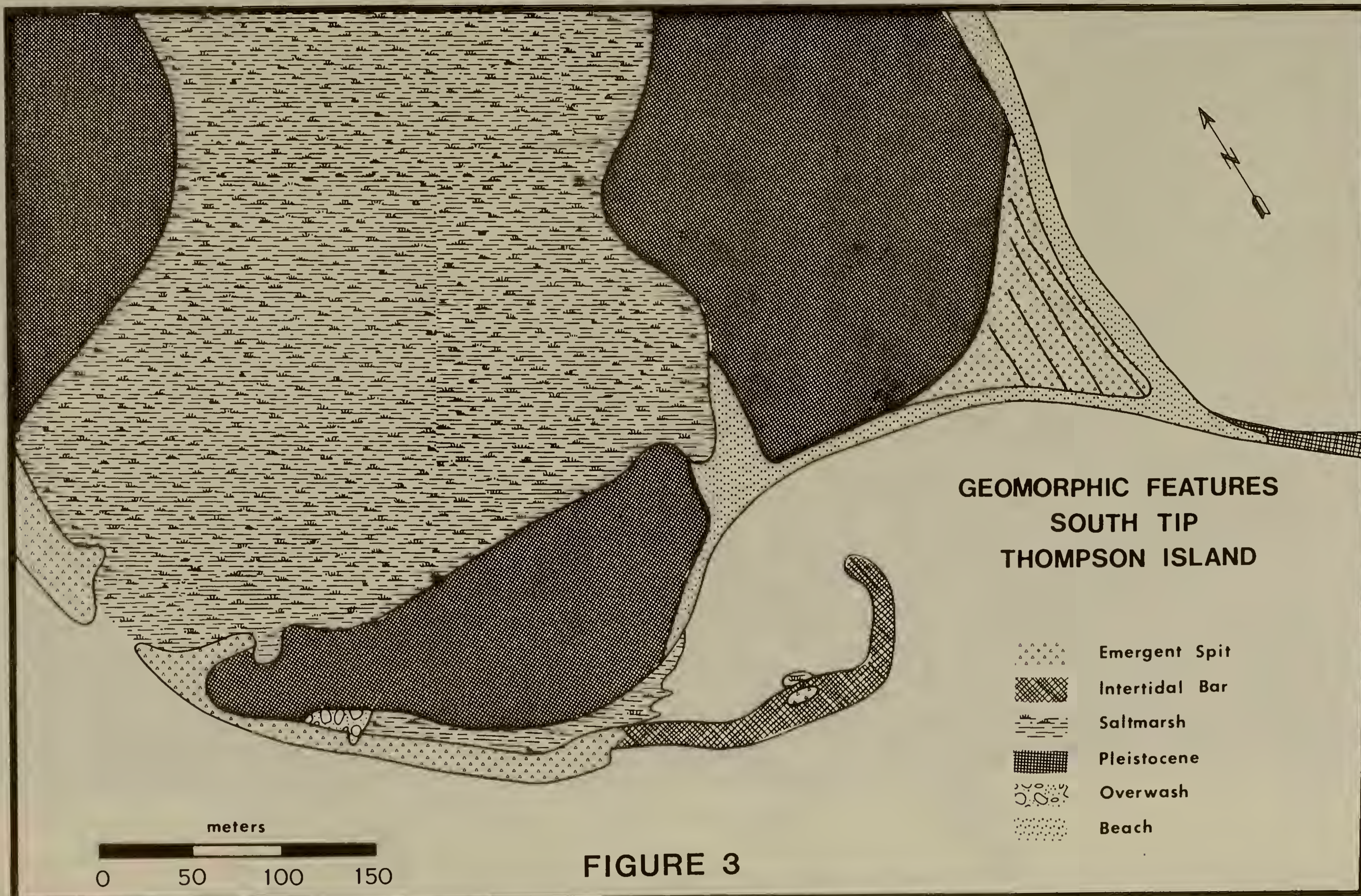

GEOMORPHIC FEATURES
SOUTH TIP
THOMPSON ISLAND
Emergent Spit
Intertidal Bar
Saltmarsh
Pleistocene
Overwash
Beach
meters
0
50
100
150
N

FIGURE 3

The spit ends in a bar that extends across a channel to Squantum Head. This bar represents the initial stages of tombolo formation. The bar has not emerged due to scouring by tidal currents. Since the bar has been relatively stable since 1847, it appears to represent an equilibrium between longshore sediment input and tidal scour.

North Cuspate Spit

The gravel comprising the North Cuspate Spit has a low sand content and no dune formation. The spit is greater than 2550 years old, based on Carbon-14 dates of the enclosed salt marsh (Clifford Kaye, USGS, personal communication). The spit has grown seaward on a base of nearshore muds (Figure 4).

The gravel has a high percentage of Cambridge Argillite pebbles, which may underly the glacial sediments in this area. This results in a dominance of flat (disc or blade) shapes. As is typical of gravel beaches, flat shapes are preferentially transported landward, while round (spheroids and rods) shapes are transported seaward. Storm wave activity results in net onshore transport and the formation of flat-pebble accumulative ridges above the high water line. The ridges form the emergent portions of the spit. The accumulative ridges have been deposited successively on the southern face of the spit, at least partially due to the abundance of source material from that direction. Tracings of the ridge crests from aerial photographs have shown that the spit was initially more parallel with the original shoreline of the submerged glacial materials (Figure 5). Successive storm ridge accumulation has reoriented the shore to a position nearly parallel to oncoming southerly waves.

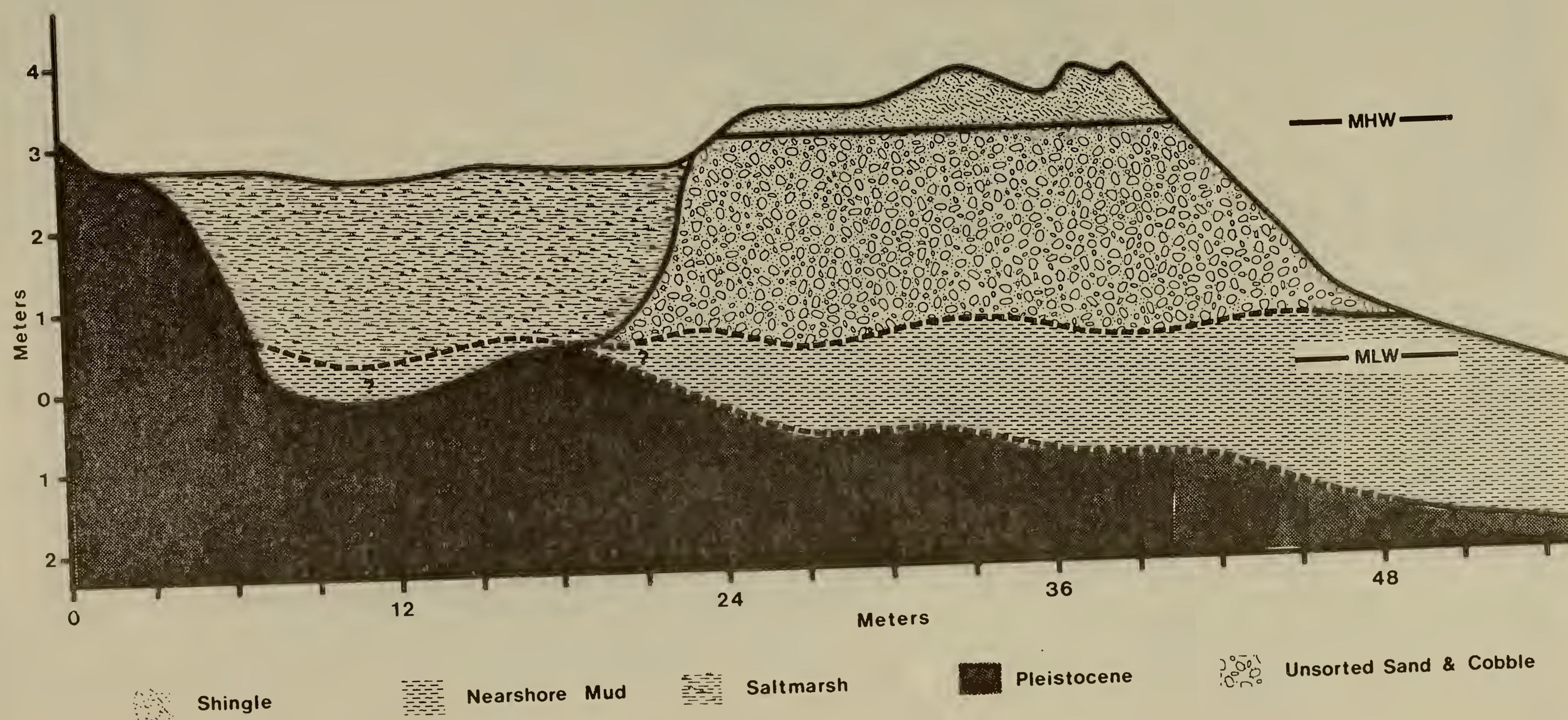
The north-facing shore of the spit truncates the gravel ridges, indicating longterm retreat. This shore is exposed to higher wave energy, and lacks a longshore sediment source. The backbeach in this area forms an overtopping ridge, or accumulation of gravel that is periodically overtopped by storm waves. The overtopping does not form distinct channels, so a landward-dipping slipface exists on the entire length of the ridge. Since the ridge is composed of flat pebbles, and currents during overtopping are primarily unidirectional, the internal geometry of the ridge is dominated by seaward-dipping imbricated pebbles.

The occurrence of accumulative ridges or overtopping ridges in the backshore corresponds to longterm shoreline changes. Since 1847, the north face has retreated at 15 cm/year, while the south face has advanced at 10 cm/year. Therefore, this landform is mobile and is migrating in a southerly direction. This mobility has been noted on other cuspate spits in the harbor, such as Bass Point, Long Island.

The spit encloses a small salt marsh system. The connection of the marsh with salt water is through an overwash channel. Since ridge growth is the response of intertidal gravel to most wave activity, the channel has only opened 3-4 times/year for the past three years. The brackish conditions presently in the salt marsh has lead to an active Phragmites sp. population overlying Spartina peat.

FIGURE 4

CROSS SECTION
NORTH SPIT, THOMPSON ISLAND



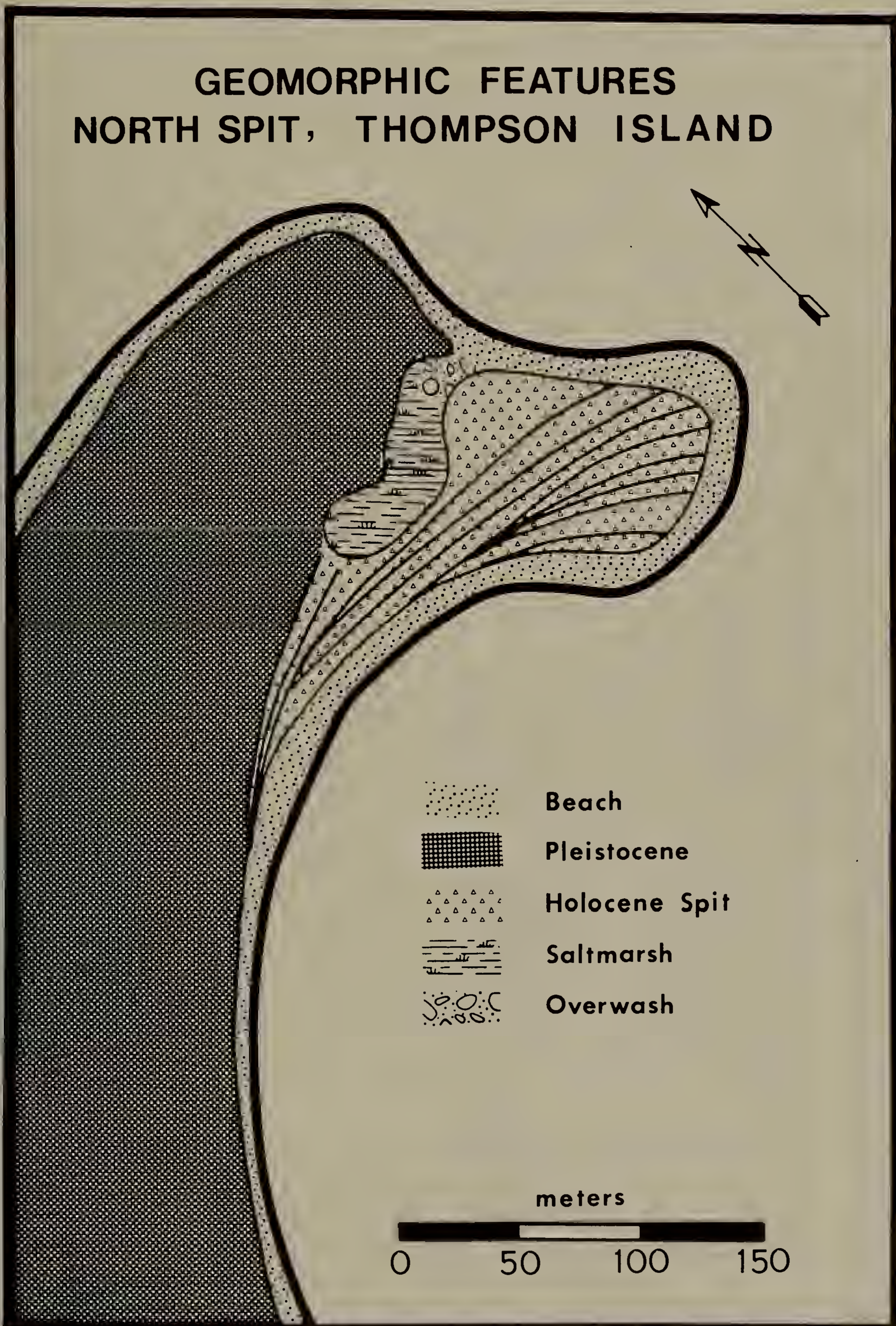

GEOMORPHIC FEATURES
NORTH SPIT, THOMPSON ISLAND
N
Beach
Pleistocene
Holocene Spit
Saltmarsh
Overwash
meters
0
50
100
150

FIGURE 5

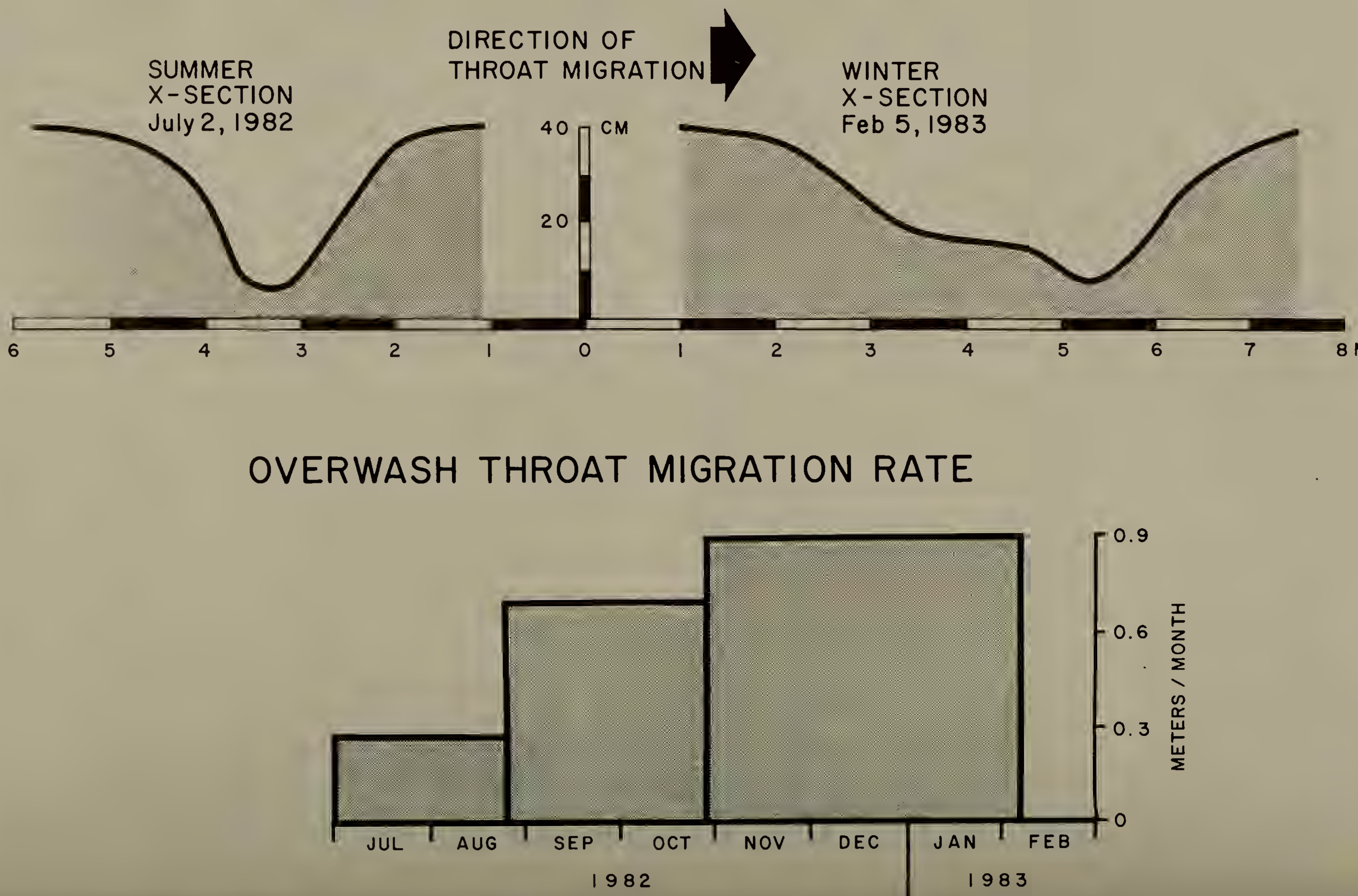
Soundings in the marsh show a maximum depth of about 3 meters (Figure 4). A steep slope exists at the boundary between the marsh and the first ridge of the enclosing spit. This steep slope is comparable to the present form of the downdrift end of the South Longshore Spit. The South Longshore Spit is probably an analog of the initial form of this spit.

Throughout the saltmarsh are several random cobble deposits resulting from winter sediment transport by ice. Freezeup around the island is not uncommon due to the low wave energy. The high tide range causes continual floating and grounding of shore ice while freezing is taking place. This is effective for entrainment of beach materials into the ice. Since freezeup is not usually complete throughout the harbor, waves continually redistribute the shore ice. This has resulted in ice accumulations up to 1.7 meters in height around the spit. However, gravel ice-push ridges have not been observed. This may be due to the continual wave action and large tide range. The ice foot is always fractured, due to tidal flexing, and can more readily be pushed landward by storm wave action. Although shore-ice forms are large, beach profiles before and after freezeup in 1981 on the south shore of the spit showed no measureable change in the beach. The major impact of ice on this shoreline is the redistribution of saltmarsh peat. The ice-transported peat blocks have established a fringe marsh along much of the eastern shore of the island.

The south face of the North Cuspate Spit is an unobstructed length of beach with a known net longshore drift direction (north). Sampling of sediments at both the low water line and storm ridge crest have provided an indication of longshore sorting of gravel by low wave energy. Preliminary data indicates an increase in length, a decrease in sphericity, and more negative OP index in the direction of net transport. This suggests that larger, flatter, and more disc-like particles are preferentially moved alongshore by waves. Observations of tracers have indicated that this preferential transport is not due to an increased efficiency once the particle is initiated. They are too large for intermittant suspension in this wave environment and their pivotability is extremely low, so that they cannot efficiently roll in the swash zone. Deployment of tracers for 3-6 tide cycles during non-storm conditions showed that spherical shapes consistantly moved farthest. However, the majority of pebbles (over 90%) that moved at all were discoid. These trends occured when tracers were placed in a low mound. If the tracers were placed one pebble thick over a sandy gravel substrate, virtually no movement was ovserved.

These preliminary studies indicate that in low energy settings, the high porosity of a pebble substrate plays a significant role in initiation. Increased pore pressure near the plunge point of the wave can initiate motion. Spherical shapes had minimal resisting area to substrate pressure surges relative to their mass, and tend to be locked into a fixed position in the rough bed. While initiation is more difficult, they are readily transported by rolling once motion begins. Flat shapes have larger resisting areas to substrate pressure surges relative to their mass, and are not locked into a fixed position in the rough bed.

FIGURE 6 OVERWASH THROAT FORM



They are readily initiated, but movement per initiation is only a few millimeters.

Since the longterm trend shows larger, flatter particles are found farther downdrift, the controlling transport mechanism appears to be the preferential initiation of flatter particles.

South Longshore Spit

The South Longshore Spit differs from the other spits on the island and most spits in the harbor in that it does not represent a regional drift convergence point. This is due to the sheltering from the east by South Cuspate Spit. The longshore sediment source has been cut off for the past few decades by a tidal inlet updrift of the spit (Figures 2 and 3).

The spit is composed of two geomorphically distinct regions. The updrift end of the spit has existed prior to 1770. This portion of the spit is fully emergent. The supratidal areas are composed of flat pebbles forming a continuous overtopping ridge. This updrift region borders on a narrow saltmarsh, and is close to welding onto the adjacent shoreline. The landward migration rates between 1982 and 1984 have averaged 4.6 meters/year. This rate is probably influenced by the lack of a longshore sediment source.

A gravel overwash throat has existed on this portion of the spit since at least 1978, and may have opened as a result of the February 9, 1978 nor'easter (Blizzard of '78). The processes associated with this overwash system differ greatly from overwash on sandy barriers. The overwash channel is a semi-permanent feature, and migrates in the downdrift direction similar to a tidal inlet. Migration rates increase in winter months, and have averaged 7.5 meters/year between 1981 and 1984. The cross-sectional area of the throat also increases during winter months, presumably due to larger volumes of overwash flow (Figure 6). The migration of the throat has lead to the deposition of a 30 meter long fan/platform overlying the adjacent saltmarsh.

Overwash is not primarily a storm event. The maximum throat elevation is typically near mean high water. Most tides above that level, or neap tides during higher wave energy events, do overwash. Accumulative gravel ridges regularly migrate up the beachface and block the throat, but are readily breached by seepage (q.v) and runup.

Sediment transport during overwash is driven by wave bores entering the throat and downslope flow of water. Wave bores have not been observed to be effective at initiation without accompanying flow. Since there is a lag of about ½ hour in the filling of the marsh (by way of the lagoon) during rising tides, water levels are higher on the seaward side. The first stage of overwash is seepage through the barrier. The elevation of the backbarrier is about 10-15 cm higher updrift of the throat as compared to downdrift areas due to prior deposition of the fan/platform. Seepage was consistantly first observed adjacent to, and downdrift of the throat position. While the volume of water that entered the marsh as seepage was not significant in filling the marsh basin, it attained sufficient velocities to winnow and transport sand from the gravel ridge and deposit seepage lobes landward of the ridge slipface.

The flow through the overwash throat took place from ½ hour before high tide, and ended at high tide. By high tide, the marsh basin had filled so water no longer flowed through the throat. Overwash flow and accompanying wave bores transported all sizes of material through the throat. Transport across the fan/platform was dominated by down-gradient flow. While coarse material tended to move landward, sand-sized material accumulated as levees on the channel margins. The levees were best developed when wave action was low (which was most typical). As water levels increased on the fan/platform, the levees consistantly were first overtopped and breached at the same position: the downdrift side of the throat, which is adjacent to the saltmarsh surface. This area had the steepest gradient as the existing fan/platform raised basin elevations both landward and updrift of the throat position. Once a crevasse formed in the levee, some flow was diverted from the existing landward-oriented channel. A microdelta composed entirely of sand encroached laterally on the marsh, while all transport of coarser material was landward. The wave bore moving landward through the throat apparently played a significant role in the initiation and transportation of coarser material. The area of sand deposition through the levee crevasse is the same area that sand seepage lobes were deposited earlier in the event. In future overwashes, the throat will migrate downdrift and the cobble layer will be deposited over the sand.

The characteristic depositional unit resulting from this gravel overwash is a reverse graded, gravel-over-sand sequence. The gravel layer may show faint coarse/fine layering. This probably represents wave setup, which causes high velocity surges through the throat with 2-5 minute periods. The sandy levees were not found preserved in the sequence.

The downdrift end of the South Longshore Spit may have accreted since 1893, as it is not shown on a survey of that date. It is covered during normal high tides (therefore, technically it's a bar), although its crest elevation is typically near mean high water. This portion of the spit has retreated at an average rate of 0.48 meters/year between 1982 and 1984. Since this gravel spit/bar does not support dune grasses, nor is there any aeolian transport, a mechanism for emergence of this feature above mean high water is not obvious. Most of the length of this downdrift segment of the spit does not border on saltmarsh, as does the updrift segment. The downdrift segment encloses a small lagoon comprised of subtidal muds.

One position on the downdrift end of the spit is regularly emergent above mean high water (Figure 7). The emergent gravel "hump" is adjacent to an isolated marsh clump in the lagoon. The emergent hump was up to 15 cm higher in the spring and summer when Spartina grasses behind it were tall, and lower or non-existant in the winter when grasses were not present.

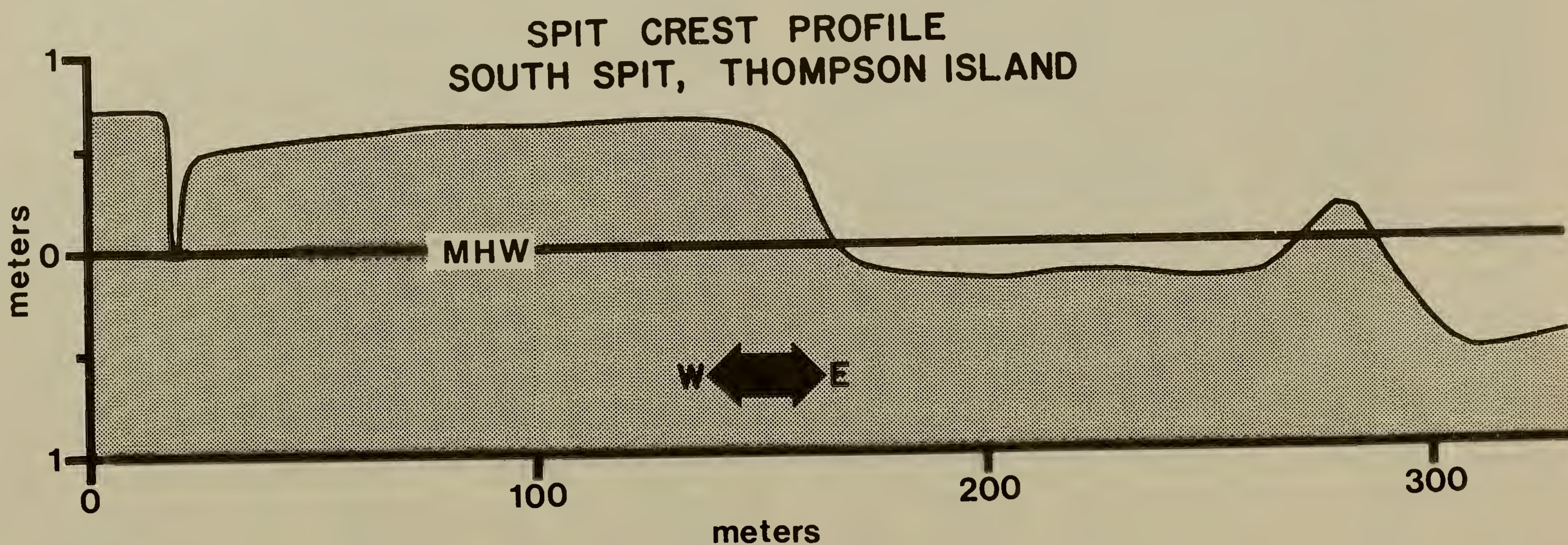


FIGURE 7. Profile along crest of South Longshore Spit

Since the emergent updrift segment of the spit ends abruptly where the adjacent marsh ends, and the hump location corresponds to the presence of lagoonal saltmarsh, the saltmarsh appears to be a controlling factor in gravel spit emergence.

The landward side of the downdrift spit/bar was composed of a continuous 2½-3 meter high slipface. The dip of this slipface was most typically 25-32^{o}. This large, steep face played a role in a sediment transport process resulting from ripples (ht = 2-8 cm) formed in the lagoon. West to southwest winds in the lagoon generated ripples that broke on the spit/bar slipface. The result was downslope transport of sediments in the swash zone of the breaking ripples. During rising tides, this samll-scale erosional scarp migrated up the slipface as water rose. During falling tides, erosion and downslope transport due to ripple swash created a small, lower slope (10-20^{o}) platform below the ripple plunge point. As the tide fell, the ripple swash zone moved down to this lower slope platform and little transport took place. A further fall in water level placed the ripple swash zone on a higher slope again, and the process was repeated. The result at low tide was a series of micro-scarps and micro-ridges extending down the slipface of the spit. Theses "water-level lines" typically had a vertical spacing of about 10 cm.

The formation of water-level lines played a role in the downslope transport of sediment on the landward, steep spit/bar slipface regardless of the approach angle of the ripples. The processes associated with water-level lines also apparently play a significant role in the longshore transport of sediment on the landward side of the spit/bar. Due to the short period ($\approx$1 sec), short wavelength ($\approx$10 cm) of the incoming ripples and steep nearshore gradient ($\approx 30^{o}$), the ripples do not refract as they approached shore. As the prevalent southwest wind direction aligns with the long axis of the spit/bar, ripples often approached nearly perpendicular to the shore (Figure 8). During high wind events, a wave bore was established along the shore (similar to face-travel of waves on a vertical seawall). Longshore currents have been recorded up to 40 cm/sec. Extremely high rates of drift occur in the small area of the ripple swash zone. This transport does play a measureable role in the form of the spit, as the downdrift tip periodically recurves into the lagoon due to normal wave refraction. Following periods of intense southwest winds, the recurve is eroded and planed-off from the landward side of the spit. If the longterm role of water-level line processes is comparable to short term observations, then the position and form of the South Longshore Spit results from a drift convergence

Acknowledgments

The support and cooperation of Thompson Island Education Foundation is gratefully acknowledged. The assistance of Kenneth Leach, Marcia Berman, and Ernest Waterman is appreciated.

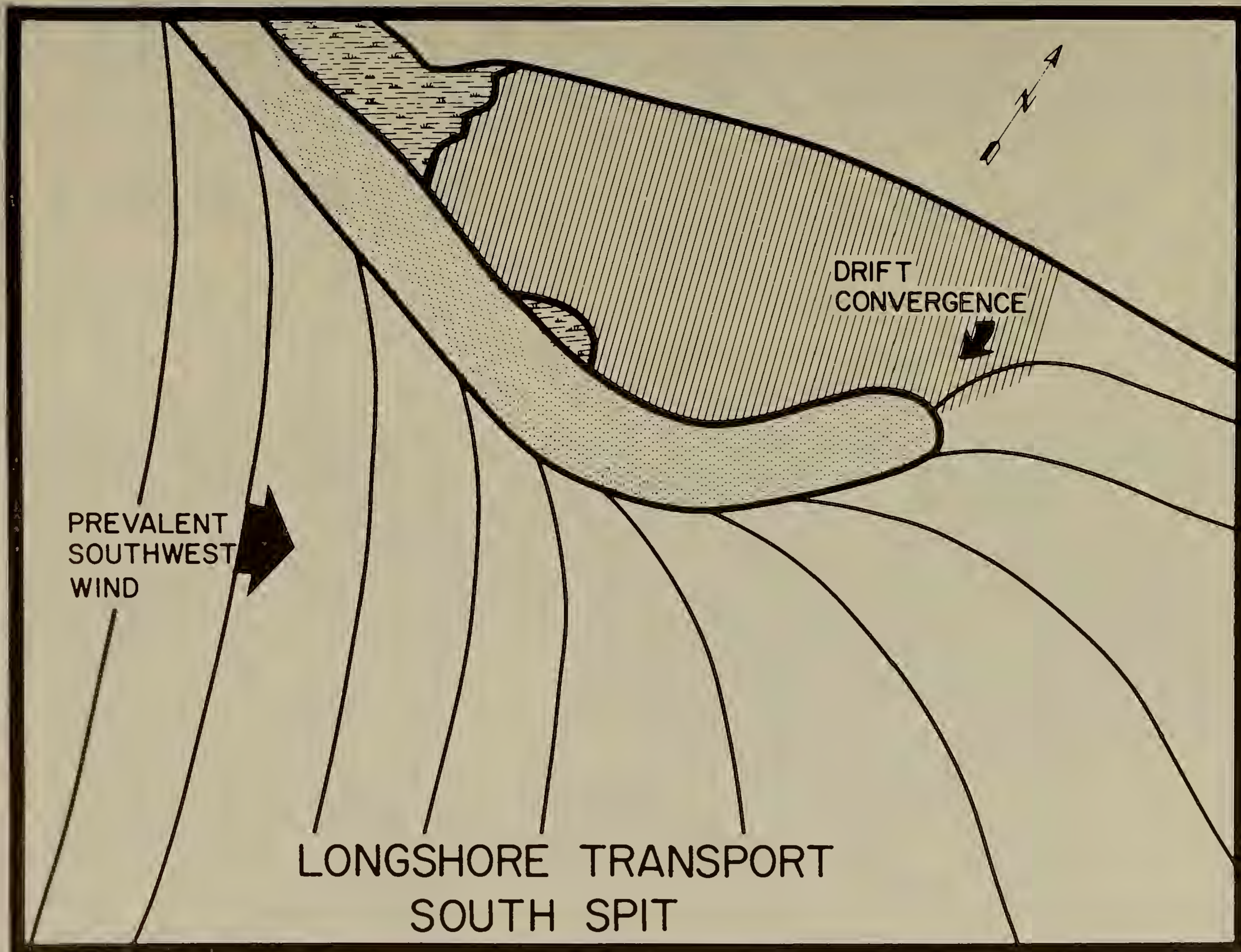


FIGURE 8. Longshore drift by ripples inside lagoon converges with regional waves at spit tip

HAZARDOUS WASTE PROBLEM SITES

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Introduction

Love Canal. Virtually overnight, engineer William Love's ill-fated attempt to construct a navigable power canal between branches of the Niagara River became synonomous with hazardous waste. Within a matter of months, the discovery of other major sites in Grey, Maine; Lowell, Massachusetts (Silresim); and West Point, Kentucky ("Valley of the Drums"), catapulted the question of how industries dispose of their wastes from the purview of a few regulatory agencies into the public eye. Sunday supplements, made-for-TV movies, and even comic strips all reflected the country's heightened concern with the real, potential, or imagined threats to public health and safety which chemical wastes - and, by extension, the companies which produced these wastes - represented.

Out of this explosion of hazardous waste issues into the public consciousness was born the Comprehensive Environmental Response, Compensation, and Liability Act (CERCLA). Among its other provisions, CERCLA provided for the creation of a fund to be used by the federal government to clean up "uncontrolled" hazardous waste sites, sites where the parties responsible for the problem were unwilling or unable to pay for the clean-up. This fund, the "Superfund," is allocated to various sites across the country in accordance with a complicated ranking and selection system.

With its long heritage of such industries as tanning and metal plating, and its high concentration of high-tech firms which generate large volumes of a variety of spent solvents and similar wastes, it is no surprise that New England has its share of superfund hazardous waste sites. In southern New England, their occurrence is directly related to the proximity of the manufacturing facilities. In northern New England, the smaller number of industries is compensated for by the greater amount of open, isolated land in which to illicitly dispose of hazardous wastes.

This field trip will visit two or three hazardous waste sites in northern Massachusetts and/or southern New Hampshire. In selecting possible sites for this field trip, the trip leaders have had to consider such disparate items as location, significance of geology to overall site conditions, accessibility, the need for personal safety precautions (i.e., disposable coveralls, respirators), and the status of pending litigation. Due to the general sensitivity of industries to being identified as hazardous waste sites to busloads of strangers, only uncontrolled sites were considered. Unfortunately, it is not possible to predict in July which sites will be available for visiting in the fall. This decision will be made in late summer, and field trip logs will be provided to participants on the day of the trip.

The following sections of this paper will provide an overview of the relevant aspects of New England hydrogeology, and will describe conditions at a few of the most likely candidate sites for the actual field trip.

Hydrogeologic Framework

Broadly speaking, the geology of New England consists of glacial deposits overlying bedrock. Where a significant water-saturated thickness of permeable glacial deposits (sand and gravel) is involved, the deposit is considered an aquifer, a source of water supply. Although many private wells and occasional municipal and industrial wells draw water from fractures in bedrock, most large-yield production wells in Massachusetts are located in these sand and gravel aquifers.

Hazardous materials released onto or into the ground will tend to move downward under the force of gravity through the unsaturated zone until the water table (the top of the saturated zone) is reached. In most cases, the introduced materials will move with the groundwater as it slowly moves toward a downgradient discharge point such as a river or lake. The rate of movement will generally be very slow, on the order of a foot per day, although the rate may be substantially higher or lower depending on such factors as the nature of the soils involved and the nature of the introduced material.

The situation is complicated when the amount of material introduced exceeds the amount which can be dissolved in the groundwater. In certain situations, large releases of undiluted material may result in lenses of the material "floating" on top of the water table or sinking through the uppermost permeable unit to move as a body along the bottom of the unit, depending on whether the material in question is less or more dense than water. This phenomenon commonly occurs in situations involving gasoline or oil spills, where the lens on top of the water may be thick enough to recover as pure product. The movement of introduced material as discrete bodies is much less amenable to analysis than the movement of this material dissolved in groundwater.

The situation is further complicated by the fact that materials introduced into the ground may interact with the soils involved in any number of ways. The soils may result in significant attenuation of the material or may merely retard its movement. A site involving a mixture of materials may thus have associated with it a number of contaminant "plumes," each with a slightly (or significantly) different size and shape. Some contaminants (e.g., metals, PCBs) tend to be retained in the soil, and may not even reach the water table.

For analytical purposes, the hydrogeologic environment is often considered to be a readily definable, fully predictable entity. This, of course, is science fiction; while the general rules of hydrogeology are clean, their application to any individual parcel of land is somewhat ambiguous. Coupled with this uncertainty is the often haphazard nature of releases of materials into the environment, which are seldon uniform in rate, consistent in location, or constant in composition. Thus, the prediction of the rate and direction of movement of contaminant plumes, which is critical to the evaluation of impacts and the necessity for remedial actions, is generally the most complex and least precise of the tasks involved on a hazardous waste site.

Regulatory Framework

As noted above, the superfund was created by CERCLA. The procedures governing the actions related to evaluating and remediating a hazardous waste site are outlined in the National Contingency Plan (NCP), published in the Code of Federal Regulations as 40 CFR 300, and described in more detail in a host of guidance documents and handbooks published by the Environmental Protection Agency (EPA) and others.

The NCP covers all phases of the hazardous waste site evaluation procedure, from initial identification through final remedial action. Of most interest to the various engineering and geotechnical consultants involved in hazardous waste work are the "RI/FS" activities described below.

Remedial Investigation

The Remedial Investigation (RI) is the exploratory phase of the site study. The purpose of the RI is to evaluate the nature and extent of contamination; identify the on-site and off-site receptors, and assess the risks which the site poses to these receptors; assess the various remedial options possible; and select the most appropriate options for further study and refinement in the subsequent Feasibility Study. Additionally, the RI is meant to identify any Immediate Remedial Measures (IRMs) which should, for reasons of public health and safety, be taken even before the completion of the RI, such as the removal of leaking drums, the emptying of lagoons, etc.

Feasibility Study

The objectives of the Feasibility Study (FS) are to evaluate in more detail the recommended remedial alternatives from the RI, and to decide upon the most feasible and cost-effective option. The FS includes such tasks as bench-scale testing of various treatment options, and preliminary design of the final remedial option. The Feasibility Study is the last step prior to the actual implementation of whatever remedial action is chosen.

Hydrogeology of Selected Hazardous Waste Sites

In the preliminary field trip announcement, it was indicated that this trip would probably include the Woburn (Industriplex) and Tyngsboro (Charles George Landfill) hazardous waste sites. As of this writing (July), it appears more likely that the sites to be visited will be among those described in this section. However, as noted above, the final selection will not be made until late summer.

Gilson Road Site, Nashua, New Hampshire

As the first Cooperative Agreement signed under CERCLA, the Gilson Road case represents a precedent-setting example of response to an uncontrolled hazardous waste site. The site, a former gravel pit, had been used for the illegal disposal of a variety of liquid and solid wastes prior to its discovery and subsequent closing by the state in 1979. Estimates indicated that several million gallons of liquid wastes had been disposed of at the site, primarily via bulk dumping into a subsurface disposal trench. The principal contaminants of groundwater at the site included volatile organic constituents, alcohols, and metals.

The surficial geology of the site was mapped at the quadrangle scale by Koteff (1973), while New Hampshire state bedrock and surficial maps were prepared by Billings (1956) and Goldthwait (1950), respectively. Both the published information and the data developed as part of the site investigation indicate that the site is underlain by deposits of stratified sand and gravel 30 to 90 feet thick. The bedrock (biotite schist of the Merrimack Group) surface is relatively level across most of the site, but drops rapidly from 40 to 50 feet MSL to -10 feet MSL toward the western edge of the site. It is believed that this rapid drop-off is due to the presence of a preglacial valley in that vicinity.

The hydraulic properties of the aquifer were evaluated based on data from pumping tests conducted as part of the site study. From this data were derived the estimates of groundwater flow velocities and volumes which were subsequently used to evaluate off-site impacts and to design the remedial measures.

The hydrological and chemical data developed during the field studies of the Gilson Road site were integrated into an engineering analysis of the site, which in turn was used to identify and evaluate alternative remedial measures. The final remedial scheme involved both hydrologic isolation via a slurry wall and impermeable cap, and groundwater interception and treatment of the most contaminated portion of the aquifer.

During the year it took to finalize the conceptual design and funding of the remedial measures, the contamination plumes moved a sufficient distance to require that a larger area of aquifer be isolated than originally planned. To minimize further movement of the plume prior to the actual completion of the isolation system, an innovative groundwater recirculation system was installed to temporarily limit contaminant migration. This system was designed using a computer flow model based on the aquifer hydraulic data described above.

At present, the isolation system is complete, and the treatment plant is under construction.

O&G/GLCC Site, Kingston, New Hampshire

The Ottati and Goss/Great Lakes Container Corporation (O&G/GLCC) site consists of 35 acres located immediately west of Route 125 in Kingston, New Hamphsire. The site occupies an east-west trending topographic valley which drains eastward to a marsh area beside Country Pond, and is bounded on the north and south by brooks which also flow to this marsh. Portions of the site have been used for drum reconditioning and, allegedly, for the disposal of hazardous materials since at least the late 1950's. Groundwater contaminants consist primarily of volatile organic compounds (VOC) and acid extractible/base neutral (ABN) compounds.

The geology of the O&G/GLCC site and surrounding area has been described in detail in a number of published studies. Sundeen (1971) mapped the bedrock of the Haverhill quadrangle, while the surficial geology of the area was mapped by Bradley (1964) as part of a USGS Water Supply Paper for southeastern New Hampshire. The study area is also included in the state surficial (Goldthwait et al., 1951) and bedrock (Billings, 1956) maps. The bedrock is mapped as biotite schist of the Merrimack Group; the bedrock surface has a relief of about 50 feet across the site, with the lower elevations related to a converging pair of preglacial valleys. Surficial deposits consist (Bradley, 1964) of outwash and ice contact sand and gravel deposits.

The issues of concern at this site include not only the flow of contaminants through the sand and gravel aquifer, but also the flow of contaminants in the underlying bedrock, and the potential impact of the O&G/GLCC site on nearby Country Pond. The pond itself is used for swimming, boating, and fishing, and private wells around the pond probably capture some water from it.

Since the RI report on this site is still in draft form as of this writing, the conclusions regarding the extent and impacts of contamination cannot yet be discussed. However, this information is expected to be available by the time of the field trip.

Municipal Landfill, Dover, New Hampshire

The Dover Municipal Landfill includes 47 acres of city property and 8 acres of private property. Landfilling of the site began in 1954 and continued through 1979; final closure was completed in May 1980. During its active period, the landfill received the solid municipal wastes from the City of Dover, treated tannery wastes from the City's sewage treatment plant, and a variety of industrial wastes, organic solvents, and waste oils. In addition, sealed drums were probably landfilled prior to 1975. The City of Dover operated the landfill and maintains current ownership.

The major environmental concern posed by the Dover Landfill site is its proximity to, and potential contamination of, two primary municipal water supply sources: the Bellamy Reservoir to the south and the Calderwood Well to the north. Previous hydrogeologic studies confirmed that organic solvents were moving from the landfill toward the well, but soil conditions are such that it was difficult to verify the movement of groundwater toward the reservoir. In 1981, shallow domestic supply wells adjacent to the landfill were found to be contaminated by organic solvents, and city water mains were extended to provide water to the area.

The Dover Landfill is located on a relatively flat terrace underlain by outwash deposits of stratified sands with localized gravel and silt inclusions. The outwash deposits are underlain by marine deposits consisting of an upper unit of stratified sands, silts, and silty clays, and a lower unit of silty clay. The outwash deposits and upper marine deposits apparently form an upper aquifer perched above the relatively impermeable lower clay unit. The aquifer is 40 to 50 feet thick beneath the landfill and thins to the west as the surface of the lower marine clays rises.

The marine deposits are underlain by ice contact sediments of stratified sand and gravel deposits of variable thickness. Previous studies suggest that this unit outcrops north of the site near the Calderwood Well before being confined beneath the marine clays in the area beneath the landfill. This sedimentary unit is a major regional aquifer in this area; the Calderwood Well is screened in it. The confining nature of the clay unit is evidenced by water level readings at multilevel wells screened above and below the unit. The ice contact deposits are underlain by glacial till and by biotite schist believed to belong to the Berwick Formation of the Merrimack Group.

While it appears that pumpage of the Calderwood Well influences the groundwater flow regime in the lower aquifer beneath the landfill, the extent to which flow paths in the upper aquifer may bring contaminants from the landfill to the well has yet to be determined. Moreover, the extent to which upper aquifer groundwater moves toward the reservoir, and the operating conditions under which this might occur, are unknown.

Work on the Dover Landfill RI/FS is just beginning as of this writing, but should be well underway by the fall.

Silresim Site, Lowell, Massachusetts

The Silresim Chemical Corporation established a chemical waste reclamation facility in Lowell in 1971. When, seven years later, the company declared bankruptcy and abandoned the site, it left behind approximately one million gallons of hazardous materials in drums and tanks. This material was removed from the site in several phases during the period 1978-1981 at a cost approaching three million dollars.

The site is underlain by a series of silts and sands deposited into ancient post-glacial Lake Shawsheen. The deposits are nearly 100 feet thick at the site; this, in conjunction with subsurface data from surrounding areas, suggests that the valley of the preglacial Merrimack River lies beneath the the site. While regional groundwater flow is northward toward the Merrimack River, there appear to be components of flow toward the Concord River to the east, and toward a small pond to the south. While a shallow silty stratum may serve to limit downward movement of groundwater beneath the site, the precise effect of this stratum on groundwater flow, and the extent to which it serves as an impermeable zone beneath the uppermost aquifer, have yet to be determined.

Soils and groundwater at the Silresim site contain significant levels of organic compounds. The contamination plume associated with Silresim can be traced to a depth of at least 50 feet, and downgradient for a distance of at least 800 feet. To date, impacts of the Silresim site have not been observed in off-site surface water or soil samples.

The Silresim Chemical Corporation buildings have been razed, and a clay cap has been laid over the site. The scope of the RI/FS for the Silresim site is currently under discussion, and it is possible that the RI/FS will be underway at the time of the NEIGC trip.

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A4. SEDIMENTOLOGY AND MULTIPLE DEFORMATION OF THE KITTERY FORMATION, SOUTHWESTERN MAINE AND SOUTHEASTERN NEW HAMPSHIRE

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The purpose of this field trip is to examine the well-preserved sedimentary and structural features preserved in the Kittery Formation, a thick sequency of turbidite and contourite (?) deposits in southwestern Maine and southeastern New Hampshire. We will examine exposures of the Formation along the Marginal Way, a public shoreline footpath in the town of Ogunquit, Maine (Fig. 1), and exposures at several localities in the Kittery area (Fig. 1, Fig. 6). Emphasis at the Marginal Way locality will be on structural interpretation, and in the Kittery localities, on sedimentologic interpretation as to environments of deposition and direction of sedimentary transport. Hussey is responsible for the geologic mapping and interpretation at the Marginal Way, and Rickerich and Bothner for the Kittery area.

Geologic Setting

The Kittery Formation, the lowest unit of the Merrimack Group (Table I) crops out in a 5-10 km wide belt extending from near the Seabrook area, New Hampshire (Novotny, 1963) northward to the Biddeford Pluton in Kennebunkport, Maine (Hussey, 1962), and in the center of the Exeter anticline between Dover and Exeter, New Hampshire (Billings, 1956) (Fig. 1). The Kittery Formation is conformably overlain by the Eliot Formation, and is in ductile fault contact with the Rye Formation (Hussey, 1980 ; Carrigan, 1984).

The Kittery Formation consists of interbedded purplish gray calcareous and somewhat feldspathic metaquartzite, and very fine-grained dark gray phyllite or biotite schist (metapelite). Metaquartzite bed thicknesses are extremely variable normal to bedding and very uniform parallel to bedding. Metaquartzite beds range in thickness from 1 cm to about 3 meters, averaging about 25 cm; metapelite from 1 cm to a few meters, but are generally less than 10 cm thick. (Rickerich, 1983) Thicker metaquartzite beds commonly have basal zones of coarse sand to granule-sized grains including quartz (both milky and bluish) and minor feldspar.

In the exposures around Kittery studied by Rickerich (1983), a variety of sedimentary structures are present. The bases of many metaquartzites in this area are characterized by scour features and primary deformation features. Undulatory erosional surfaces are present at the bases of some metaquartzites, and often distinct metaquartzite beds are in erosional contact. Channeling less than 50 cm in width is occasionally present at or near the base

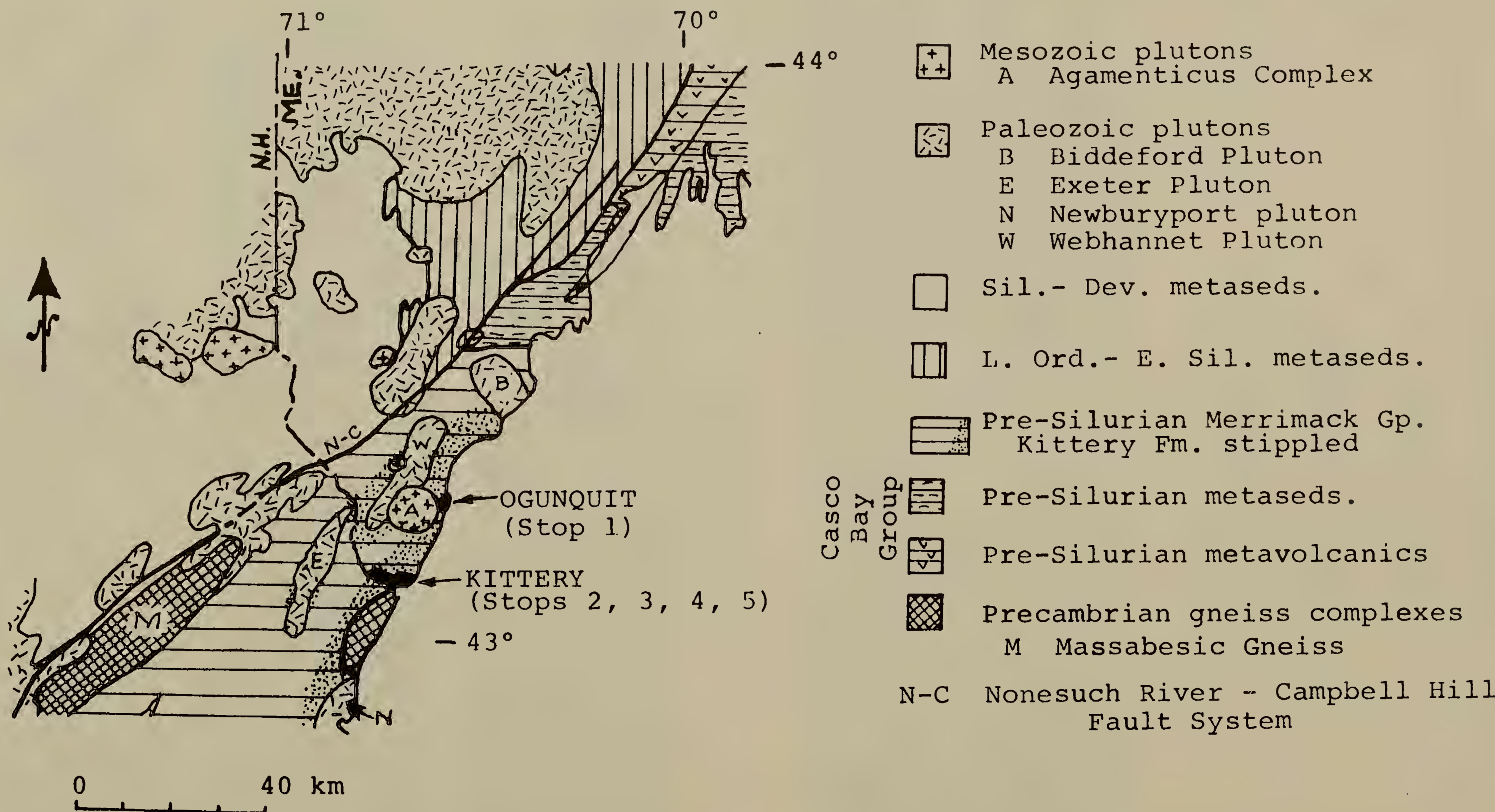

Fig. 1. Location map and generalized geologic map of southwestern Maine and southeastern New Hampshire (after Hussey and Pankiwskyj, 1975; and Lyons et al, 1982).

of some beds. Flute marks are rare, but both load and flow flames are frequently present in many metaquartzites.

Basal layers of some metaquartzites often contain thick, well-defined, laterally continuous parallel laminae; large-scale cross laminae; and dewatering or early cleavage structures. Thick parallel laminae tend to be spaced about 1 cm apart and usually occur in thick-bedded coarse-grained metaquartzites. Most large-scale cross-laminae are also restricted to coarse-grained thick-bedded metaquartzites and are frequently dunal in nature, although some large-scale antidunal cross-laminae are present. The dewatering or early cleavage structures are weak laminae that dip 10° to 26° in an upcurrent direction and are most often restricted to narrow laterally-continuous zones in mataquartzite beds. Basal layers of almost all metaquartzites are normally graded.

The basal layers of fine-grained metaquartzites, and the upper layers of coarse-grained metaquartzites frequently contain a range of delicate sedimentary structures. Thin (~1mm) usually close-spaced, laterally-continuous bed-parallel laminae generally underlie small-scale crossbedding. Small-scale crossbedding is most often plano-lenticular. Upward-steepening climbing ripple structures are occasionally present.

Each of these structures can be interpreted as having been deposited over a specific range of current flow conditions, i.e., velocity, duration, load, and competency. Specific suites of sedimentary structures, bedding characteristics, and grain size characteristics are helpful in identifying the nature of the current that deposited a particular bed. Thick-bedded medium to coarse-grained poorly-sorted normally graded metaquartzite beds with high metaquartzite/metapelite ratios, sharp upper and lower contacts, flame structures, thick parallel laminae, large-scale cross laminae and dewatering structures are indicative of high velocity deposition. Thinner-bedded, medium to fine-grained graded metaquartzites with low metaquartzite/metapelite ratios, often containing well-developed extremely thin parallel laminations and small-scale crossbedding are interpreted as having been deposited by low-velocity currents.

TABLE I

Stratigraphic Column,
Southwestern Maine and Southeastern New Hampshire

Siluro-Devonian	Shapleigh Group		Interbedded metapelite and metasandstone; rusty metapelite
E. Ordovician (?) to Precambrian (?)	Merrimack Group	Gonic Fm	Metapelite
		Berwick Fm	Medium-bedded to massive biotite and calc-silicate granofels and gneiss
		Calef M	Black phyllite
		Eliot Fm	Thin-bedded calcareous metapelite and metasiltstone
		Kittery Fm	Thin to thickbedded calcareous metasandstone, metasiltstone, and metapelite
Precambrian (?)		Rye Fm	Blastomylonitic calc-silicate gneiss, biotite gneiss, and metapelite gneiss

The apparent decrease in depositional energy upwards in each bed, the range of sedimentary features observed, the lateral continuity of parallel laminae, and the laterally constant thickness of each bed are all excellent evidence that the vast majority of beds in the study area were deposited by turbidity currents. In fact, most of the suites of structures observed in metaquartzite beds are analogous to specific turbidite suites discussed by Walker (1967, 1978), Ricci-Lucchi (1975), Mutti (1977), and Keith and Friedman (1977).

High velocity turbidity current deposits are often in close stratigraphic juxtaposition with low velocity turbidity current deposits. There are several possible sedimentologic explanations for this occurrance and the simplest explanations fit nicely within the context of a submarine fan model (Rickerich, 1983). Specifically, turbidite sequences in the Kittery area which contain evidence of rapid deposition have characteristics typical of the upper or central area of a submarine fan lobe (Walker, 1978). Turbidite sequences which contain evidence of low energy deposition are typical of the fringe area of a submarine fan lobe. Migration, switching or progradation of submarine fan lobes can account for stratigraphic variability of bedding styles. Temporal fluctuation in turbidity current magnitude can also explain stratigraphic variability.

The load flames, rip-up clasts, cross-bedded units, large and small-scale parallel laminae, and normal grading in the thick and thin-bedded metaquartzites near Fort McClary (Stop 2) are typical of the primary sedimentary features observed throughout the map area. A massive, thick-bedded metaquartzite near the Henry H. Cook Memorial School, Kittery (Stop 3) contains large-scale cross laminae and scour features that are less common in the map area, but are typical of high current velocity features observed in some beds in the Kittery Formation.

The range of bedding styles and sedimentary features at Fort McClary and the Cook Memorial School outcrops supports the likelihood that the Kittery Formation is a metaturbidite and demonstrates the stratigraphic diversity of bedding styles which led to the classification of the paleodepositional environment of the Kittery Formation as the lobe facies of a submarine fan (Rickerich, 1983).

Not all of the metaquartzites in the Kittery Formation are believed to be turbidites. The Squash Island outcrop (Stop 4) contains a unit which is tentatively identified as a contourite. This unit is thick-bedded, fine to medium grained, well-sorted, crossbedded and ungraded. Several possible contourites were observed in the Kittery area.

The paleocurrent azimuth mean for contourites in the study area is 308° with a standard deviation of 30°. In contrast, the paleocurrent azimuth mean for all turbidites in the area is 264° with a standard deviation of 60°, thus indicating an easterly source area for Kittery turbidites. It is not inconsistent for contourites and turbidites to be interbedded (Lonsdale and Hollister, 1979; Stow and Lovell, 1979).

Several important trends can be distinguished across the map area from southeast to northwest. The Pleasant Street outcrop (Stop 5) in South Eliot, Maine, is the most northwesterly and uppermost exposed section of the Kittery Formation in the study area. The average bed thickness at this outcrop is less than at the previous stops, the metaquartzite/metapelite ratio is lower, few high-velocity depositional features are present, and the metaquartzites are finer grained. These observations are representative of general trends from southeast to northwest across the map area and are also consistent with westward turbidity current flow.

The finer-grained, thinner-bedded more pelitic Eliot Formation is exposed a few hundred meters to the northwest of Stop 5. The contact between the Kittery and Eliot Formations is believed to be gradational.

The Kittery Formation has been regionally metamorphased from low Greenschist facies in the Wells Beach area to high Greenschist or possibly low Epidote Amphibolite facies in the Gerrish Island area. (See field trip B-4 by Swanson and Carrigan, this guidebook.)

Three major fold phases affect the Kittery Formation. The earliest, F_1, consists of SW-facing recumbent isoclines of possible regional extent. Parasitic mesoscopic scale recumbent F_1 isoclines are best observed along the Marginal Way in Ogunquit (Stop 1). F_2 folds are upright open to steeply overturned tight folds with plunges up to, but generally considerably less than, 30°; plunge reversals are common. Where tight, these folds have a well-developed axial plane cleavage (S_2) along which, in hinge zones, transposition of bedding is marked. F_3 folds are relatively open, overturned generally dextral fold sets with strongly developed strainslip cleavage, S_3, also involving significant bedding transposition in hinge zones. Cleavage to be seen along the Marginal Way (Stop 1) is correlated with S_3.

The Kittery Formation has been intruded by a great variety of magma types over a long period of time from Middle Ordovician (or perhaps earlier) to Cretaceous times. The oldest reported radiometric age for igneous rocks intruding the Kittery Formation is for the Newburyport pluton (Fig. 1) which intrudes the Kittery Formation in the Newburyport-Seabrook area. Zartman and Naylor (1984) report a zircon age of 450±15 Ma for this pluton. The Webhannet Pluton (Fig. 1) with a zircon age of 403±12 Ma and a Rb/Sr age of 390±10 Ma was intruded at the end of deformation of the Acadian Orogeny (Gaudette, et al, 1982). An early Mississippian age (341±12 Ma, Rb/Sr) is reported for the post-tectonic Biddeford Pluton (Fig. 1) (Gaudette, et al, 1982). Post-tectonic alkalic rocks of the Agamenticus Complex with a Rb/Sr age of 228±5 Ma (Foland and Faul, 1977) were intruded in Early Triassic time. These rocks are similar to, but somewhat older than, alkalic rocks of the White Mountain Magma Series in New Hampshire. Post-tectonic funnel intrusions and cone sheets of the Cape Neddick Complex (a small composite pluton in the town of York just southeast of the Agamenticus Complex) give a K/Ar age of 116±2 Ma (Foland and Faul, 1977) and is the youngest pluton, being emplaced during Cretaceous time. Between the time of injection of the Cape Neddick and Agamenticus Complexes two suites of basic dikes, with occasional felsic dikes, were emplaced. These are abundant in coastal zone exposures of southwestern Maine and southeastern New Hampshire, and will be well seen at Stop 1. The earlier suite of dikes may have been emplaced during Triassic to Jurassic time and the later suite in Jurassic to Cretaceous time (Swanson, 1982, p. 137).

The 450 Ma age for the Newburyport pluton places a minimum age of Middle Ordovician on the deposition, deformation, and possibly metamorphism of the Kittery Formation. Lyons, et al (1982) discuss relations of the

Berwick Formation and Massabesic Gneiss that suggest a possible late Precambrian age for the Merrimack Group as a whole, and thus for the Kittery Formation. Correlation of the Merrimack Group with the Vassalboro Formation of latest Ordovician to earliest Silurian age, despite the striking lithic similarity, now appears untenable.

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Road Log

Meeting Time and Place: Assemble in front of the Information Center at the Rest Area on the northbound lane of I-95 in Kittery at 8:00 a.m. sharp. People coming from the north should exit from the Maine Turnpike at the York tollbooth and proceed south on U. S. 1 for 3.2 miles then turn right into road marked "Rest Area." Please be prepared to consolidate into as few vehicles as possible--parking at Stop 1 may be quite limited or costly. We will return to the Rest Area to pick up vehicles before proceeding to Stop 2 (Lunch Stop).

From the Rest Area, exit to U. S. 1 around the north side of the Information Center. Road log begins at the junction with U. S. 1.

Mileage	
0	Junction Rest Area access road with U. S. 1. Turn left on U. S. 1.
1.4	Cross over York River.
2.8	Jcn, U. S. 1A. Stay on U. S. 1.
3.2	Jcn, I-95. Stay on U. S. 1.
6.4	Jcn, U. S. 1A. Stay on U. S. 1.
10.5	Ogunquit Square. Turn sharp right onto Shore Road.
11.3	Leave Shore Road. Proceed straight ahead to Perkins Cove.
11.5	Park in public parking lot. If full, park in pay lot ($2-$3).

Stop 1. Marginal Way. Walk northeast along the Marginal Way, a paved public footpath. We will proceed to the north end of rock exposures near the Ogunquit River (approximately 1/2 mile) and then work our way back to Perkins Cove.

The area seaward of the footpath is a nearly continuous exposure of the Kittery Formation, and abundant basic and felsic dikes that intrude it (Fig. 2). In general, bedding is not as well preserved here as a little further south and in the Kittery area. However, bedding styles, thicknesses, and metaquartzite/metapelite ratios vary greatly, probably reflecting the same variations in local submarine fan sedimentation as described for the Kittery area. Thicker beds commonly have coarse sand bases, and quick grades to thin metapelite tops. Thinner beds commonly have slower grades. Cross lamination is present, but relatively rare.

Bedding along this part of the Kittery exposures varies from vertical to gentle easterly dips, the latter predominating (Fig. 3).

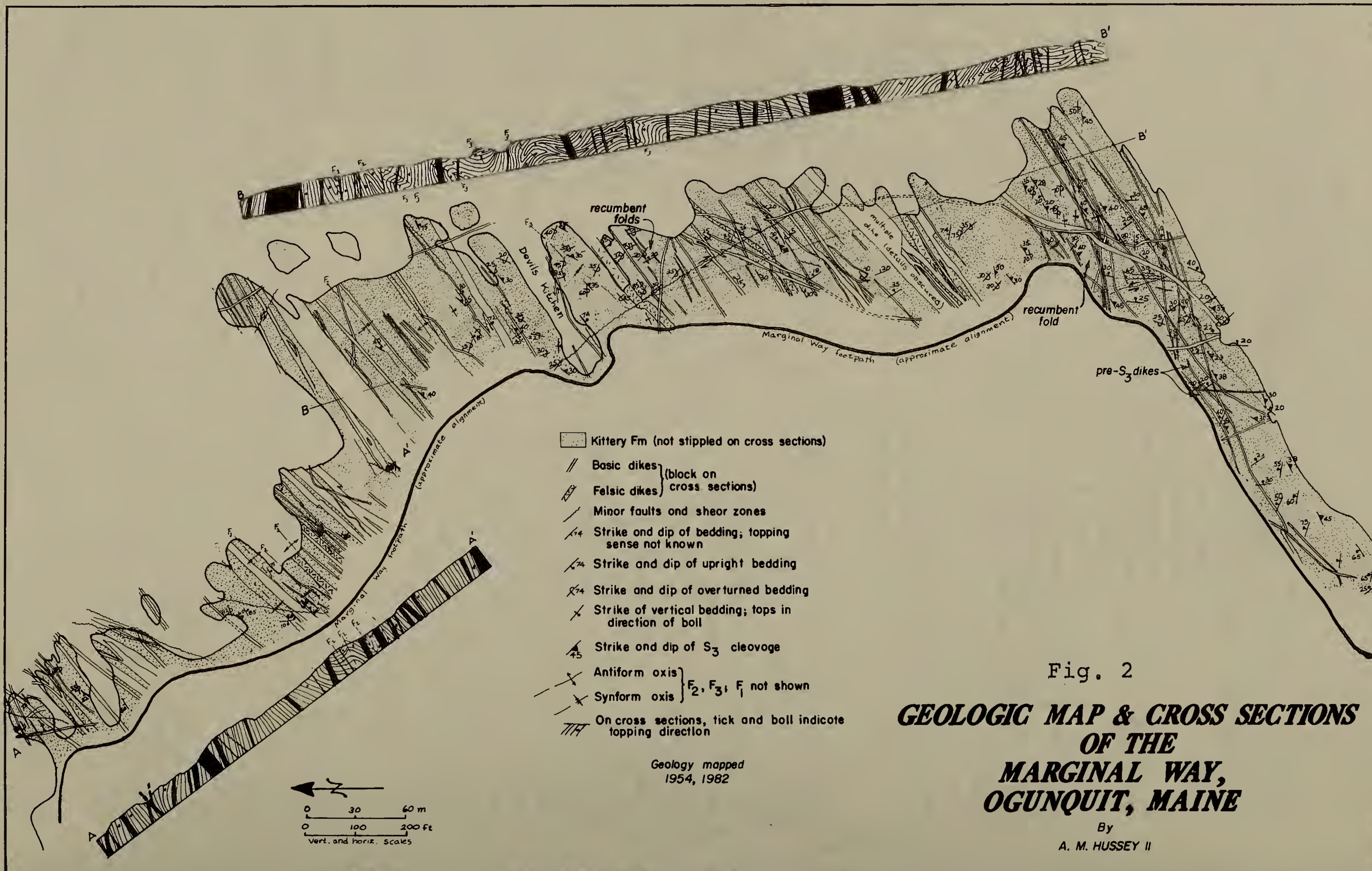


Fig. 2

GEOLOGIC MAP & CROSS SECTIONS OF THE MARGINAL WAY, OGUNQUIT, MAINE

By

A. M. HUSSEY II

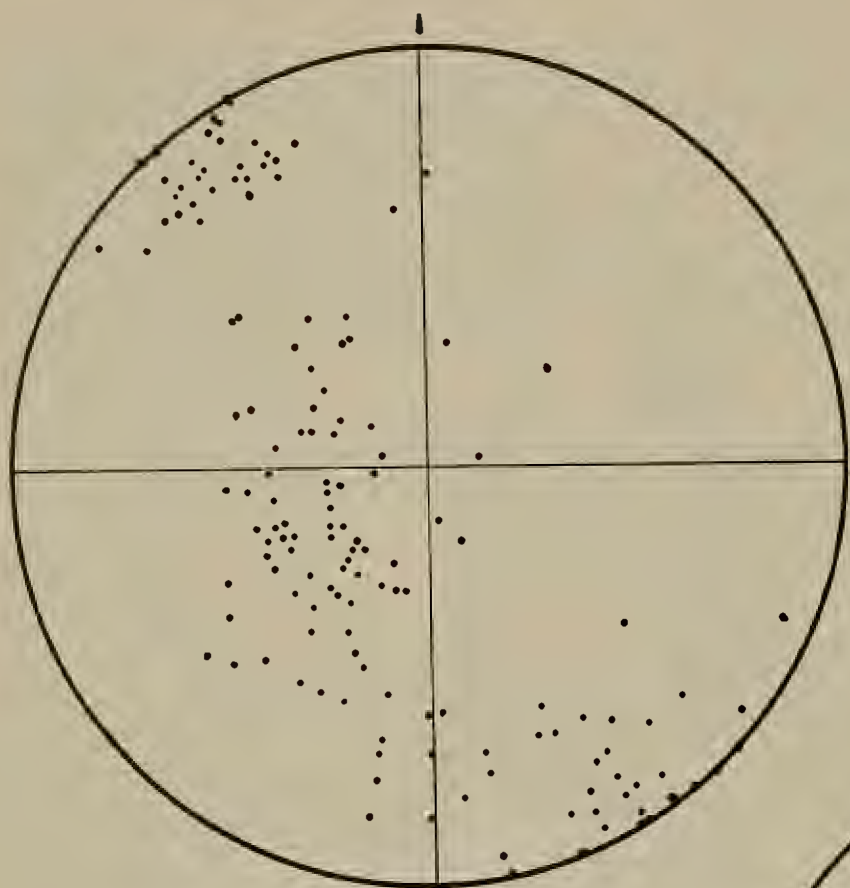


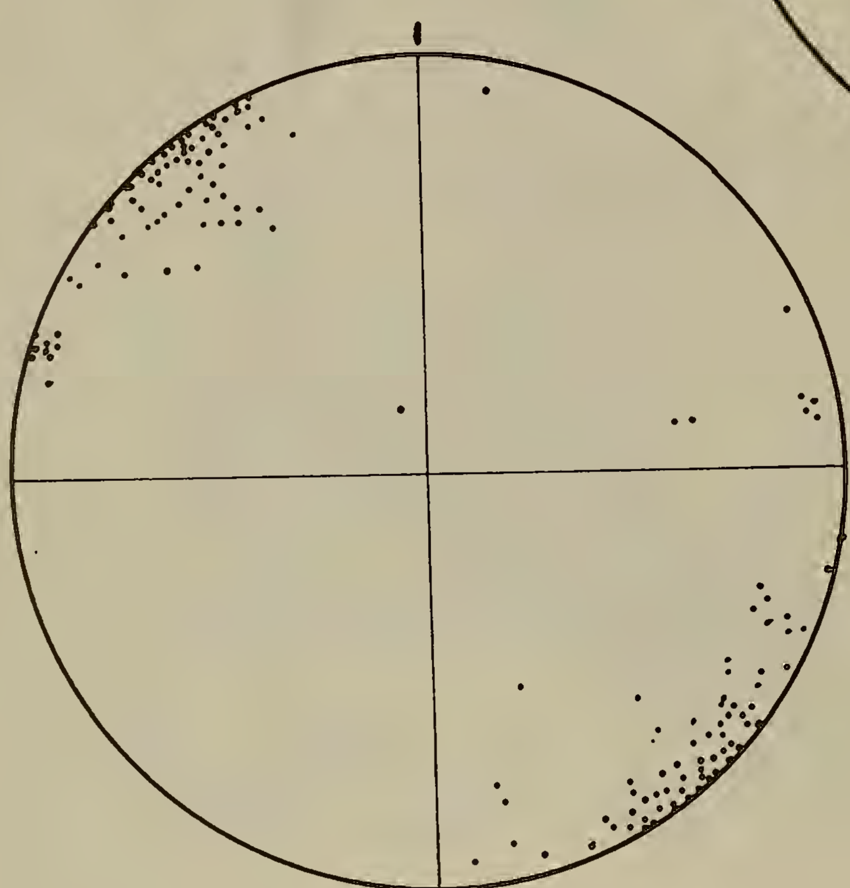
Fig. 3 (left). Stereographic projection (lower hemisphere plot) of poles to bedding, Marginal Way, Ogunquit, Maine.

Fig. 4 (above). Stereographic projection (lower hemisphere) of:

- • poles to S_3 cleavage
- o poles to AP, F_3
- × axes, F_3
- □ poles to AP, F_2
- ■ axes, F_2
- △ poles to AP, F_1
- ▲ axes, F_1

Marginal Way, Ogunquit, Maine

Fig. 5 (below). Stereographic projection (lower hemisphere plot) of poles to dikes, Marginal Way, Ogunquit, Maine.



Low dips reflect the scarcity of tight upright F_2 folds. F_2 folds here are gentle and open except toward the north near the Ogunquit River where bedding dips are much steeper. No cleavage is associated with F_2 here. Mesoscopic F_1 folds are common, particularly south of Devil's Kitchen (Fig. 2). They are generally asymmetrical with long inverted limbs and short upright limbs, implying they are parasitic to larger scale recumbent isoclines. Axial planes and axes vary moderately in their structural orientation (Fig. 4), but, in general, F_1 folds face southwest. Local bed-parallel cleavage may be an axial plane cleavage to these folds. F_3 folds are open asymmetric inclined to overturned dextral fold sets plunging uniformly to the northeast at gentle to moderate angles (Fig. 4). Strain-slip cleavage (S_3) with moderate dip to the southeast parallel to axial planes of F_3 is strongly developed in metapelitic beds (Fig. 4). Local transposition of bedding along S_3 is marked.

Approximately 125 dikes have been mapped along the Marginal Way. These are mostly basic dikes (basalt, diabase, basalt porphyry), but also included are blue-weathering alkaline trachytes (probably not tinguaites--not feldspathoidal) and buff weathering trachyte porphyry. The blue trachytes are everywhere older than basic dikes; none have been observed to cut basic dikes. These blue trachytes are probably related to the nearby alkaline quartz syenite and syenite of the Agamenticus Complex (Fig. 1). Most of the dikes trend between N50E and N60E, and have steep dips (Fig. 5), parallel to the trend of the Bald Head dike swarm described by Swanson (1982). Examination of the geologic map of the Marginal Way (Fig. 2) suggests a later dike set varying in strike between N25E and N-S, and having steep dips. These are roughly parallel with the trend of, and may correlate with, the Gerrish Island swarm described by Swanson (1982). Two dikes have been mapped that are not folded but appear to have an S_3 fabric imposed on them (Fig. 2). In thin section they are very fine grained and irregularly textured, and consequently, mineral identification is very difficult. Are these early igneous dikes or might they possibly be clastic dikes? The fact that they are unfolded favors the former interpretation.

Numerous rusty-weathering shear zones or faults are present (Fig. 2). They are generally steep to vertical; carbonate mineralized (calcite and ankerite); penecontemporaneous with dike emplacement (some basic dikes are cut by the shears, others not); and mostly linear but occasionally sinuous. Some do not appear to offset bedding or earlier dikes, whereas others involve offsets of a few centimeters to a few meters. Offset movement is generally oblique or strike slip, both dextral and sinistral.

Return to vehicles at Perkins Cove not later than 11:45 a.m. Turn around and proceed back toward Ogunquit.

Mileage

11.7 Straight onto Shore Road

12.3 Turn left onto School Street. (Do not proceed to Ogunquit Square because left turn onto U. S. 1 is generally prohibited.)

Mileage

12.6 Left turn onto U. S. 1.

22.8 Turn right into Rest Area access road. Retrieve vechicles. Return to U. S. 1. Road log resumes at Junction with U. S. 1. Turn right (south) onto U. S. 1.

23.7 Turn left on Haley Road.

25.1 Norton Road to left. Stay on Haley Road.

26.6 Turn right on Maine Route 103 at stop sign.

27.3 Turn right into Fort McClary picnic area.

Stop 2. Lunch (Fig. 6). Lunch materials for those who have not brought it may be obtained at Bisbee's Market 1/4 mile towards Kittery.

After lunch we will walk west to parking lot at Fort McClary State Memorial for the following:

(1) Comparison of fold styles in parking lot exposures.

(2) SW of the Fort along shore of Portsmouth Harbor, beneath the stone turret in overturned beds, note load flames, rip-up clasts, normal grading, a strongly cross-bedded unit.

(3) Note bedding variability while walking west.

In westernmost outcrop note:

(a) coarse-grained medium-bedded metaquartzites, load structures, slow grades, large-scale parallel laminations (beds upright);
(b) preserved ripples (beds upright);
(c) extremely thin bedded metaquartzites (overlying (a) and (b), beds overturned) with low metaquartzite/metapelite ratio.

Turn right on Route 103 out of picnic area parking lot.

28.7 Pass Naval Shipyard on left.

28.9 At light, turn left on Whipple Road. (Stay on Route 103.)

29.0 Bear left onto Williams Avenue.

Stop 3 (Fig. 6).

29.1 Harry H. Cook Memorial School. West of the school on the shoreline in massive thick-bedded overturned metaquartzites, note large-scale cross-lamination, scour features, and small-scale cross beds in upper part of bed.

Backtrack to Route 103; bear left onto Route 103.

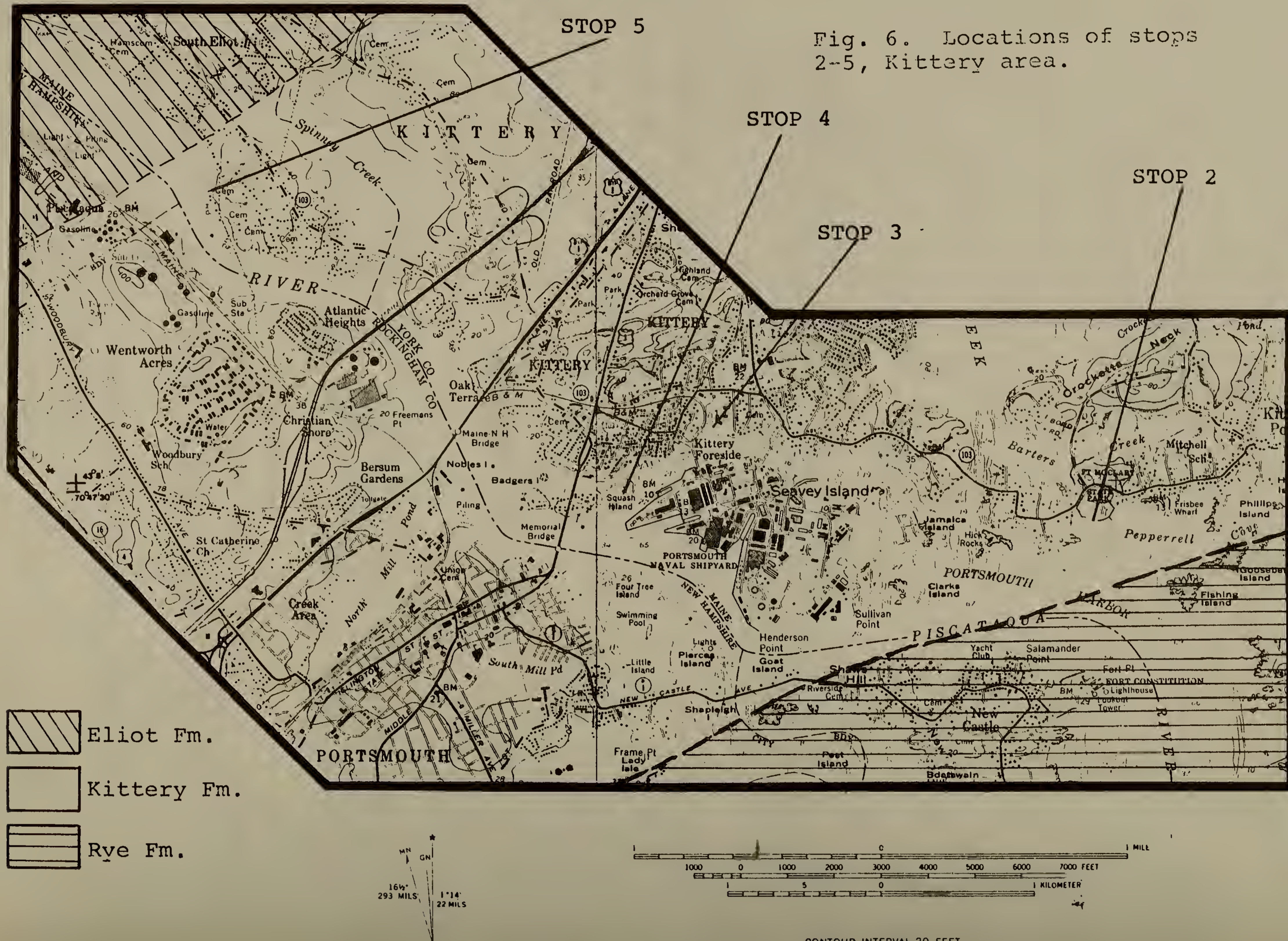


Fig. 6. Locations of stops 2-5, Kittery area.

Mileage

29.2 Bear left at stop sign (stay on Route 103).

29.4 Bear right at light (stay on Route 103).

29.8 Bear left onto U. S. 1 at light, continue until you cross bridge to Badgers Island.

30.2 Bear left onto Island Avenue. Park behind Chase's Minit Market, walk up Island Avenue to the east, cross through property of large white house (subsequent visitors must obtain permission!) to Squash Island (Fig. 6).

Stop 4. Outcrop is toward the eastern end of the island on the south side. A distinctive 2.5 mile-thick cross-bedded unit is interpreted as a contourite. Note sorting, grain size, consistency of cross-bed orientation, and lack of graded bedding.

Return to vehicles. Return to U. S. 1. Turn right on U. S. 1.

30.5 Bear left at first stop light (Government Street).

30.7 Straight at stop sign. Back on Route 103.

31.1 Bear left on South Eliot Road. (Stay on Route 103.)

32.1 Bear left on Pleasant Street.

32.7 Park behind Advent Christian Church on right.

Stop 5. (Fig. 6). Outcrop is SSE along the Piscataqua River. Note decrease in bed thickness, grain size, metaquartzite/metapelite ratio as compared to stops 3, 4, and 5. Compare this outcrop to outcrops of the Eliot Fm. 300 m to the northwest along shore.

-END OF TRIP-

IGNEOUS ROCKS OF THE NASHOBA BLOCK, EASTERN MASSACHUSETTS

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INTRODUCTION

Eastern Massachusetts is underlain by a distinctive sequence of igneous, metamorphic and sedimentary rocks quite different from those found in most of the rest of New England. The region can be subdivided into three northeast-trending belts (Fig. 1), separated by major fault zones (Zartman & Naylor, 1984). The easternmost block, underlain by Late Precambrian to lower Paleozoic rocks of the Boston Platform, has been identified as belonging to the fragmented Avalon microcontinent (Skehan & Rast, 1976; Rast et al., 1976; Rast, 1980). To the northwest, across the Bloody Bluff Fault zone, lies the Nashoba Block, a suspect terrane (Zen, 1983a) of generally high metamorphic grade with distinctly different geologic features. Farther west, across the Clinton-Newbury Fault zone lies the Merrimack Trough, a region with a geological and intrusive history that contrasts with both the Boston Platform and the Nashoba Block.

The stratified rocks of the Nashoba Block are largely high-grade metasediments and amphibolites in the mid-to-upper amphibolite facies. The rocks in the western portion of the area include the Tadmuck Brook Schist and schists and gneisses of the Nashoba Formation and the Fish Brook Gneiss (Bell & Alvord, 1976). Further east lies the Marlboro Formation, a thick sequence dominated by amphibolites. The absolute ages of all these formations, except the Fish Brook, are still in question and range from Ordovician to Precambrian (Zen, 1983a,b). Olszewski (1980) obtained a 730 m.y. U-Pb age from volcanic zircons in the Fishbrook, and our Nd isotope studies of the Marlboro Fm. metabasalts suggests they formed 450-550 m.y. ago (see discussion in DiNitto et al., this volume).

The igneous rocks of the Nashoba Block contrast strongly with those of the Boston Platform. Neither the large, Late Precambrian intrusions (Dedham Granodiorite, Milford Granite) nor the Ordovician to Devonian alkaline intrusions (Quincy, Peabody, Cape Ann Granites) which characterize the Boston Platform are present in the Nashoba Block. In contrast, the Nashoba Block contains abundant metabasaltic flows in the stratified sequence (Marlboro Fm. and Boxford member of the Nashoba Fm.) and was intruded by Ordovician to Silurian calc-alkaline plutons (Sharpners Pond Diorite, Straw Hollow Diorite, Assabet Quartz Diorite, older dioritic phase of the Indian Head Hill pluton) and deep-seated peraluminous granites (different phases of the Andover Granite, perhaps extending over a 40 m.y. period - see below).

Roughly half of the exposed Nashoba Block is underlain by igneous or meta-igneous rocks (Fig. 2); deciphering the evolution of this terrane requires that we understand why the magmas formed and the nature of their source materials. The units to be visited on this trip are briefly outlined below; the stops are shown in circles on Fig. 2.

MARLBORO FORMATION

The Marlboro Fm. is dominated by amphibolite with subordinate pelitic schist

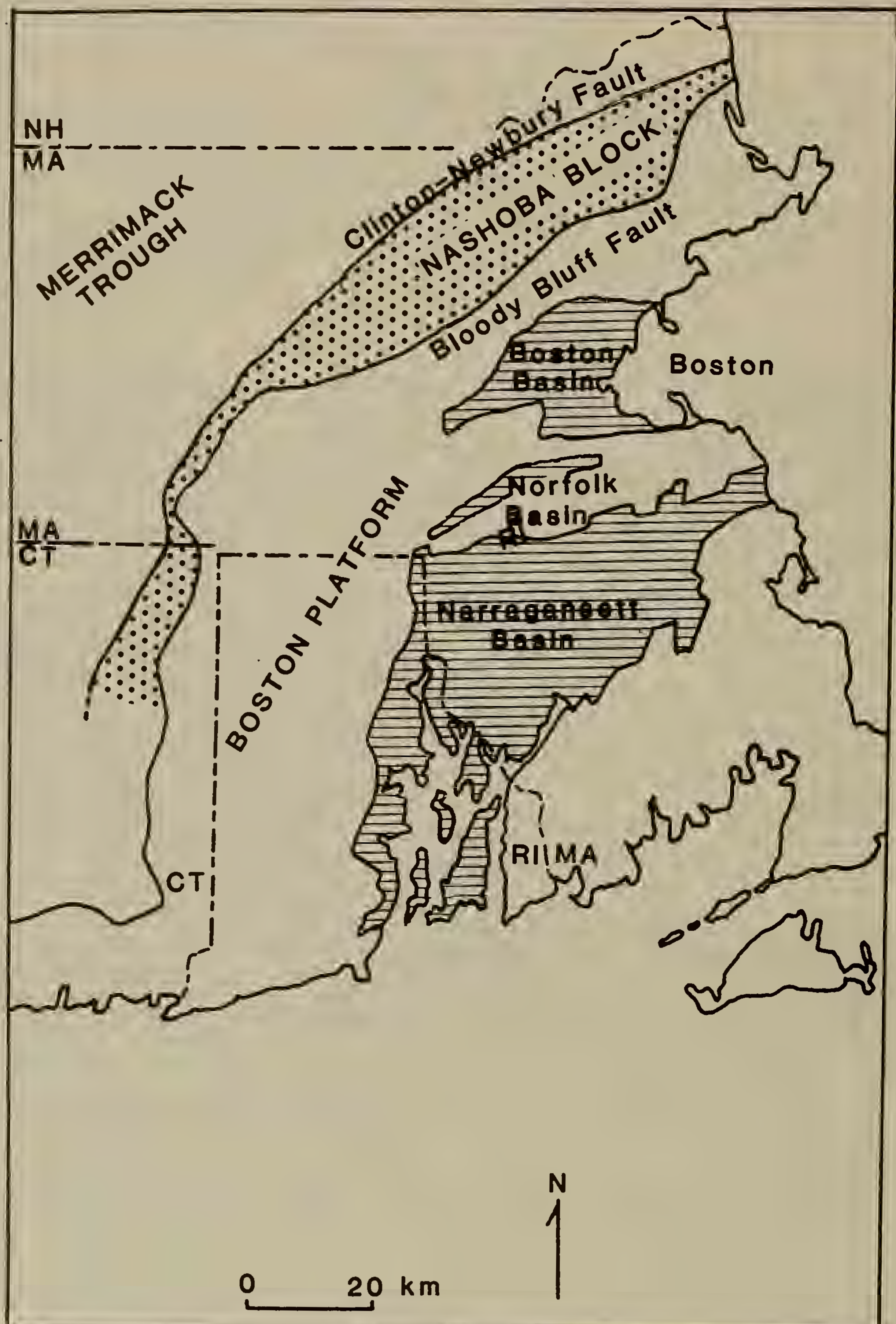

Figure 1. Location of the Nashoba Block, eastern Massachusetts. Sedimentary basins of the Boston Platform shown by lined pattern: Boston Basin, Late Proterozoic-Early Paleozoic; Narragansett and Norfolk Basins, Carboniferous. Outlined square shows area of Figure 2.

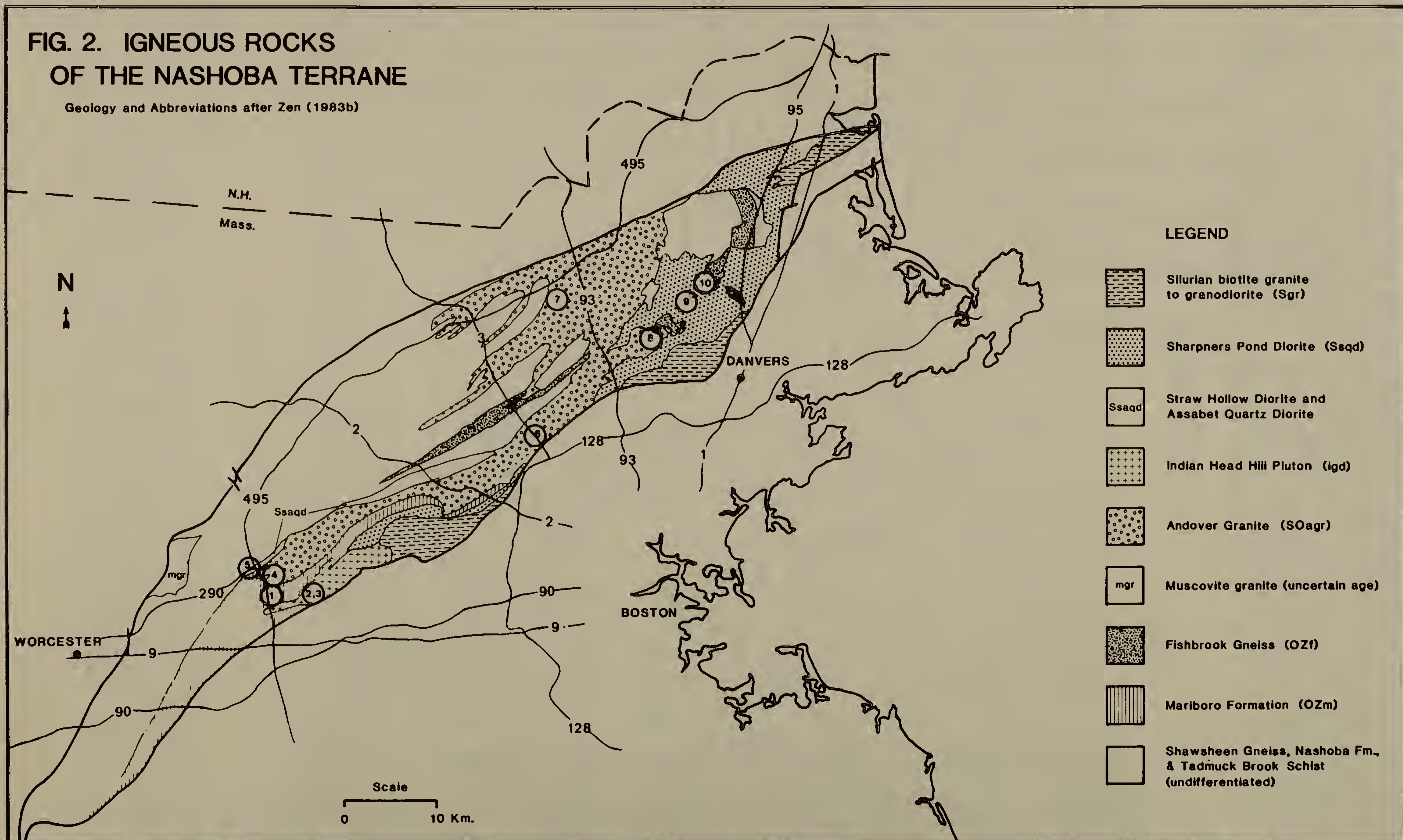

FIG. 2. IGNEOUS ROCKS OF THE NASHOBA TERRANE

Geology and Abbreviations after Zen (1983b) and felsic gneiss; DiNitto et al. (this volume) have subdivided the Marlboro into 5 members. The results of our geochemical studies of the Marlboro Fm. amphibolites are described in the trip by DiNitto et al., and are not repeated here. Briefly, they show that the amphibolites are basaltic in origin and most likely formed in either an arc or marginal basin setting during the time period 450 - 550 m.y. ago.

DIORITIC INTRUSIONS

Intermediate composition plutons form a prominent part of the Nashoba Block (Fig. 2). They define a typical calc-alkaline trend (Fig. 3). Zartman & Naylor (1984) obtained a 430 m.y. age for the Shaprners Pond Diorite; we have determined a Rb-Sr whole rock age of 402 m.y. for the older dioritic phase of the Indian Head Hill pluton (Fig. 4).

Castle (1964) recognized three gradational phases of the Sharpners Pond Diorite - hornblende diorite (oldest), biotite-hornblende tonalite, and biotite tonalite (youngest); aplite dikes are common. Two or more of these phases commonly coexist at most outcrops; not uncommonly, older, more foliated melanocratic diorite is intruded by younger, less foliated, more leucocratic tonalite (Stop #9). The Straw Hollow Diorite at Stop #5 contains at least two phases - an older, more foliated, finer-grained phase is intruded by a younger, more leucocratic phase.

REE (Figs. 5 and 6) and Nd isotope data (Fig. 7) confirm this heterogeneity. In general, the REE contents of these intrusions are quite similar to those in plutons of similar bulk composition developed at convergent plate boundaries (e.g., Sierra Nevada, Noyes et al., 1983). On the 143/144 Nd evolution diagram (Fig. 7), note the generally non-intersecting lines for samples within the same pluton; this isotopic heterogeneity probably arose from incomplete mixing of assimilated crust during ascent of the magma. For example, the two Straw Hollow curves (solid circles on Fig. 7) do not intersect except at an unreasonably old age; the field relations of these two samples can be studied at Stop #5. The upper point represents the older, darker phase at Stop #5; it is intruded by a slightly lighter colored, coarser grained phase which shows evidence (both Sr and Nd isotopes) of being more contaminated with older crustal material than the older diorite. (An evolution curve for a granulite facies metasedimentary xenolith from the Boston Platform - Hill & Ross (1983) - is shown for reference. This type of material makes a satisfactory contaminant.) The Sharpners Pond Diorite is similarly heterogeneous. A plausible explanation for these relationships is that 430-400 m.y. ago, mafic magmas with 87/86 Sr = 0.7035 and 143/144 Nd = 0.51250 developed at a convergent plate boundary and assimilated varying proportions of Proterozoic to early Paleozoic crustal rocks. Olszewski (1980) has documented the existence of such Proterozoic rocks in the Nashoba Block.

ANDOVER GRANITE

Castle (1964) used both mineralogical and textural criteria to divide the Andover Granite into six facies: muscovite granite-gneiss, biotite granite-gneiss, fine-grained granite gneiss, undifferentiated granite gneiss, binary (2 mica) granite, and pegmatitic granite. Wones & Goldsmith (in prep.) distinguish an older, more foliated Andover from a younger, more pegmatite-rich Andover. The older phase(s) of the Andover may predate or be coeval with the Sharpners Pond Diorite; Castle (1964) described both gradational contacts between the Andover and Sharpners Pond, as well as xenoliths of Sharpners Pond within the

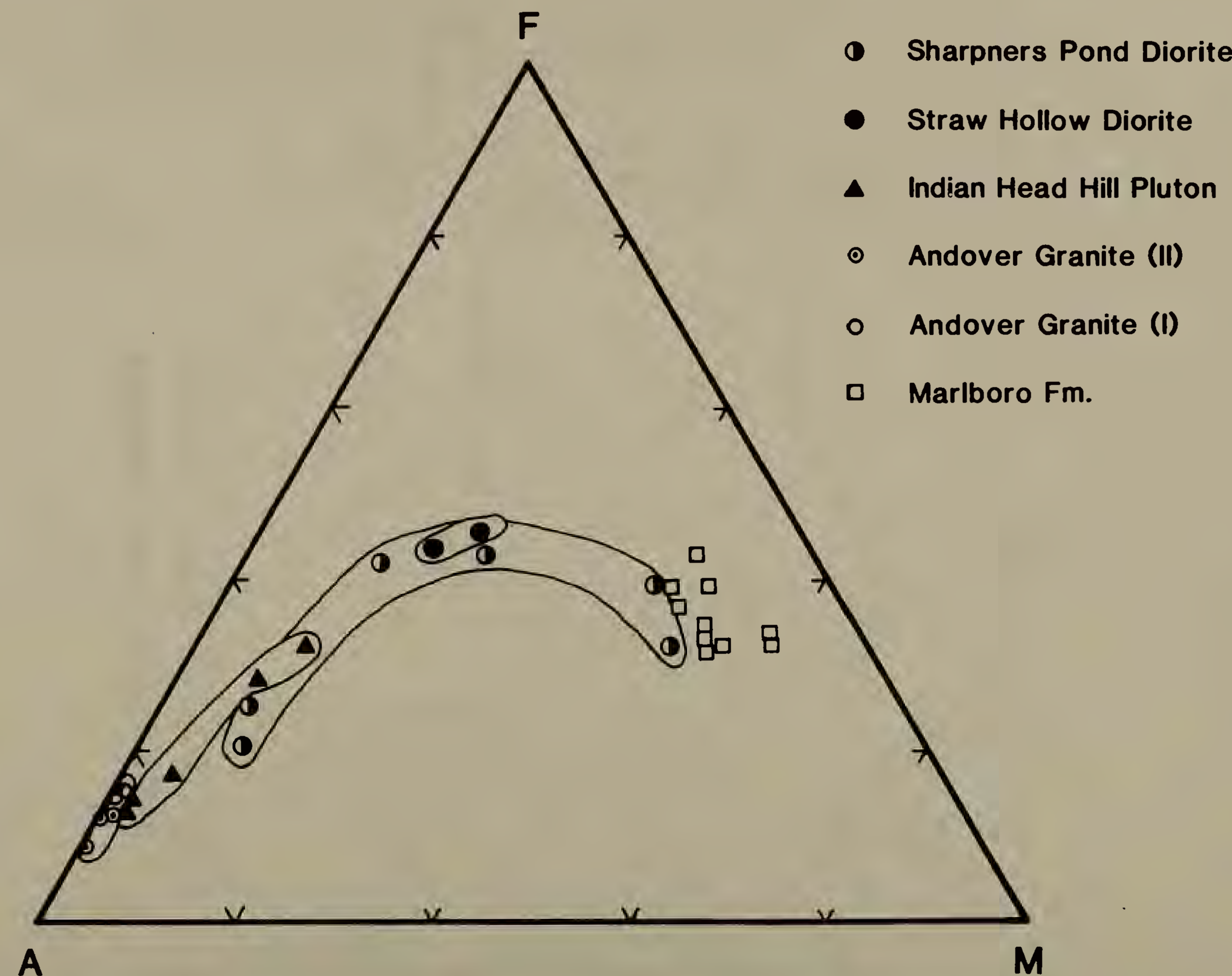

Fig. 3 AFM Diagram, Nashoba Block

Indian Head Hill Pluton
(Older, foliated phase)

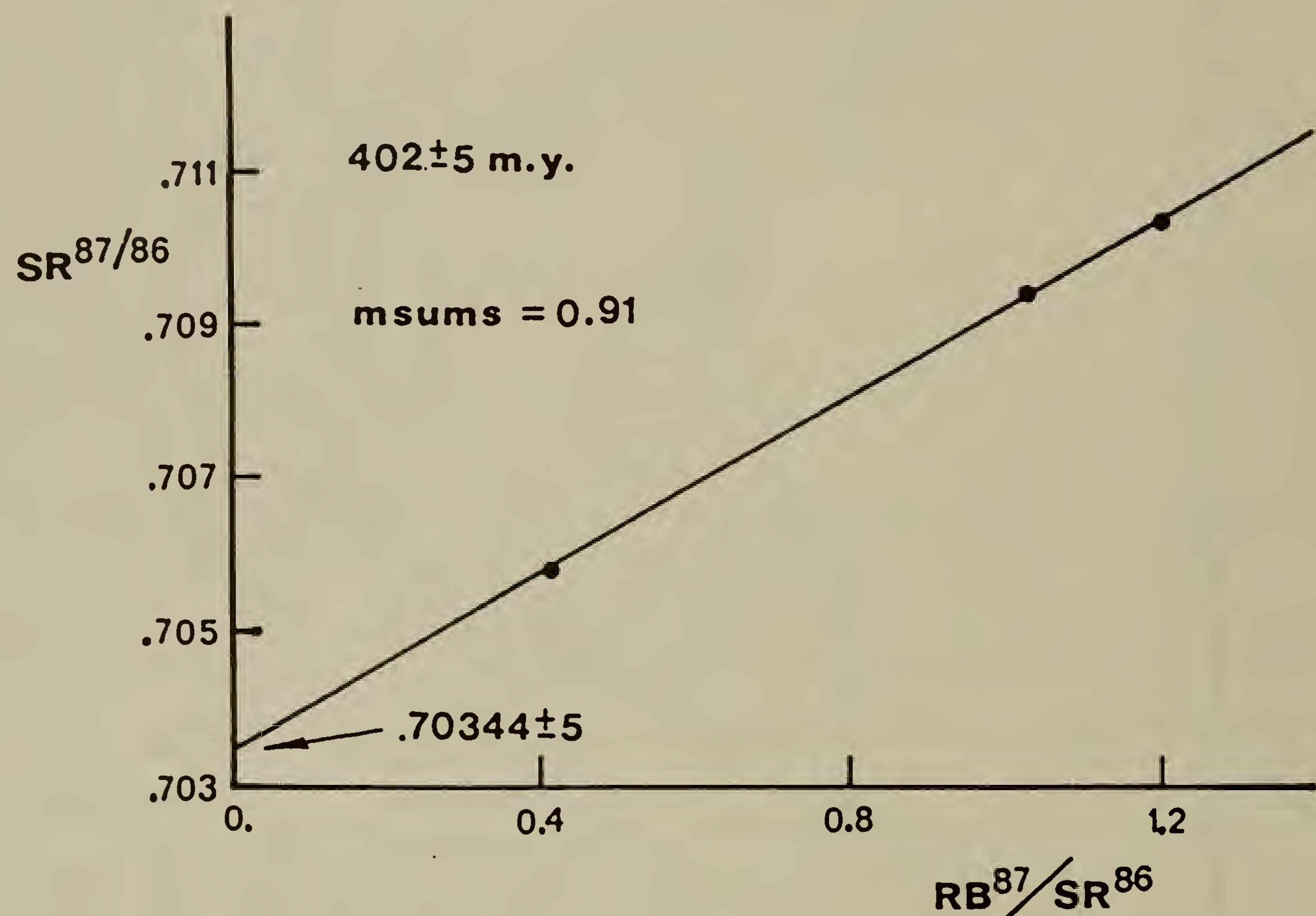

Fig. 4

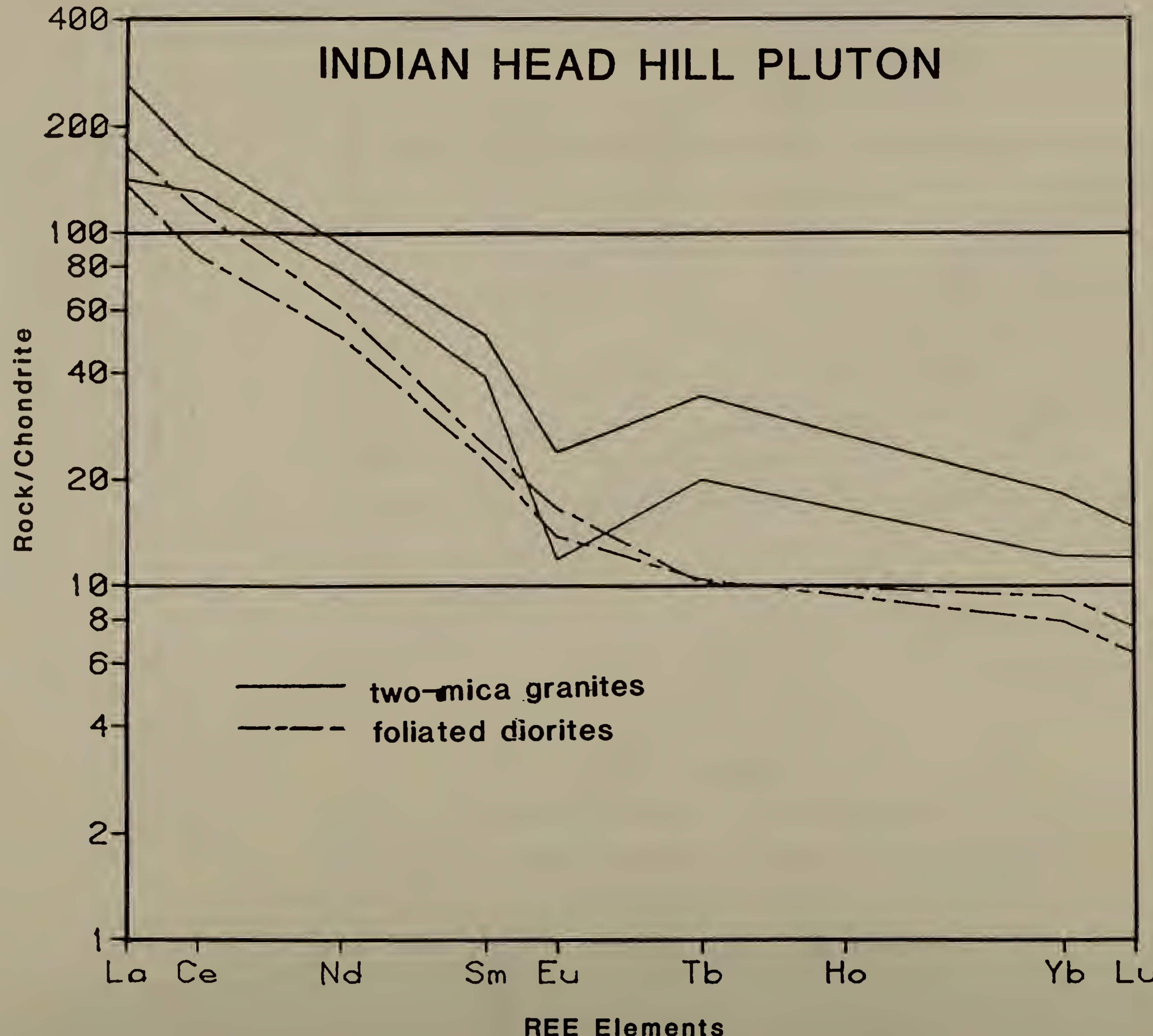
INDIAN HEAD HILL PLUTON
400
200
100
80
60
40
20
10
8
6
4
2
1
Rock/Chondrite
La
Ce
Nd
Sm
Eu
Tb
Ho
Yb
Lu
REE Elements
two-mica granites
foliated diorites

Fig. 5

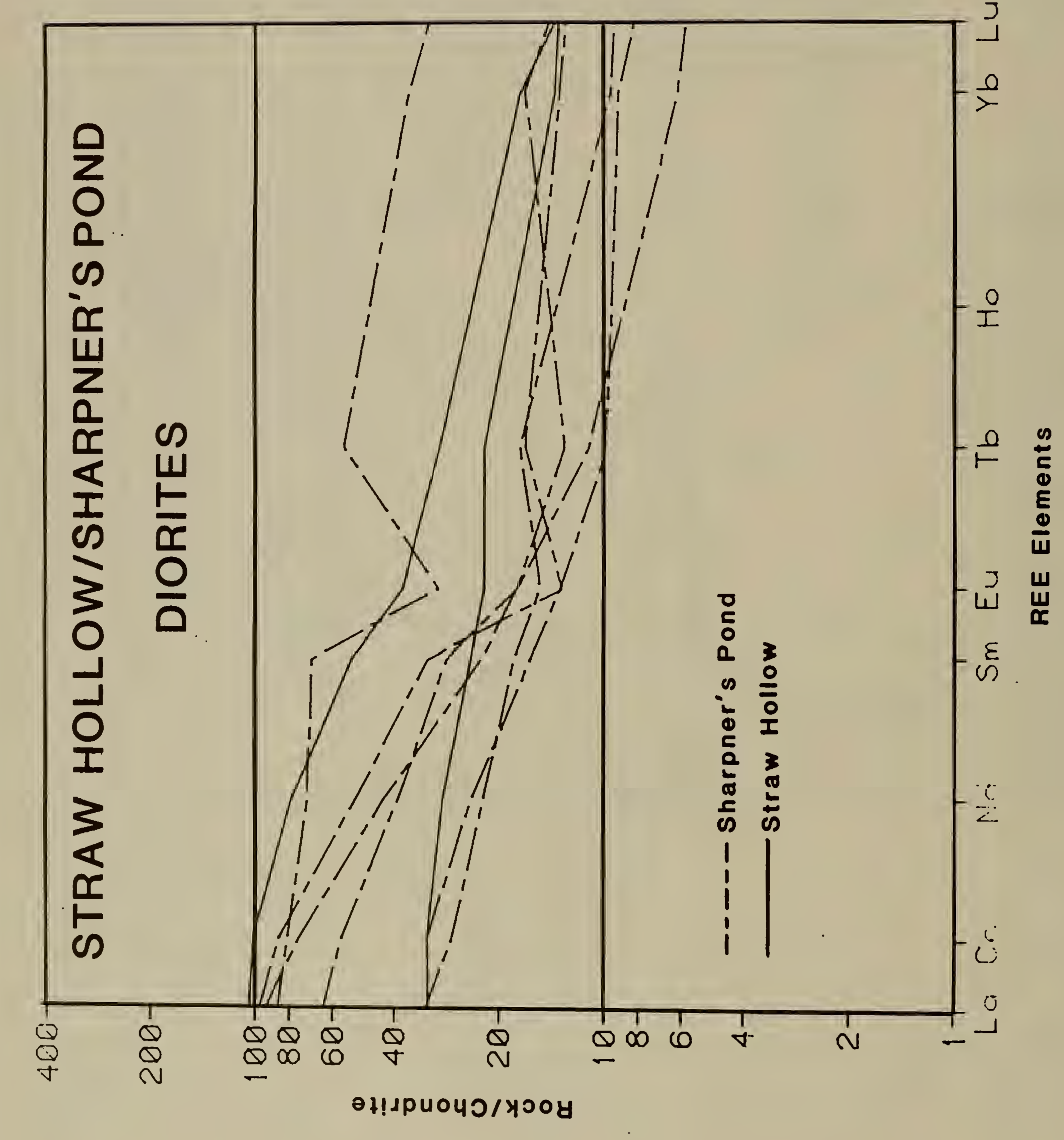
STRAW HOLLOW/SHARPNER'S POND
DIORITES
Rock/Chondrite
400
200
100
80
60
40
20
10
8
6
4
2
1
La
Ce
Nd
Sm
Eu
Tb
Ho
Yb
Lu
Sharpner's Pond
Straw Hollow
REE Elements

Fig. 6

Nd EVOLUTION DIAGRAM NASHOBA BLOCK PLUTONS

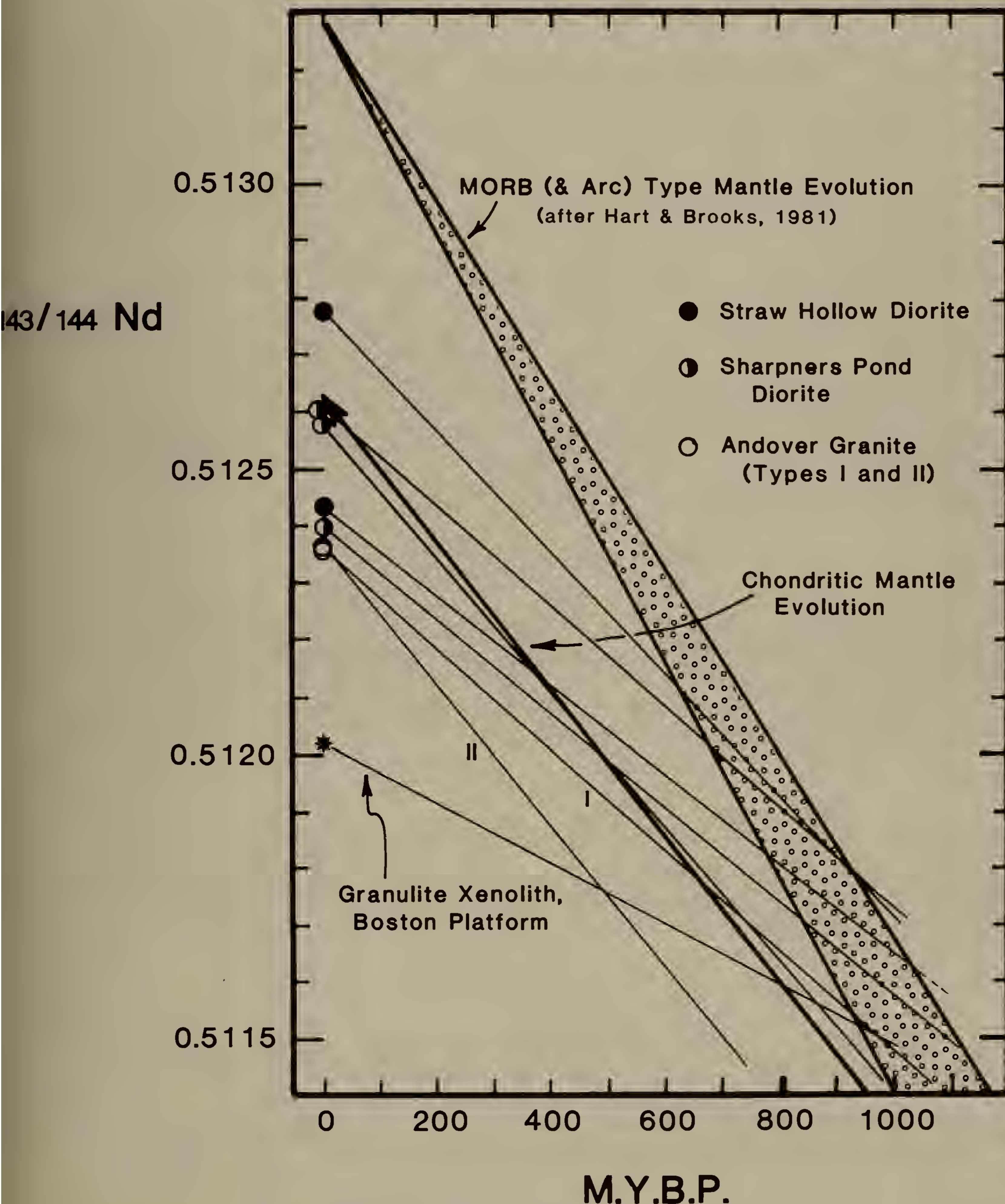

Fig. 7

less foliated phase of the Andover (see Stop #8).

Our data suggest that there are (at least) two fundamental units within the Andover - herein designated Andover I and Andover II. Type I Andover is a mildly peraluminous biotite-muscovite-garnet granite while type II is a strongly peraluminous muscovite-garnet granite. Petrographic evidence suggests that type II contains a greater restite component than type I (abundant sillimanite/muscovite reactions, etc.). The two types are geochemically quite distinct - on the AFM diagram (Fig. 3) type II plots closer to the alkali apex than type I; type II has a much flatter REE pattern (Fig. 8), consistent with it containing a higher restite component. Fig. 9 is a composite Rb-Sr isochron plot containing our data plus those of Zartman & Naylor (1984) and Handford (1965). While some scatter is clearly evident, it is interesting to note that most of the data fall close to one of two limiting isochrons, one with an age = 455 m.y. and the other 415 m.y. Both give essentially the same initial 87/86 Sr = 0.7043 - 0.7044. These ages are consistent with the field relations described above, with Zartman & Naylor's (1984) age of 430 m.y. for the Sharpners Pond Diorite, and with their suggested range of ages for the Andover.

The differences between types I and II extend to Nd isotope systematics as well (Fig. 7). The type II Andover sample has a flat REE pattern, thus a relatively high 147Sm/144Nd ratio, which forces 143/144 Nd to evolve rapidly in time (steep slope on Fig. 7). 450 m.y. ago this sample had 143/144 Nd similar to the granulite facies metasedimentary xenolith from the Boston Platform, supporting the argument that type II contains a greater restite component than type I. The type I sample has a model age (time at which the evolution curve for the rock intersects that for a model chondritic mantle) of 650 m.y., providing additional evidence for a significant quantity of at least Late Proterozoic crustal detritus in the source for the Andover (Nashoba Fm.?).

FISH BROOK GNEISS

The oldest known formation in the Nashoba Block is the Fish Brook Gneiss, a heterogeneous unit dominated by leucocratic gneisses (Bell & Alvord, 1976) displaying a characteristic "swirled foliation" (Zen, 1983b)(Stop #10), whose protolith is inferred to be tuffaceous. Olszewski (1980) dated zircons of igneous morphology (including samples collected at Stop #10) and obtained an age of 730 m.y., proving the existance of Proterozoic basement under at least part of this terrane.

ACKNOWLEDGEMENTS

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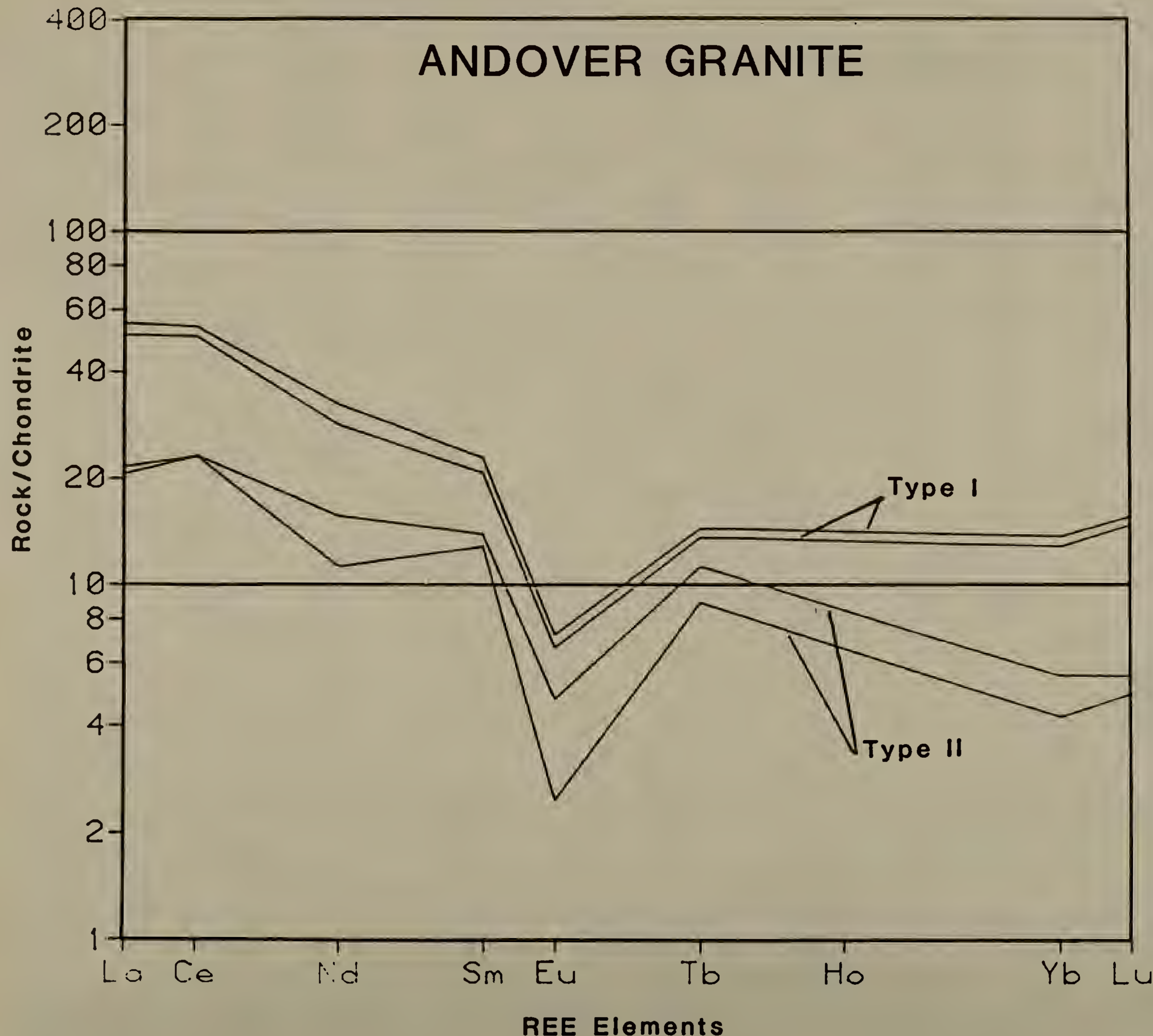
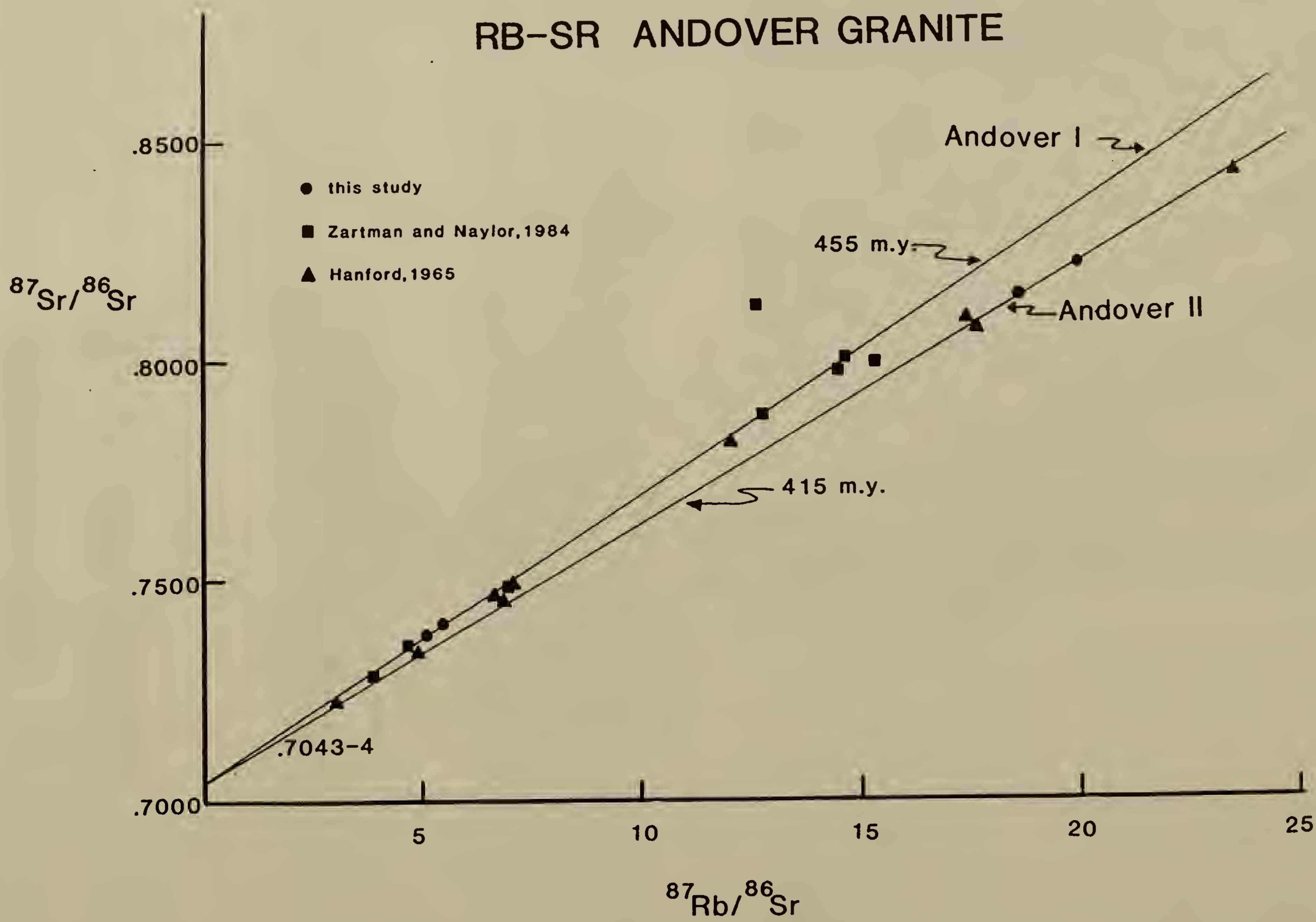

ANDOVER GRANITE
400
200
100
80
60
40
20
10
8
6
4
2
1
Rock/Chondrite
Type I
Type II
La
Ce
Nd
Sm
Eu
Tb
Ho
Yb
Lu
REE Elements

Fig. 8



RB-SR ANDOVER GRANITE
this study
Zartman and Naylor,1984
Hanford,1965
$^{87}Sr/^{86}Sr$
.8500
.8000
.7500
.7000
Andover I
455 m.y.
Andover II
415 m.y.
.7043-4
5
10
15
20
25
$^{87}Rb/^{86}Sr$

Fig. 9

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ROAD LOG

MEETING SPOT and STOP #1. Parking lot of Holiday Inn at junction of Rte. 20 and I-495 in Marlborough, MA. Examine outcrops of Marlboro Fm. amphibolite (Sandy Pond Member) along the east side of the parking lot. Poison ivy abounds. The amphibolite is tightly folded; micaceous horizons are crenulated.

Look for coticules: siliceous layers composed of spessartine, quartz, albite and magnetite; the protolith for these may have been Mn-cherts. This sequence has been cut by garnet-tourmaline-muscovite-bearing pegmatites, provisionally related to the Andover Granite; note that the pegmatites have been sheared.

Mileage in the left-hand column will be cumulative from this location (cum.); and the right-hand column will record mileage from site to site (s/s) noted in the text of the log.

cum.	s/s	
0.0		Exit Holiday Inn parking lot, turn left (east) on Rte. 20.
1.5	1.5	Proceed through the town of Marlborough, straight on Rte. 20. Type locality of the Marlboro Fm. on left (north) side of Main Street (Emerson, 1917). See DiNitto et al. (this volume) for additional details of this locality.
1.6	0.1	Junction with Rte. 85; continue straight on Rte. 20.
1.7	0.1	Left turn, follow Rte. 20.
1.8	0.1	Marlboro Fm. amphibolite on left side of road.
2.1	0.3	Right turn, follow Rte. 20.
3.9	1.8	Traffic lights; continue straight on Rte. 20.
4.3	0.4	Turn right at entrance to shopping mall (Zayre's); go to outcrops at right side of parking lot, behind the Shawmut bank. STOP #2. This location is at the north end of Indian Head Hill, the type locality of the INDIAN HEAD HILL PLUTON. Originally, Emerson (1917) mapped this as part of the Dedham Granodiorite, but it is quite different from the Dedham; Hepburn & DiNitto (1978) designated it as a separate unit. The pluton is composite and contains an older, foliated dioritic phase intruded by a non-foliated biotite granite. The latter phase is exposed here, cut by pegmatites. Observe the pegmatites, and contrast them with those clearly associated with the Andover, to be seen later in the day. Despite the proximity to the Bloody Bluff fault, this granite is undeformed.
4.5	0.2	Return to entrance to parking lot, turn left (west) on Rte. 20.
4.9	0.4	Traffic lights (at Mobil Station); turn left on Farm Rd. Pass small airport on left.
5.3	0.4	Turn left on Broad Meadow Road. Continue straight to a fork in the road.
5.9	0.6	At the fork, bear left towards the entrance to Gulbankian Mobile Home Village.

6.0 0.1 Turn left into the main entrance (not the office) of the trailer park. Continue straight, to a cross roads.

6.2 0.2 Left at cross street.

6.3 0.1 T-intersection, outcrop on left beside mailboxes. PARK ON THE BLACKTOP. STOP #3. NO HAMMERS PLEASE. The owners have always granted permission to study these outcrops, but be sure to ask first. These rocks constitute part of the older, dioritic phase of the Indian Head Hill pluton; a 3 point Rb-Sr isochron yields an age of 408 m.y. (Fig. 3). The older diorites have been cut by aplites, and later by the fine grained, granite seen at the last stop. Return to main road.

6.4 0.1 Right turn on main road.

6.5 0.1 Right turn onto Broad Meadow Road.

6.6 0.1 Right turn takes you to Office of Gulbankian Mobile Home Village If time permits, we will examine outcrops beside the gray house. Do not block buildings. STOP 3A. Good pavement exposures of the fine-grained granite intruding foliated diorite, as at Stop 3. Return to Broad Meadow Rd.

6.8 0.2 Right turn on Broad Meadow Rd.

6.9 0.1 Junction with Perimeter Rd.; continue straight on Broad Meadow Rd.

7.4 0.5 Turn left (west) on Farm Road.

7.8 0.4 Turn right on Phelps St.; go to Rte. 20.

8.4 0.6 Turn left on Rte. 20 (west). Continue through the town of Marlborough, heading for I-495.

9.3 0.9 Left turn, follow Rte. 20.

9.6 0.3 Right turn; follow Rte. 20.

10.1 0.5 Bear left; follow Rte. 20.

11.4 1.3 Junction with I-495; take I-495 North towards Lowell-Lawrence. Prepare to pull over just past the first bridge.

11.9 0.5 STOP #4. ANDOVER GRANITE, containing garnet, muscovite and sillimanite, cut by pegmatites. This location is near the southernmost exposure of the Andover Granite. Contrast the mineralogy and structure of this outcrop with that of the youngest granite in the Indian Head Hill pluton seen in the previous two stops.

Continue north on I-495.

12.9	1.0	Exit I-495 north at Exit-25A, "Hudson, To 85". Prepare to pull off on right side at end of exit ramp.
13.3	0.4	Park on the right just past the electrical box. STOP #5. STRAW HOLLOW DIORITE. We will spend roughly one hour at this stop; we could easily spend a day. This area is cut by a splay of the Assabet River fault zone; look for evidence of both deep (blastomylonite) and late shallow (breccia) faulting. Does the Straw Hollow intrude the blastomylonite, or is the mylonite developed from the Straw Hollow? The Straw Hollow contains at least two phases, a finer grained, more foliated phase intruded by a coarser grained, lighter colored, less foliated phase. These units are geochemically distinct (see discussion in text). Both have been cut by granitic pegmatites and aplites (presumed to be related to the Andover Granite), which have themselves been extensively deformed. In addition, the Straw Hollow contains gabbroic pegmatite segregations genetically related to the diorite. (The curbstones along the highway are the ubiquitous Chelmsford Granite, from the Merrimack Trough - the next tectonic block to the west.) Continue east towards Rte. 85.
14.3	1.0	Traffic light, intersection with Fitchburg St.; continue straight. (Fitchburg St. to the right leads to the Assabet Valley Regional Vocational School, which has a large outcrop of foliated Andover Granite next to the main entrance.)
14.7	0.4	Turn right on Rte. 85 South.
16.0	1.3	Marlborough High School on left.
16.5	0.5	Traffic lights; straight through.
16.8	0.3	Traffic lights; turn left on Lincoln St. (east).
17.1	0.3	Traffic lights; straight through, now on Rte. 20 (east). Continue on Rte. 20 to Weston.
18.9	1.8	Traffic lights, Mobil station on right; straight through.
20.2	1.3	Traffic lights; straight through.
20.3	0.1	Sudbury town line.
24.8	4.5	Bear left, stay on Rte. 20. Wayland town line.
27.1	2.3	Continue east on Rte. 20.
28.7	1.6	Weston town line, sign on right.
29.6	0.9	Left turn on Boston Post Rd., toward Weston Center.
30.2	0.6	Left turn, past Gulf station, on Concord Rd.

30.6	0.4	Road crosses railroad tracks (bridge).
31.0	0.4	Bear left, stay on Concord Rd.; sign "To Campion Center".
31.7	0.7	Campion Center on right.
31.8	0.1	Bear right.
31.9	0.1	Right turn into Weston Observatory of the Dept. of Geology and Geophysics, Boston College, then left to Observatory parking lot. LUNCH STOP.
32.1	0.2	Exit Weston Observatory, left turn.
32.2	0.1	Left turn on Concord Rd.
33.0	0.8	Bear right, stay on Concord Rd.
33.9	0.9	Bear left, through Weston Center.
34.6	0.7	Left turn on Rte. 20 (east).
35.8	1.2	Left turn to Rte. 128 (north). Continue north on Rte. 128 to Rte. 3 (north).
44.3	8.5	Take exit 43N to "Lowell - To Rte. 3N". Continue on Rte. 3 (north).
46.8	2.5	Take exit 26 - "Rte. 62 - Bedford - Burlington".
47.0	0.2	Turn left (west) on Rte. 62. Prepare to stop.
47.3	0.3	Outcrop of Andover Granite.
47.4	0.1	Park either on south side of Rte. 62 across from Earl Rd.,or else along Earl Rd. STOP #6. ANDOVER GRANITE. Foliated Andover 2-mica granite, with massive dioritic inclusions, cut by garnet-apatite bearing pegmatites. Continue east on Rte. 62.
47.8	0.4	Take Rte. 3 (north) toward Lowell.
51.8	4.0	Rest area.
55.9	4.1	Take Exit 30N to the Lowell Connector. (Note - due to construction-imposed detours at the time this road log is being compiled, the mileage at this point will be different from that on the trip. The log should regain reliability at Exit 38 from I-495, to Rte. 38 south.)
56.8	0.9	Detour for I-495 south.

58.7	1.9	Rejoin I-495 north.
61.2	2.5	Take Exit 38 from I-495, to "Lowell - Tewksbury - Rte. 38".
61.4	0.2	Left turn (south) to Tewksbury.
63.5	2.1	Traffic lights; straight.
63.7	0.2	Left turn on North Street.
63.8	0.1	Straight through intersection, continue on North Street.
64.0	0.2	North Street School on right.
64.3	0.3	Railroad tracks; parking for next stop if many cars on trip.
64.4	0.1	Outcrops of ANDOVER GRANITE under power lines. STOP #7. Foliated, massive garnet-bearing Andover Granite; contrast with previous stop. Cut by pegmatites and finer grained granite.
		Continue south on North Street towards Rte. 38.
65.3	0.7	Stop sign; continue straight.
65.4	0.1	Turn left (south) on Rte. 38.
65.8	0.4	Traffic lights; continue south on Rte. 38.
67.6	1.8	Traffic light; straight through, continue on Rte. 38.
68.3	0.7	Left turn on Salem Rd. (east).
68.4	0.1	Cross South Street; continue on Salem Rd.
69.7	1.3	Cross railroad tracks.
69.9	0.2	Cross railroad tracks.
70.3	0.4	Turn right on Middlesex Ave. (south).
70.6	0.3	Turn left on Rte. 62 (east).
71.0	0.4	Cross over I-93.
71.3	0.3	Intersection; continue straight on Rte. 62.
71.9	0.6	Gravel pit on left.
72.8	0.9	Turn left on North Street.
72.9	0.1	M. Murphy School on left.

73.2	0.3	Traffic lights; straight through on North Street - cross Main St.
73.7	0.5	Hillview Country Club on left.
73.9	0.2	Flashing light - straight through on North Street.
74.1	0.2	Bear right, stay on North Street.
75.0	0.9	T-intersection; turn right on Haverhill Street.
75.3	0.3	Turn left (west) on Aspen Rd.
75.5	0.2	Take the first left onto Colonial Hill Rd.
75.6	0.1	Outcrop on left - <u>STOP #8</u>. Small quarry in housing development shows 2-mica granite (Andover) intruding diorite (Sharpners Pond). Loose blocks near road show best relations. Reverse direction in cul-de-sac.
75.8	0.2	Turn right on Aspen Rd.
76.0	0.2	Turn left on Haverhill St. (will soon become Jenkins St.). Proceed north on Haverhill St.
76.1	0.1	Intersection with Marblehead Rd., flashing yellow light. Straight through; stay on Jenkins St.
78.1	2.0	Intersection - straight through.
79.7	1.6	Turn right (south) on Salem Turnpike - a fast, two-lane road. In 2 miles you will make a left turn off this road.
81.9	2.2	Turn left (east) on Sharpners Pond Road (yes, there is a Sharpners Pond).
82.5	0.6	Outcrop of <u>SHARPNERS POND DIORITE</u> on both sides of road. <u>STOP #9.</u> This is a typical example of the Sharpners Pond, aplites and felsic diorite cut foliated, more mafic diorite.
		Continue east on Sharpners Pond Rd.
83.7	1.2	Park at guardrails blocking straight continuation of road. Walk 1/4 mile along dirt trail to excavations which expose <u>FISH BROOK GNEISS</u> - <u>STOP #10</u>. These exposures typify the "swirled foliation" characteristic of the Fish Brook. Although Bell & Alvord (1976) included the Fish Brook in the stratigraphic sequence of the Marlboro/Nashoba, Olszewski (1980) obtained zircons from this and other Fish Brook localities which yielded Proterozoic (730 m.y.) ages. If the protolith to the predominant swirled gneiss is waterlain tuff (Bell & Alvord, 1976) and the zircons are igneous (Olszewski, 1980), then the Fish Brook Gneiss represents the oldest (meta)igneous unit so far documented in the Nashoba Block. Return to cars.

		Proceed left on paved road past parking area.
84.7	1.0	Turn right on Lacy St.
86.4	1.7	Turn right on Lawrence St. (east).
86.8	0.4	Turn right on Main St.
87.7	0.9	Turn left on Middleton Rd. towards East Boxford.
88.1	0.4	Turn right towards Topsfield.
88.2	0.1	Turn left towards Boxford Village.
89.4	1.2	Turn right onto I-95 (south).
91.0	1.6	Outcrop on right of Sharpners Pond Diorite with multiple injection dikes, just before exit to Endicott Rd.
95.4	4.4	Take Centre St. Exit to Danvers.
95.7	0.3	Outcrop of Salem Gabbro-Diorite - a major component of the Boston Platform, but not found anywhere in the Nashoba Block. This outcrop is easily accessible from the parking lot of NEIGC headquarters, and may be contemplated in comfort from the lounge.
95.8	0.1	Turn right to Dayton St., then left on Dayton to motel entrance.
96.0	0.2	Lobby of the Inn at Danvers (Best Western) - NEIGC '85 headquarters.

MAFIC DIKES FROM BOSTON TO CAPE ANN

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Introduction

From Boston north to Rockport, Massachusetts (Figure 1), hundreds of mafic dikes ranging in thickness from a few millimeters to at least 122 meters intrude rocks of Precambrian to Devonian age. The ages of the dikes are uncertain but field relationships suggest they range in age from Precambrian to Mesozoic (La Forge, 1932; Billings, 1976; Ross, 1981a; Kaye 1983). Whole-rock K-Ar ages determined for 11 of the dikes range from 190±6m.y. to 383±23m.y. (Zartman and others, 1970; Ross, 1981b; Krueger, unpublished data). The principal objective at this stage of this investigation is to attempt to group the dikes into petrologic suites on the basis of field evidence (primarily textures, cross-cutting relationships, and strikes), similar major and minor element chemistry, and similar petrographic character. Additional K-Ar age dates are planned to assist in unraveling the relative and absolute ages of the groups.

The main petrographic types recognized so far include compositional and textural varieties of dolerite and altered dolerite, camptonite, and hyalo-monchiquite (Ross, 1982, 1983). These can be grouped into tholeiitic and alkaline suites on the basis of their whole-rock major and minor element chemistries (Ross, 1982, 1983). The majority of the dolerites trend NW but lesser NE and E-W trends are evident and both lamprophyres trend NE (Ross, 1982, 1983a). This suggests that 3 or 4 distinct dike swarms of different ages and trends exist, with each swarm consisting of a range of chemical/petrographic types.

Petrography

Nine petrographic types have been identified among the 103 dikes thus far examined in thin section. Of these , 45 are dolerites (8 plagioclase-phyric), 52 are altered dolerites (19 plagioclase-phyric), 2 are plagioclase megacryst-rich and biotite-bearing altered dolerites; 1 is the Medford biotite-bearing dolerite; 1 is a xenolith-,xenocryst-, and megacryst-rich camptonite; 1 is an olivine-phyric camptonite; and 1 is a hyalo-monchiquite.Their petrographic properties are summarized in Table 1. For a more complete description see Ross, 1981b. Selected modal analyses are listed in Table 2. The distinction between altered and fresh dolerite is somewhat arbitrary. Virtually all the dikes show at least slight deuteric alteration (saussuritization, ±uralitization and olivine alteration to iddingsite, bowlingite, or smectite ± calcite). Fresh dolerites show only very slight to moderate amounts of saussurite and uralite $\pm$ hydrothermal alteration. The altered dolerites show moderate to total alteration of plagioclase, pyroxene, and total alteration of olivine when present.

Dike Chemistry

Fifty-six of the over 100 dikes sampled have been analyzed for major and minor element composition by whole-rock X.R.F. analyses of fused glass beads. The mean chemical compositions of the various petrographic types are listed in Table 3. A breakdown of the distribution of tholeiitic and alkaline types among the petrographic varieties is shown in the first two columns of Table 1.

The broad range in compositions (Table 3, Figure 2) is a function of the presence of dikes and dike swarms of vastly different ages derived from different sources. Substantial variations within a single swarm of dikes is also to be expected and even within an individual, thick differentiated dike such as the Medford dolerite dike (Figure 2). Similarites and differences in major and minor element chemistries, in themselves, are insufficient criteria for distinguishing one dike group or individual dike from another. Successful correlations are only possible when field, petrographic and chemical (including trace-elements) criteria are used. These can be refined by K-Ar age dating and perhaps paleomagnetic studies also.

With the above complications in mind, and the fact that 56 analyses are a small sample (more in progress), of the hundreds of dikes present in the study area, some generalizations are possible. Forty-one of 56 dikes (including 2 lamprophyres) plot in the tholeiitic field on an alkali-silica diagram (Figure 2). An even greater proportion of the fresh dolerites (25 of 28) are tholeiitic whereas the altered dolerites are more evenly divided (19 tholeiitic, 17 alkaline, Table 1). The above includes porphyritic varieties. Most of the tholeiites are quartz normative but a few are olivine normative. Most of the alkaline dikes are olivine normative.

The effect of alteration on the chemistry of the altered dolerites is uncertain. Altered samples of basalt flows from the Hartford and Newark Basins contain less CaO, MgO, and SiO_2, and more Fe^{3+}, Na_2O, and H_2O than their relatively unaltered counterparts and perhaps reflect low-temperature interaction of basalt and seawater $\pm$ hydrothermal alteration (Puffer and others, 1981). Similar effects might be expected within the present study area although dikes would not have had direct contact with seawater upon intrusion. Hydorthermal alteration by saline subsurface waters could have occured in addition to other hydrothermal alteration (during Permian thermal distrubance?) and deuteric alteration. It is possible then, that some of the alkaline altered aphyric and porphyritic dolerites low in SiO_2 and plotting near the discriminant on Figure 2 might actually be tholeiitic.

Dike Trends

The 103 dikes examined so far can be grouped according to their strikes into the following categories: NW (azimuths between 280° and 350°), N-S (azimuths between 350° and 10°), NE (azimuths between 10° and 80°), and E-W (azimuths between 260° and 280°). The distribution of dike types within these trend categories are summarized in Table 4. Of the 28 tholeiitic and alkaline dolerites (including porphyritic dikes), 11 (3 alkaline) trend NW, 9 NE, 6 (1 alkaline) E-W, and 2 N-S. Of the 25 altered tholeiitic and alkaline dolerites analyzed (including porphyritic dikes), 19 (10 alkaline) trend NW.

Ages of Dike Swarms and Sets

At least 3 dike swarms representing perhaps as many as 14 dike sets are present (Table 5). A preliminary estimation of dike ages presented in Table 5 is based primarily on field evidence and K-Ar (Whole-rock and biotite) age dates. It is expected that this age scheme will be modified and refined as work progresses.

Acknowledgements

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Table 1. Summary of Petrography of Mafic dikes of Eastern Massachusetts

Main Chemical Type[1]	Petro-graphic Type[2]	Mineral Assembages[3]						Main Trend[5]
		Groundmass					Plagioclase Phenocrysts[4]	
		Plag.	Aug.	Oliv.	Bio.	Amph.		
A?	OPC	abun.	abun.	rare	abs.	rare	comm. (< 3)	NE
T(21), A(2)	AP	ubiq.	ubiq.	rare	rare	rare	rare (< 5)	NW, NE (E-W, N-S)
T(4), A(1)	PP	ubiq.	ubiq.	rare	rare	rare	ubiq. (> 5)	NW (E-W)
T-A(1)	MED	abun.	abun.	rare	abun.	abs.	rare (< 10)	NE
T(9), A(7)	AD	ubiq.	ubiq.	rare	rare	rare	rare (< 5)	NW (E-W)
T(8), A(1)	PPAD	ubiq.	ubiq.	rare	rare	rare	ubiq. (> 5)	NW
A(2)	MCAD	ubiq.	ubiq.	abs.	ubiq.	rare	ubiq. (to 120)	NW
A(1)	CAMP[6]	abun.	abun.	rare	comm.	comm.	abun. (> 10)	NE
A(1)	HM[7]	abs.	comm.	comm.	abs.	abun.	abs.	NE

1 Numbers = number of dikes; T = Tholeitic, A = Alkaline. OPC dike not yet analyzed.

2 AP = aphyric, PP = plagioclase-phyric, Med = Medford dolerite, AD = altered dolerite, MC = megacryst-rich, CAMP = camptonite, HM = hyalo-monchiquite, OPC = olivine-phyric camptonite

3 Ubiq. = ubiquitous, abun. = abundant, comm. = common, abs. = absent.

4 Size limits in mm in parentheses, abundant olivine phenocrysts (< 5mm) in type OPC.

5 Includes dikes that have not been chemically analyzed; lesser trends in parentheses.

6 One dike only; phenocrysts and megacrysts of andesine, biotite kaersutite and xenocrysts of microcline perthite and wide variety of crustal and mantle xenoliths are abundant.

7 One dike only; other major phases are glass and ocelli filled with analcime, zeolite, and/or calcite.

Table 2. Representative modal analyses of mafic dikes, Northeastern Massachusetts[1]

Mineral	Type of Dike[2]						
	APa	APt	PPADa	PPADt	MEDa-t	MCADa	HMa
Total plagioclase	63.2	50.9	45.1	64.5	68.7	76.1	-
phenocrysts	trace	trace	4.6	trace	-	53.4	-
Total Augite	25.3	40.6	40.8	27.1	18.5	12.9	4.2
Uralitized	3.9	4.4	36.8	23.4	1.8	11.2	-
Olivine[3]	-	trace	3.0	trace	trace[4]	-	2.4
Biotite	-	0.2	-	-	6.2	7.3	-
Kaersutite	-	-	-	-	-	-	32.0
Opaque Accessories	8.9	7.3	4.6	8.3	2.6	3.1	13.4
Nonopaque Accessories	2.6	1.0	0.5	0.1	4.0	0.6	-
Ocelli	-	-	-	-	-	-	7.4
Glass/mesostasis[5]	-	-	6.0	-	-	-	40.6

1 Based on 1000 points per thin section.
2 See footnote 2 on Table 1 for key to dike type symbols; a = alkaline, t = tholeiitic
3 Usually replaced by chlorite, smectite, carbonate, iddingsite or combinations of these. Commonly fresh when present in Cape Ann dikes.
4 Totally altered in dike interior; about 10% fresh and altered microphenocrysts in chilled margins.
5 Mesostasis of very fine-grained altered plagioclase, pyroxenes in PPAD; glass plus tiny microlites of mostly kaersutite in HM.

Table 3. Mean chemistry of eastern Massachusetts mafic dikes.*

OXIDE	THOLEIITIC** DOLERITES AP	THOLEIITIC** DOLERITES PP	THOLEIITIC** ALTERED DOL. AP	THOLEIITIC** ALTERED DOL. PP	ALKALINE** DOLERITES AP	ALKALINE** DOLERITES PP	ALKALINE** ALTERED DOL. AP	ALKALINE** ALTERED DOL. PP	MED-FORD DOL.	PLAG-MEGA-CRYST RICH ALT. DOL.	CAMP-TONITE	HYALO-MONCH-IQUITE
SiO_2	40.21	51.26	50.10	49.58	46.25	52.78	48.02	47.81	49.71	51.35	50.36	49.71
Al_2O_3	15.90	16.80	16.56	16.64	15.20	15.96	15.83	16.31	15.78	16.52	17.46	17.70
TiO_2	2.62	2.26	2.54	2.55	3.74	2.30	3.55	3.25	3.17	2.16	2.01	2.56
Fe_2O_3	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
FeO	10.97	8.95	11.04	10.66	12.70	11.39	12.62	11.40	10.96	10.71	11.01	11.80
MnO	0.23	0.18	0.23	0.23	0.21	0.43	0.23	0.19	0.22	0.28	0.16	0.21
CaO	9.49	8.36	7.65	8.87	8.78	5.08	6.84	8.70	8.64	6.43	5.59	4.47
MgO	6.36	5.47	5.95	5.83	6.14	2.83	5.40	5.75	4.57	2.90	4.77	3.43
K_2O	0.88	1.56	0.82	1.04	1.32	2.75	1.70	1.41	1.24	2.30	2.60	2.14
Na_2O	1.88	2.73	2.61	2.10	2.93	3.31	3.10	2.71	3.14	3.76	3.18	5.12
P_2O_5	0.46	0.43	0.50	0.50	0.73	1.17	0.71	0.47	0.60	1.19	0.86	0.86
Number of Analyses	23	4	9	7	2	2	8	2	9	5	2	1

* XRF analyses presented inanhydrous from normalized ot 100 wt.%.
Fe_2O_3 assumed 2.00% (Hooper, 1981). Analyses done with a Philips P.W. 1410 XRF spectrometer on glass beads using the method of Hooper and Atkins (1969) as modified by Hooper and others (1976).

** AP = aphyric, PP = plagioclase-phyric.

Table 4 Summary of dike trends

Dike Type	Frequency of dike trends[1]			
	NW	NE	E-W	N-S
All types[2]	51	21	21	6
Dolerites	16	5	12	3
Plagioclase-phyric	4	2	2	-
Altered Dolerites	17	6	5	1
Plagioclase-phyric	12	4	2	1
Medford Dolerite	-	1	-	-
Plagioclase Megacryst-rich altered dolerite	2	-	-	-
Olivine-phyric camptonite	-	1	-	-
Camptonite	-	1	-	-
Hyalo-monchiquite	-	1	-	-

1 See text for azimuth limits of each category.
2 If unclassified dikes included: 71NW, 32NE, 30E-W, 9N-S.

Table 5. Summary of dike swarms, sets and ages for northeastern Massachusetts.

Swarm	Set	Estimated Age[1]
NW	tholeiitic dolerite	Mz?
	alkaline dolerite	Mz?
	tholeiitic altered dolerite	Pre-Mz?
	alkaline altered dolerite	Pre-Mz?
	plagioclase megacryst-rich altered dolerite	Pre-Mz?
NE	tholeiitic dolerite	Mz
	tholeiitic altered dolerite	Mz
	Medford dolerite	Mz
	lamprophyres	Mz
E-W	tholeiitic dolerite	Pre-Mz?
	alkaline dolerite	Pre-Mz?
	tholeiitic altered dolerite	Pre-Mz?
	N-S tholeiitic dolerite	Mz?
	N-S tholeiitic altered dolerite	Mz?

1 Mz = Mesozoic

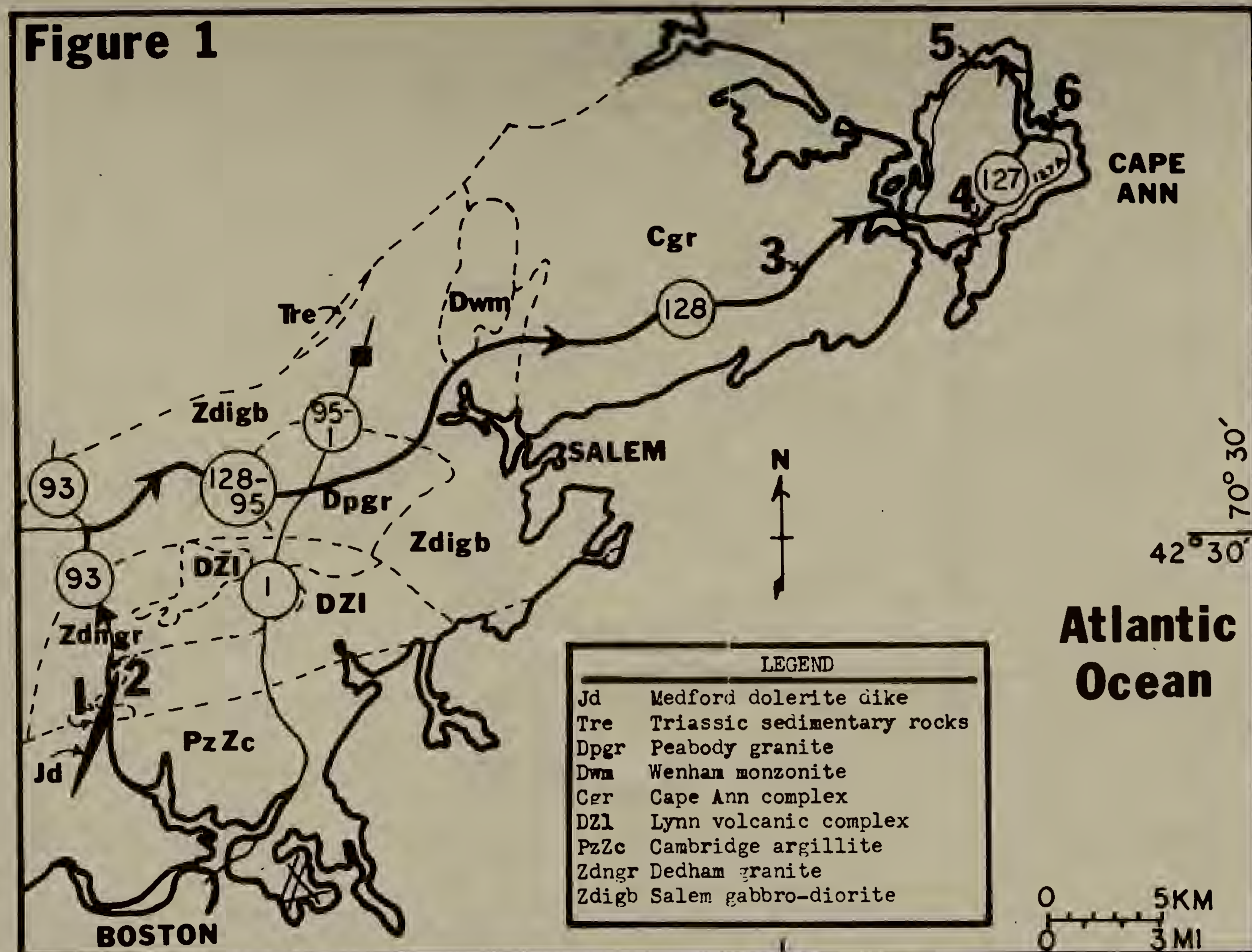

Figure 1. Location map showing field trip route (highway numbers circled) with stops 1-6 numbered. Geology (dashed contacts) from Bedrock geologic map of Massachusetts (E-an Zen, 1983).

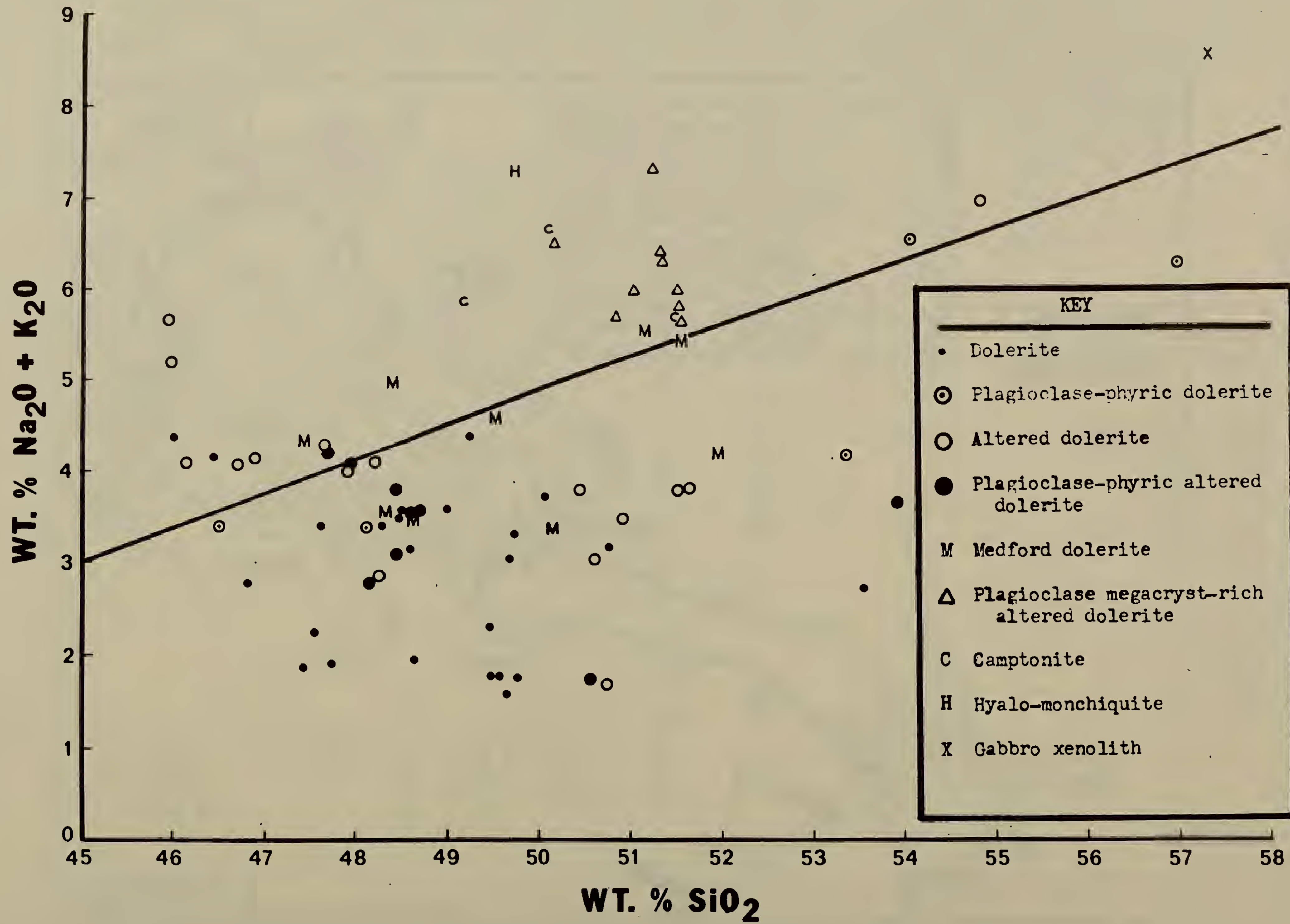

Figure 2. Alkali-silica diagram showing 56 dikes from eastern Massachusetts.

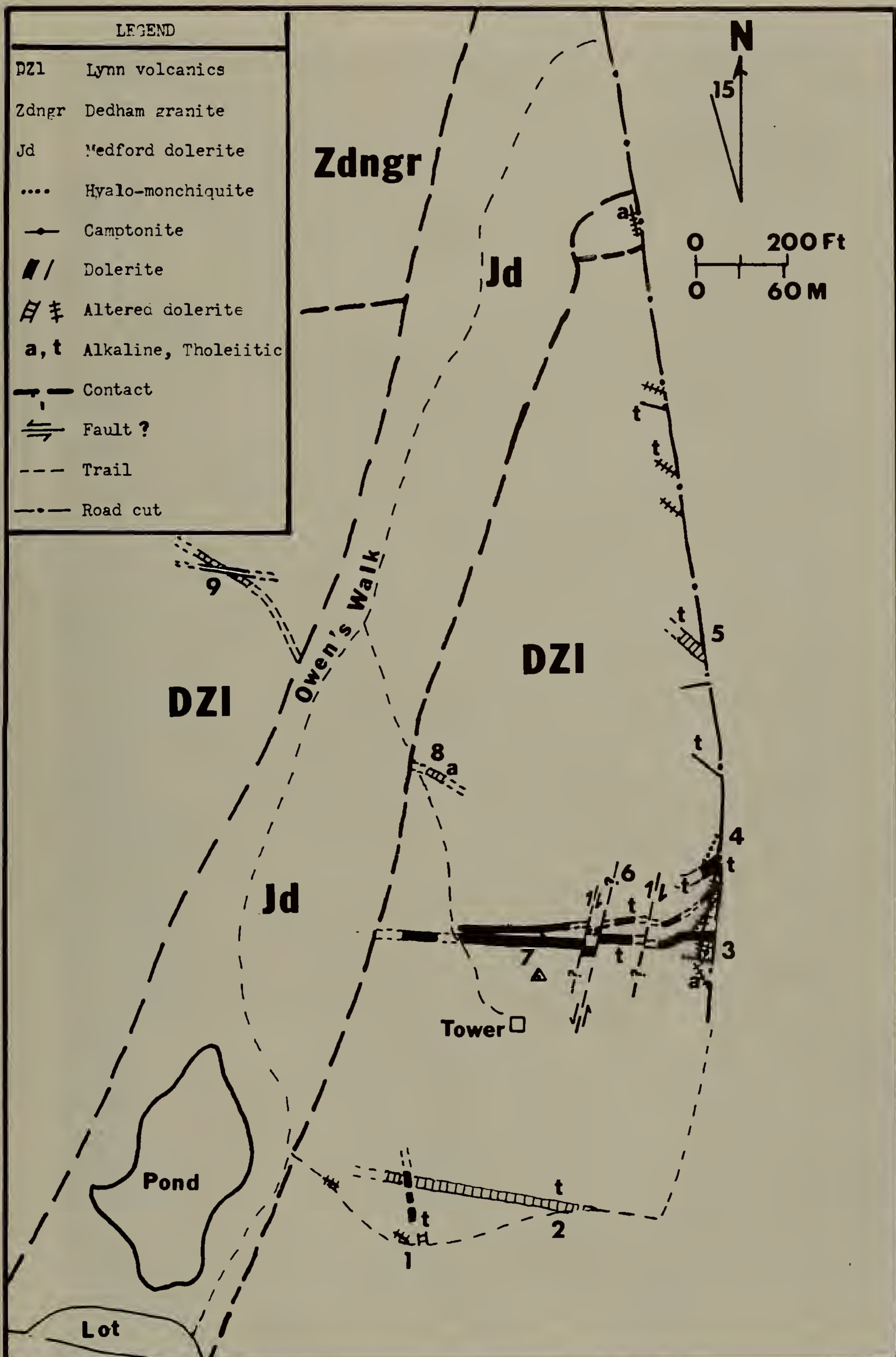

Figure 3. Geologic map of Pine Hill area after Wilson, 1901; Ross 1981; and Ross, unpublished mapping. Only dikes chemically analyzed and/or examined in thin section are shown.

Figure 4. Photograph of south end of road cut along Route I-93 at Pine Hill (Stations 3 to 4 in Figure 3). North is to the right. Dikes 2 and 4 are also seen at Stations 6 and 7 on Figure 3. See text for discussion.

Road Log

The assembly point is in the Bellevue Pond parking lot at Pine Hill just west of Interstate Route 93 in Medford, Massachusetts. Proceed south from the Inn at Danvers via Rte. 95 (merges with Rte. 1 to the south) 3.4 miles to the Rtes. 95/128/1 interchange. Exit right just past light for 95 south and Massachusetts Turnpike (90), Providence R.I., and N.Y. Turn right in 0.5 miles at sign for Burlington, Mass. Pike. Proceed 8.2 miles to exit for Rte. 93 south and Boston. Proceed south on Rte. 93 5.1 miles and take exit 7 for Rte. 28 Felsway West and Winchester. Proceed 0.4 miles and take first sharp right in rotary to South Border Road, Winchester. 0.2 miles to Pine Hill parking lot on right at white pipe gate. Park in lot or in front of gate if closed.

Mileage

0.0 Stop 1. Subdued outcrops of the Medford biotite-bearing dolerite dike occur in the grassy area between South Border Road and the parking lot. You are standing in the approximate center of the dike which trends N15°E through Bellevue Pond to the north of the parking lot (Figure 3). Walk north about 300 feet on the trail along the east (right) side of Bellevue Pond and go right at its junction with a dirt road.

Several dikes cutting Lynn volcanics (Pine flow unit of Zarrow, 1978) crop out along the north side of the road. Three dikes are exposed at the corner (Sta. 1, Figure 3). What appears to be an en echelon offset of a NW-trending dike is actually 2 dikes. The 3 segments to the west and extending north up the hill are a single tholeiitic dolerite. The segment to the east (right) and nearest the road is a tholeiitic altered dolerite and is finer-grained than the other dike. To the west of both is an obscure outcrop at road level of a tholeiitic, plagioclase-phyric altered dolerite that may be the extension of a similar dike seen to the NW in the road (Figure 3). A well-exposed tholeiitic altered dolerite occurs in the next curve (Sta. 2) and trends (Az 277°) across the hill until it is eventually cut by the tholeiitic dolerite seen at Station 1. Continue along the trail which turns left (north) as it nears Rt. 93. Follow along the right side of the cyclone fence north to the large road cut along the exit ramp and Rte 93. Numerous dikes will be examined along this road cut. The most noteworthy are the dikes from Station 3 to Station 4 (Figures 3 and 4), where cross-cuting relationships can be examined. Dike 5, (Figure 4) crosses the roadcut at low angle to produce broad surface exposures. It cuts dikes 2, 3, 4, and 6. The NE-trending olivine-phyric camptonite (left of Figure 4) cuts 2 thin E-W trending dikes (fresh and altered dolerite). Both these younger dikes trend slightly NE and the dikes they cut all trend within 10° of due east-west.

The hyalo-monchiquite (dike 7, Figure 4; Station 4, Figure 3) is distinctly blacker than the other dikes. See Table 2 for a modal analysis, and see Table 3 and Figure 2 for chemical analysis. It is the only glass-bearing dike I have observed in the region so far. Kaersutite is the principal ferromagnesian mineral.

Continue north along the road cut to where exposure ends beyond Station 5 (Figure 3). Climb the gentle slope back south-west to hole in cyclone fence. Go through hole in fence and continue southwest up narrow valley to Station 6.

The minor left-lateral displacements of dikes along a possible NE-trending fault (Billings, 1925; Koch, 1978) will be examined. Right-lateral offsets are also evident along possible NE-trending faults to the east and west of, and nearly parallel to, that mapped by Billings (1925) and Koch (1978). Extreme caution should be used when attributing apparent strike-slip offsets of dikes to faulting in cases where faults are not clearly visible in outcrop. I have seen many well-exposed examples of en echelon dike offsets in this region and on the Columbia River Plateau in which absolutely no faulting has occured. In such cases the offsets are due to magma intruding into joints or faults which are en echelon or offset to begin with and along which no movement has occured subsequent to dike intrusion. An excellent example of a dike showing lateral displacement due to flow along a joint can be seen in exposures along the Saugus River on Rte. 1 in Medford (Stop 3 in Ross, 1981b). This can also be seen in thin dikelets at Station 3 (Figure 3).

Follow the dikes at Station 6 west uphill to outcrops north of the granite observation tower at Station 7 on top of Pine Hill. The white paint stripes on the southern dike show true north. A compass needle held within about one foot of the ground will be deflected as much as 180°. This is apparently due to a history of lightning strikes at this locality. Note the xenolith of Lynn volcanics near the south margin of this dike.

The dike immediately north of the painted dike is dipping 68° N and its thickness at this exposure is apparent. Its true thickness (1.9 meters) can be seen in the ledge a few meters east. The apparent curvature of this dike on Figure 3 is produced by the erosional shift down the dip of the dike where it enters the narrow valley at Station 6. This curvature was not shown on previous maps (Billings, 1925; Koch, 1978) but was depicted on Wilson's map (1901). This effect is important to remember, especially when projecting a dipping dike across a valley in which exposures are poor or absent. This is precisely the situation between Stations 8 and 9 (Figure 3).

Enter gravel road just north of tower and proceed down hill to Station 8 (Figure 3). The NE-trending valley is underlain by the NE-trending Medford dolerite dike in which extensive nineteenth century quarrying occured at Pine Hill.

A NW-trending altered, alkaline dolerite is exposed just above the road and just uphill of where the Medford dike crosses the road. This dike dips 72°S here and at the same elevation across the valley (Station 9) where it also dips south. Mapping by Billings (1925) and Koch (1978) show these two seqments aligned perfectly along strike. They also project it into the Medford from either side without showing any erosionally produced "curvature" of the dipping dike where it crosses the valley. Wilson (1901) and my preliminary mapping (Figure 3) reflect an erosional shift. If the dike seqments at the same elevation (i.e. erosional level) across the valley are projected along strike they do not meet but rather show a left-lateral displacement of over 30 meters on both Figure 3 and Wilson's map (1901). This offset could have been produced by dilation of the Medford dike's conduit upon intrusion or by subsequent faulting. If the dilation is removed, the cross-dike seqments more closely line up but still are offset. Dilation oblique to the trend of the Medford dike also could account for this. This contradicts Billing's (1925) conclusion that the Medford dike's emplacement was by "large-block stoping" and not dilation of country rock. Additional detailed mapping is underway to reconcile differences between Wilson's (1901), Billing's (1925) and Koch's (1978) mapping. At the present, I believe dilation of the country rock during intrusion of the Medford magma was the principal intrusive mechanism.

A short distance down hill the NE-trending east contact of the Medford dolerite with the Lynn volcanics crosses the road. The actual contact cannot be readily observed here, even upon digging through soil and regolith cover (below the road). Samples from within a few centimeters have been collected just below the road but the absolute margin is weathered away. This contact can by observed (requires digging) about 250 meters to the north. Best exposures are in roadcuts on routes 93 and 28 to be observed at Stop 2 and the discussion of the contacts will be delayed until then.

The Medford dike forms subdued rounded exposures and typically exhibits spheroidal weathering. The fresh rock is a suprisingly tough cookie to break with a hammer. Its most distinctive field characteristic is its classic "diabasic" (i.e. ophitic), medium-grained texture accentuated by the color contrast of the light grayish-green labradorite (An_{70} cores, An_{29}rims, Ross, 1981b) with the darker mafic constituents (Table 2).

I have collected and analyzed (chemically and petrographically) 8 samples along a traverse (at the present station) extending from very near the east margin to the center of the Medford dike. It ranges from tholeiitic to alkaline in chemical composition (Figure 2, Table 3) indicating the magma probably underwent substantially more differentiation than typical of thinner dolerite dikes. Additional analyses are underway from this locality and at Stop 2 where the absolute chilled margin of both contacts have been sampled.

Follow the road downhill to its junction with Owen's walk and cross Owen's walk and climb up the hill to Station 9 (Figure 3) to observe the western segment of the alkaline altered dolerite at Station 8. Time permitting we will walk north along the ridge to observe several dikes including Billings'(1925) epidosite dike. In either case, we will return to the parking lot via Owen's walk.

0.0 Retrace route east from parking lot on South Border Road.

0.2 Enter rotary and cross over I-93 and exit onto I-93 northbound for Lawrence and Salem, N.H.

1.0 Pull well off to right onto shoulder, park and lock vehicles.

Stop 2 Contacts of Medford dolerite dike. Both contacts of the Medford dolerite dike can be examined in roadcuts here on Rtes. I-93 and 28 (just over fence). The dike has narrowed from about 140 meters at Pine Hill (Wilson, 1901) to about 14 meters at this locality. The east contact is exposed in the road cut about 500 feet south of the north contact. This is an apparent thickness (as we will see later at Rt. 28) since the east contact is at a low angle to the highway here. As described by Billings (1925) from localities at Pine Hill, the east contact is not sharp. It appears gradational and fused with the Dedham granodiorite. Pink potassium feldspar xenocrysts of Dedham can be seen within the Medford dike margin. This contact is sharp, however, in the road cut on Rt. 28 to be seen shortly. In thin section, the east dike margin is microporphyritic with microphenocrysts of pyroxene, plagioclase, and altered olivine psuedomorphed by bowlingite (as in the dike interior).

Walk north to the west contact of the dike (strikes Az 15°, dips 81°E). Note it is much sharper and, at this locality, faintly banded and slightly vesicular. The rock is microporphyritic in thin section and, unlike at the east contact, the olivine is fresh. The presence of olivine phenocrysts in the chilled margin with plagioclase, augite, and biotite confined to the groundmass, indicated olivine began crystallizing prior to the other phases. Additional chemical and petrographic work is underway to detect and document possible chemical and mineral compositional changes across the dike. At the west contact the dike branches around a Dedham granodiorite xenolith, showing that stoping on a small scale did occur, as is typical of dikes in the region. This would be expected as a fissure dilated upon magma intrusion into brittle rock.

Climb to the top of the road cut and follow cyclone fence a short distance north and climb through large gap at base of fence.

Both contacts can be seen in roadcuts on east (east contact) and west (west contact) sides of Rte 28. The dike is about 14.6m (48 feet) thick here and both contacts are sharp. The dike apparently terminates a short distance to the north, perhaps within Spot Pond (Wilson, 1901; Billings, 1925; Kaye, 1980).

Return to vehicles and proceed north on I-93.

2.1 A 35 m thick tholeiitic dolerite trending 273° is exposed on either side of I-93.

5.4 Take exit 11N, Rte 95N & 128 for Peabody (and Gloucester) and enter onto Rte 95/128. Switch to center lane as early as safe. Stay on 95N & 128.

21.3 Just north of the entrance from the service area onto 128 a 1.5 m thick, NE-trending tholeiitic dolerite is well-exposed. Its NW margin forms a vertical face in the nearly parallel road cut. The hackly-looking joint pattern is actually a cross-sectional view of the base of fairly distinct (for dikes in region), short columnar joints formed perpendicular to the dike contact.

25.2 Stop 3 (optional). Pull off on right shoulder, park, and lock vehicles. Very, very carefully walk across northbound lanes of Rte. 128, cross median strip and walk along eastern (median-side) roadcut of south bound lanes. Five or 6 dikes (one may branch) are exposed here and are, in order from north to south, as follows: alkaline altered (moderately) dolerite, alkaline dolerite, alkaline dolerite, alkaline altered dolerite, and (in brush) an unusual coarsely plagioclase-phyric alkaline altered (moderate) dolerite. This latter dike is identical to the dike to be seen about 12 km to the northeast at Stop 6. It is not on strike with 5.6 meter thick dike at Stop 6, it is thinner (1 meter), strikes 323° and dips 88° NE. A single chemical analysis from this locality agrees well with analyses from Stop 6. These dikes almost certainly represent a single intrusive episode of magmas derived from the same source.

Return to vehicles (watch traffic) and continue north on 128.

29.9 A thin, NE-trending (dips 52° NW), plagioclase-phyric tholeiitic dolerite is visible across 128 near the northern end of the roadcut of the south-bound lanes. This one is essentially identical to one to be seen at the next stop (Stop 4).

32.7 Enter rotary and take second exit staying on 128 N for Rockport.

33.5 Enter second busier rotary and take second exit, staying on 128 N for Rockport.

34.2 Pull of on right shoulder at base of hill just uphill from concrete retaining wall.

Stop 4, Mafic dikes and xenoliths in Cape Ann granite. A medium-grained, NW-trending, alkaline dolerite is exposed just uphill from the retaining wall on the SW (right) side of highway. It is texturally ("diabasic"), petrographically (biotite-bearing), and chemically (but with lower SiO_2) reminiscent of the Medford dike.

A few meters uphill two thin purplish felsite (quartz, microcline, amphibole, plagioclase) dikes (trend NE and dip steeply NW) are exposed. They are cut along strike (higher in roadcut) by a thin NW-trending alkaline altered dolerite dike.

Across the highway a thin NE-trending (dips 60° NW) tholeiitic, plagioclase-phyric dolerite cuts Cape Ann granite which here contains abundant xenoliths of a coarse-grained labradorite-phyric gabbro. A number of other similar zones are present elsewhere on Cape Ann (Dennen, 1976). These xenoliths indicate the presence at depth of a gabbroic pluton presumably intruded by Cape Ann granite.

Return to vehicles and continue SE to stop light.

34.3 Turn left onto Rt 127 at stop light and proceed toward Rockport.

35.6 Rockport city limits.

37.4 Turn left at confusing intersection (known affectionately as "5-corners") and remain on Rte. 127 headed for Pigeon Cove.

38.4 Cross over bridge with view left of large granite quarry.

40.8 Turn right and enter Seaside and Locust Grove cemeteries. Follow main road straight to back (NW), lower end of cemetery, turn right (no choice) and park along right side of road, bring lunch, and lock vehicles. Walk NW across road (Figure 6), and into woods. An unmarked trail can be found and followed downhill through the woods to the ocean front. Otherwise just blunder NW through the woods until you reach the ocean where we will assemble for lunch and Stop 5.

Stop 5. Dikes at Seaside-Locust Grove cemetery. Dikes present at this locality include the following (numbers refer to stations on Figure 5): dolerite (1,2, and 5), altered dolerite (4) and plagioclase-phyric altered dolerite (3). Chemical analyses of these are in progress and will be available by the time of this field trip.

Of special interest at this locality is the one meter right-lateral offset of dike 5 by a N17°E trending mylonite about 3 meters thick and dipping 77° NW (Figure 5).

The remains of what appears to be a mafic or ultramafic xenolith (someone beat me to it) occurs in dike 3.

Figure 5. Stop 5 at Seaside cemetery. Dikes referred to in text are numbered. 1 inch = 1000 feet (305 m).

Return through woods (use trail this time) to cars.

41.3 Turn left onto Rte 127 at exit from cemeteries and retrace route back on 127 to "5-corners".

44.7 Turn left at "5-corners" onto Broadway Street and continue NE on Broadway.

45.1 Turn right at T-intersection with 127A (Mt.Pleasant Street).

45.2 Turn left onto Norwood Avenue (one-way).

45.4 Turn left at stop sign onto Highland Avenue (no street sign).

45.5 Pull off right onto wide, rocky shoulder, park and lock vehicles. Take asphalt path beyond vehicles through woods to the "Headlands". Follow vegetation line along upper edge of ocean-front exposures about 60 meters until coarsely plagioclase-phyric dike encountered.

Stop 6. Plagioclase megacryst-rich alkaline altered dolerite. This dike is unusal in the region due to the large (up to 120 mm) and abundant (up to 71.6 volume percent) plagioclase megacrysts (longer than 3 cm) and phenocrysts. The dike extends at least 2.3 km north and can be seen exposed along the shore across Sandy Bay. It is also visible at low tide just across the harbor on both sides of the turn-around at the end of Bearskin Neck. A thinner, identical dike occurs at Stop 3 and probably represents the same intrusive episode.

Here at "The Headlands" the dike is 5.6 meters thick strikes N7^{o} to 14^{o}W, and dips 88^{o}E (across Sandy Bay it strikes N27^{o}W, dips 74^{o}E, and is about 8.2 meters thick). If the tide is low enough, a thin, tholeiitic plagioclase-phyric dolerite dike (strikes N10^{o}E, dips 22^{o}SE) can be seen cutting the megacryst-rich dike. The broad flat surfaces developed in the granite a few meters west of this locality is due to wave erosion along this thin dike. It is also exposed a few meters to the east.

The average of 6 whole-rock X.R.F. analyses of the megacryst-rich dike is listed in Table 3 and all 6 analyses (plus one from Stop 3) are plotted on Figure 2. A modal analysis from the center of the dike is listed in Table 2 and its petrography is summarized in Table 1.

This dike affords in excellent opportunity to observe the effects of magma flow within a narrow conduit on the distribution and orientation of entrained phenocrysts. Wave erosion has created a prominant chasm by removing the western portion of the dike from its contact, inward to a prominant joint face that is approximately parallel

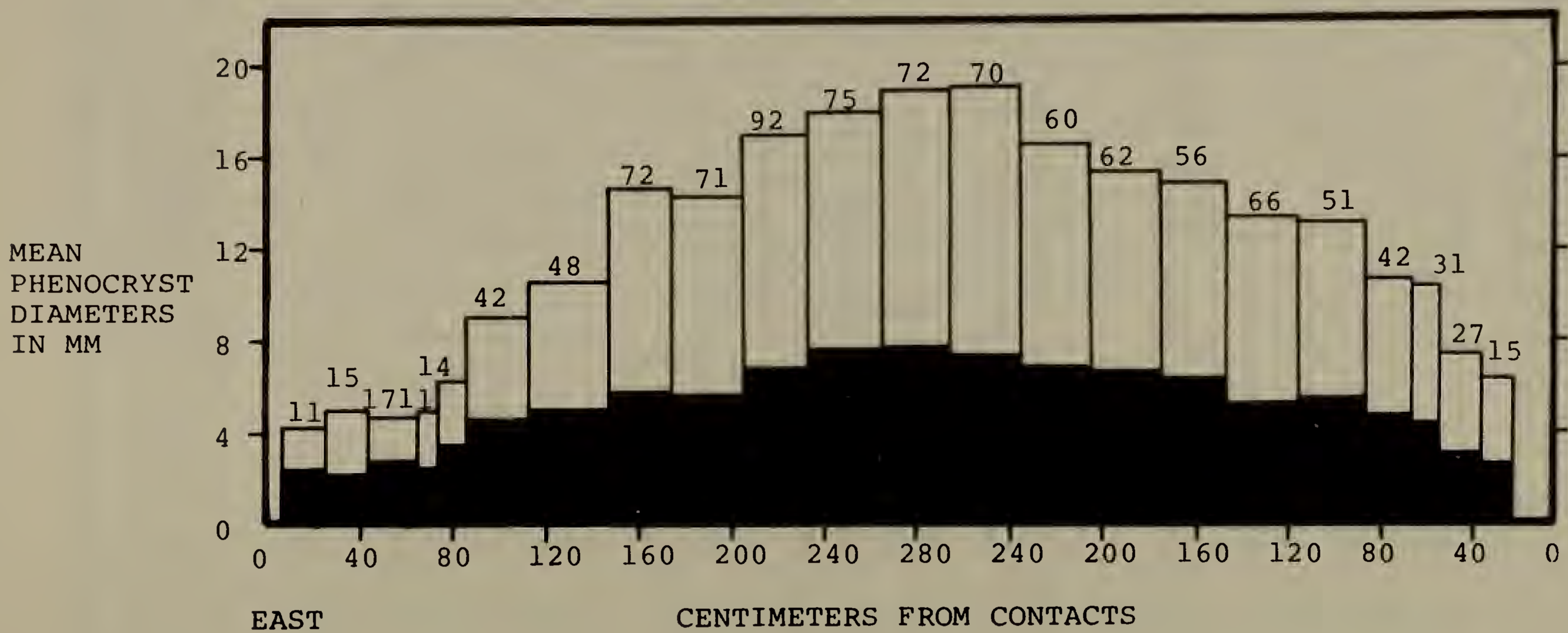

Figure 6. Histogram of average pheonocryst dimensions across the megacryst-rich dike at Stop 6. Numbers at top of bars are largest grain lengths (mm) measured. Dark bars are average megacryst widths.

the trend of the dike. As a result, a three-dimensional analysis of variations in the size, orientation and modal abundance of plagioclase megacrysts (i.e. phenocrysts longer than 3 cm) and phenocrysts was possible. The results summarized below are excerpted from Ross (1984).

Measurements were made within 600 cm^2 areas equally spaced across the dike. Means of the factors measured were calculated for each 600 cm^2 area and compared. The following dike margin-to-center trends are present: volume percent phenocrysts increases from zero (within 6cm of contact) to 71.6 percent; phenocryst cross-sectional dimensions increase from 4.1 X 2.2mm to 19.2 X 7.9mm (maximum length = 120mm) representing an increase in length/width ratio from 1.86 to 2.68 (Figure 6); phenocryst strike deviations from dike strike increase from 19.1^o to 40.9^o; phenocryst dip angle deviations from dike dip increase from 20.9^o to 39.7^o.

Mean phenocryst abundance, size, elongation (in cross section), and angular deviation from dike attitude all increase inward from the margins of the dike. This data clearly illustrates the greater concentration of increasingly larger phenocrysts towards the center of the dike by flowage differentiation. Magmatic flow alignment of phenocrysts was more effective and more nearly parallel to the dike's attitude near its margins where shear due to flow along the conduit walls was greater.

SiO_2, K_2O, FeO, MnO, and perhaps TiO_2 and MgO decrease toward the center of the dike. Al_2O_3. P_2O_5, CaO, and perhaps Na_2O increase inward sympathetically with plagioclase concentration, demonstrating the effect of flowage differentiation on bulk chemical variations across the dike.

End of trip. Retrace route south to Route 128 for connections north and south.

A GEOLOGIC TRAVERSE ACROSS THE NASHOBA BLOCK, EASTERN MASSACHUSETTS

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Introduction

The easternmost margin of the Appalachian-Caledonian orogenic system in southeastern New England is divided into three distinct tectonic blocks or terranes separated by the Bloody Bluff and Clinton-Newbury fault systems (Figure 1). Both terranes east of the Clinton-Newbury fault zone, the Boston Platform and Nashoba Block) are believed to be "exotic" to North America prior to the mid- to Late Paleozoic (e.g., Williams and Hatcher, 1983; Zen, 1983a; Zartman and Naylor, 1984). But the rocks of the Nashoba Block, between these two major fault zones, contrast markedly with those of the Boston Platform to the east and the Merrimack Trough to the west. Contrasts with the Boston Platform include the degree of metamorphism, the type of sedimentation, and the age and composition of the igneous rocks. The large, Late Proterozoic granitic intrusions and volcanic outpourings of the Boston Platform are not present in the Nashoba Block, nor are the Ordovician to Devonian peralkaline plutons. In contrast, the Nashoba Block has been intruded by a series of intermediate plutons (Straw Hollow, Sharpners Pond, Assabet Quartz Diorite) and by deep-seated peraluminous granites (Andover) roughly contemporaneous with the peralkaline intrusions of the Boston Platform (Zartman and Naylor, 1984; Wones and Goldsmith, in press; Hepburn et al., in press).

A similar contrast occurs along the western edge of the Nashoba Block, where generally low-grade metasediments of the eastern portion of the Merrimack Trough of Lyons et al. (1982) abut against the Clinton-Newbury fault. These metasediments are part of a thick deepwater sequence (flysch and calcareous flysch) generally thought to have been metamorphosed and deformed during the Acadian Orogeny. The exact age of these sediments is still debated but has commonly been thought to be Siluro-Devonian by long distance correlations to fossil-bearing strata in central Maine. However, Lyons et al. (1982) have recently suggested they may be in part as old as Precambrian. The eastern Merrimack Trough has been intruded by a number of granitic plutons (Zen, 1983b), the Ayer being the most prominent in the field trip area (Figure 2). The Ayer has been recently dated as 433 +/- 5 m.y. (Zartman and Naylor, 1984), indicating emplacement prior to the Acadian Orogeny. Thus, it appears that the rocks west of the Clinton-Newbury fault zone have experienced the effects of the Acadian Orogeny, while to the east in the Nashoba Block these effects are not evident.

This excursion is designed to give a general overview of the Nashoba Block, with stops in the major metasedimentary, metavolcanic, and igneous units in the central part of the block. While much has been learned about this interesting terrane recently, a great deal more work is needed in order to answer even the most fundamental questions (i.e. age of sedimentation and

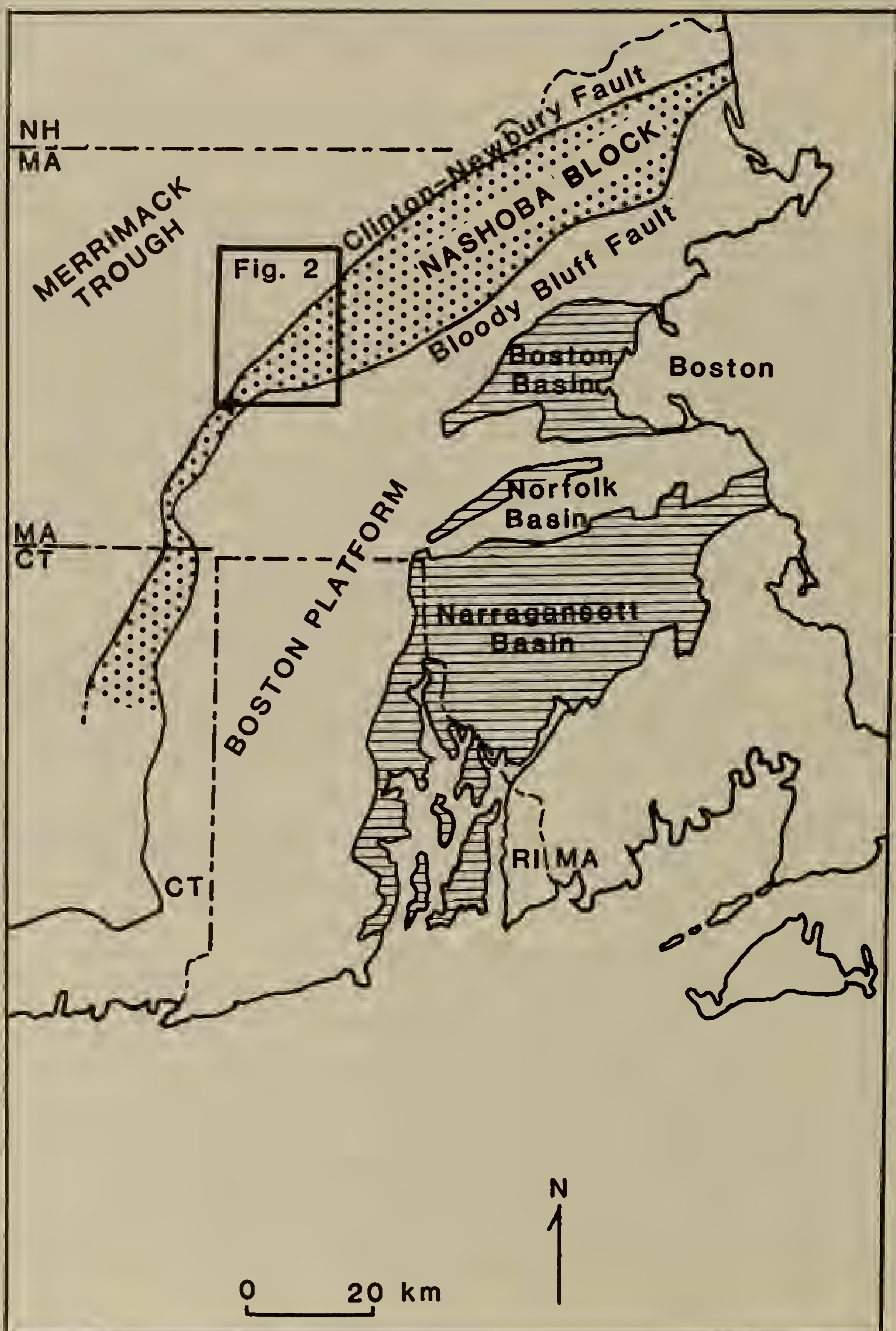

Figure 1. Location of the Nashoba Block, eastern Massachusetts. Sedimentary basins of the Boston Platform shown by lined pattern: Boston Basin, Late Proterozoic-Early Paleozoic; Narragansett and Norfolk Basins, Carboniferous. Outlined square shows area of Figure 2.

volcanism, age of metamorphism, correlations, etc.). Excursion participants may find we present more problems than answers.

Recent Work

However, interpretation of the Nashoba Block has been greatly aided by recent work, including mapping done in connection with the new Massachusetts Bedrock Geologic Map (Zen, 1983b). Complete, up-to-date references on this mapping are given in Zen (1983b) and Barosh et al. (1977) and are not repeated here. Bell and Alvord's (1976) work in stratigraphically subdividing the Nashoba Block was a major step forward. Other recent stratigraphic studies include those of Skehan and Abu-Moustafa (1976), Alvord et al. (1976), Goldsmith et al., (1982a, 1982b) and Goldsmith (in press). Radiogenic age determinations have recently been summarized by Zartman and Naylor (1984) and Olszewski (1980). Wones and Goldsmith (in press), Hepburn et al. (in press), and Hill et al. (1984) have provided recent discussions on the petrology and geochemistry of the igneous rocks of the Nashoba Block.

Nashoba Block

Stratigraphy

The stratified rocks of the Nashoba Block are largely high-grade (sillimanite and sillimanite - K-feldspar zones) eugeosynclinal metasediments and metavolcanics. Feldspar and biotite gneisses and amphibolites are dominant, along with lesser amounts of pelitic schist, calc-silicate granulite and marble. Bell and Alvord (1976) and Alvord et al. (1976) have subdivided the stratified rocks into five major formations, striking northeast, which they interpreted to be generally a westward dipping homocline, topping west. We feel that, because of the complex structural relations, faulting, and lack of fossil or other clear evidence of stratigraphic younging direction, the question of stratigraphic sequence remains unresolved. The stratified rocks are (from SE to NW): the Marlboro Fm., Shawsheen Gneiss, Fish Brook Gneiss, Nashoba Fm. and Tadmuck Brook Schist (Figure 2). Brief descriptions of these formations are given here. For more complete descriptions, refer to Bell and Alvord (1976) and the references cited in this section. On Figure 2, the designation used for the various units is largely that of Zen (1983b), so that field trip stops can be easily located on the new Massachusetts Bedrock Geologic Map.

Marlboro Formation. Predominantly amphibolites of the Sandy Pond Member. Biotite and hornblende gneisses lie between this member and the Bloody Bluff fault. Various rock types are mapped in the Marlboro between the Sandy Pond and the Assabet River fault to the northwest. These include rusty-weathering sillimanite schist and quartzofeldspathic granulite (see DiNitto et al., this volume).

Shawsheen Gneiss. (not seen on this field trip) Sillimanitic muscovite-biotite-plagioclase-quartz gneiss and minor amphibolite, not distinctly different from the Nashoba Fm. but separated from it by the Fish Brook Gneiss.

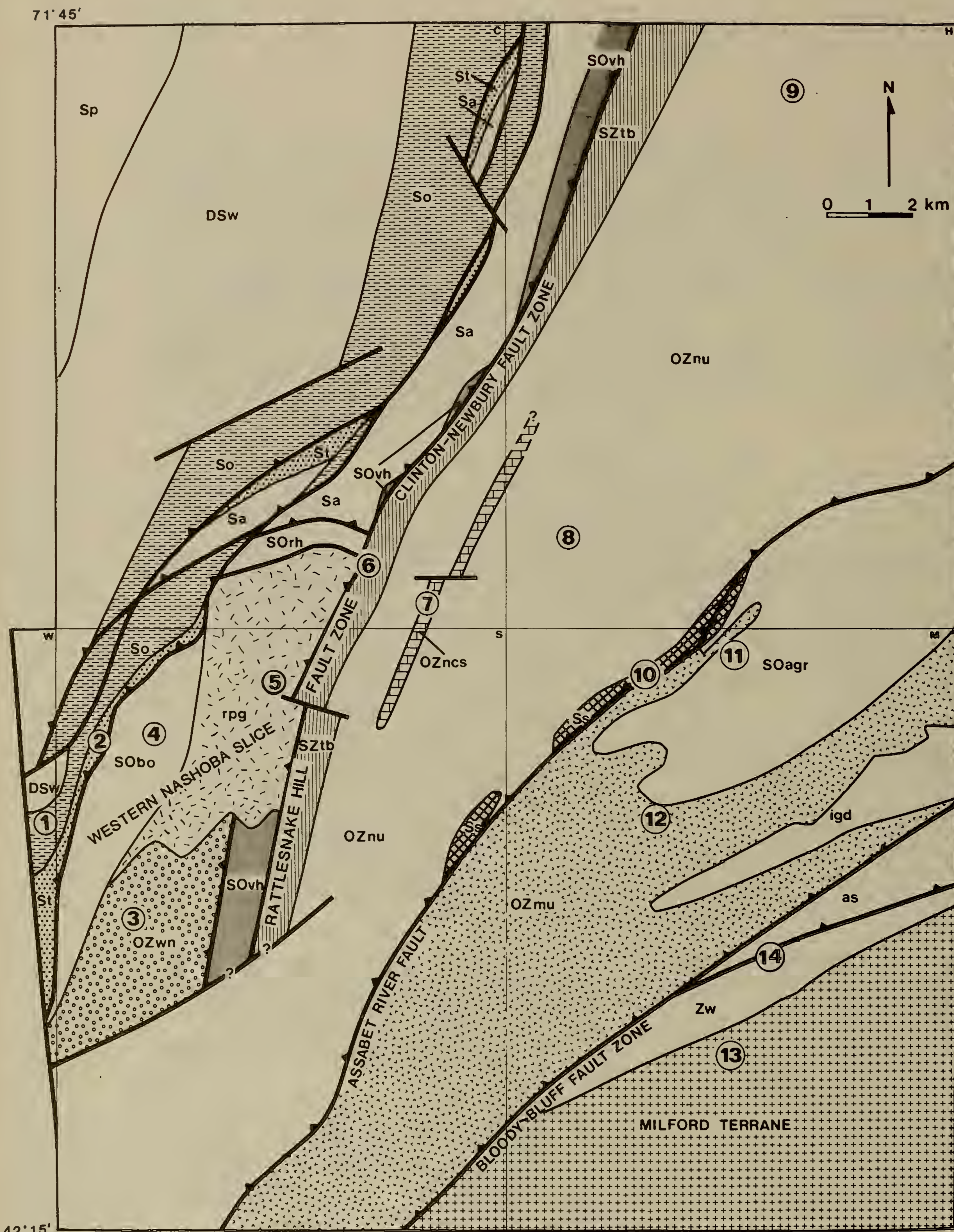
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FIGURE 2

GENERALIZED BEDROCK GEOLOGIC MAP ACROSS THE NASHOBA BLOCK

Clinton, Hudson, Worcester North, Shrewsbury & Marlborough quadrangles, east-central Massachusetts (after Bell and Alvord, 1976; Hepburn, 1978; DiNitto, 1983; Zen, 1983, & Munn; in progress)

MERRIMACK TROUGH

☐	Sa	Ayer Granite
☐	DSw	Worcester Formation
☐	Sp	Paxton Formation
▤	So	Oakdale Formation
▩	St	Tower Hill Quartzite
■	SOvh	Vaughn Hills Formation
☐	SOrh	Reubens Hill Formation

BOSTON PLATFORM

☐	as	Altered & sheared rocks
☐	Zw	Westboro Formation

NASHOBA BLOCK

▩	Ss	Straw Hollow Diorite
☐	SOagr	Andover Granite
☐	igd	Indian Head Granodiorite
▩	rpg	Rocky Pond Granite
▥	SZtb	Tadmuck Brook Schist
☐	OZnu	Nashoba Formation undifferentiated
▤	OZncs	Calc-silicate unit within the Nashoba Formation
▩	OZmu	Marlboro Formation undifferentiated
▩	OZwn	Nashoba Formation (western)

——	Contact
——	Fault
▲—	Thrust Fault
②	Field Trip Stop

QUADRANGLES

C	Clinton
H	Hudson
M	Marlborough
S	Shrewsbury
W	Worcester North

Fish Brook Gneiss. (not seen on this field trip) Generally light colored feldspathic and quartzofeldspathic gneisses.

Nashoba Fm. Bell and Alvord (1976) subdivided the Nashoba into ten members in the Billerica, Concord and Westford quadrangles. Many of these members have not been recognized in the field trip area, having been cut out by faulting or due to the lack of detailed mapping. The dominent lithologies of the Nashoba are quartz-plagioclase-biotite gneisses and schists. Varying percentages of these minerals produce much of the diversity within the formation. Lesser amounts of pelitic schist, rusty-weathering amphibolite, and calc-silicate granulite or marble are present in sufficient quantity to be locally differentiated. Migmatite and pegmatite are common, particularly in the northeast toward the Andover Granite.

Tadmuck Brook Schist. Rusty weathering, graphitic, micaceous phyllite to schist with porphyroblasts of andalusite and sillimanite locally present. In many localities, if not all, the Tadmuck Brook is a phyllonite.

Plutonic Rocks

The Nashoba Block has been intruded by a heterogeneous series of plutons ranging in composition from granite to diorite. The granites have generally been assigned to the Andover Granite. The Andover consists of a series of foliated and unfoliated peraluminous, garnet bearing, muscovite-biotite granites and pegmatites, thought to have been generated in part by anatexis of the Nashoba Formation. Zartman and Naylor (1984) date the Andover as ranging from 408 to 450 m.y. based on Rb-Sr whole rock analyses. Hill et al. (this volume), on the basis of additional Rb-Sr analyses, suggest that the Andover may represent two separate intrusions, one in the Late Silurian approximately 415 m.y. ago, and one in the Late Ordovician approximately 450 m.y. ago.

The Indian Head Hill Granodiorite (Hepburn and DiNitto, 1978) is a composite pluton, just north of the Bloody Bluff fault zone (Figure 2). The younger phase is an unfoliated biotite-muscovite granite. The older, a foliated hornblende-biotite diorite has been dated at 402 m.y. with Rb-Sr (Hill et al., this volume).

The principal dioritic rocks of the Nashoba Block (Sharpners Pond Diorite, Assabet Quartz Diorite and Straw Hollow Diorite) are generally unfoliated to weakly foliated hornblende-plagioclase-sphene bearing calc-alkaline intrusions. Zartman and Naylor (1984) give a zircon age for the Sharpners Pond of 430 +/- 5 m.y. that probably is representative for all of these intrusions.

"Western Nashoba slice"

From Newbury on the northeast to about Clinton, the Clinton-Newbury fault is a well defined fault zone which has rocks of the Tadmuck Brook Schist on its eastern side south of Lawrence. However, in the Clinton area, the fault splits into several branches (Figure 2; Zen, 1983b; Peck, 1976). Within this fault complex (southern Clinton and north-central Shrewsbury quadrangles) are an anomalous series of rocks which we call the "western Nashoba slice". This slice is bounded on the east by the Rattlesnake Hill Fault zone of Skehan and Abu-Moustafa (1976), on the west by faulting at the pronounced metamorphic break east of the Tower Hill Quartzite, and on the north by the faulted Reubens Hill Complex. The rocks in this slice are likely part of a separate thrust sheet (Munn, in progress) that includes the Boylston Schist (Grew, 1970; Goldsmith et al., 1982a) and the "western Nashoba Formation" (Nashoba, western on Figure 2). The Boylston Schist is a rusty-weathering garnet-silimanite schist seen at Stop 4. The "western Nashoba" is lithologically similar to the Nashoba Formation to the east, but has been separately designated in Figure 2 to emphasize its unique tectonic position (i.e., west of the Tadmuck Brook Schist). These formations are intruded by the Rocky Pond Granite, a fine to medium grained, white to light gray, two mica or muscovite granite that contains only a weak foliation (Stop 5). This granite was previously referred to as the Rattlesnake Hill muscovite granite by Skehan (1968) and the muscovite granite of West Berlin by Zartman and Naylor (1984). It has not yet yielded a reliable age (Zartman and Naylor, 1984). Also present in the area just west of the Rattlesnake Hill fault is a sliver of low-grade rocks, the Vaughn Hills Quartzite, quartzites and interbedded phyllites that have been correlated with the Tower Hill Quartzite and thus assigned to the Merrimack Group by Zen (1983b). We agree with this stratigraphic correlation but are as yet unsure of the tectonic implications (Munn, in progress).

Tadmuck Brook Schist

The Tadmuck Brook Schist occupies a unique position on the east side of the Clinton-Newbury fault zone. From about Littleton to Lawrence several members of the Nashoba Formation are truncated at low angle by the Tadmuck Brook (Alvord et al, 1976). Interestingly, where the Clinton-Newbury fault zone becomes less distinct north of Lawrence or where it breaks into a number of slivers, such as south of Clinton-Shrewsbury, the Tadmuck Brook is not present. Alvord et al. (1976) interpret the contact with the Nashoba as a disconformity, although they note that in places, the contact is known to be a fault.

In the area of the field trip (and to the northeast; Alvord, 1976; Bell and Alvord, 1976), the Tadmuck Brook Schist has a general tripartate division. To the west the rock is commonly a phyllite in appearance, with few recognizable porphyroblasts and common retrograde features. In its central part (as at Stop 6), it appears to be a phyllite to schist with common augen, knots and lenses containing andalusite and/or staurolite or their retrograded products. On its eastern boundary with the Nashoba it becomes a sillimanite-biotite schist.

Because of the presence of fine-grained pyrrhotite, the rock weathers rapidly and even "fresh" exposures are soon covered with a rusty rind that makes textures difficult to see in the field. Common throughout the formation are thin quartzose layers and lenses, and thicker lenses (to 10 m) of amphibolite that, as Bell and Alvord (1976) note, cannot be traced from one outcrop to the next.

Everywhere we have looked in the Tadmuck Brook Schist, there is evidence of cataclasis and recrystallization. Fluxion structure (Higgins, 1971) is common. The comminution and metamorphic recrystallization histories are complex and likely reflect several periods of repeated fault movement and at least two episodes of metamorphism (Munn, in progress). Large, remnant knots of coarse-grained andalusite and staurolite lie within a finer grained micaceous matrix that has been later overgrown by biotite, andalusite, and on the east by sillimanite. The youngest metamorphism produced the largely unoriented porphyroblasts seen at Stop 6.

Many questions about the metamorphism still remain. Is the younger metamorphism the result of thrusting cooler rocks into contact with the hotter Nashoba Block, or is it related to a later event, perhaps the same one that formed sillimanite in the Nashoba Block itself? Undoubtedly, late movements on the Clinton-Newbury have complicated the picture. Younger shear zones whose rocks show less recrystallized textures crosscut the Tadmuck Brook Schist.

What was the protolith of the Tadmuck Brook Schist? Clearly, the micaceous mineralogy and presence of aluminosilicates indicate an originally shaley protolith, but the coarse, multigranular textures in knot remnants indicate that its immediate precursor was not a fine-grained rock. The large amphibolite lenses suggest that at least some Nashoba material is present in the Tadmuck Brook. Numerous rusty weathering schistose units within the Nashoba might have served as the shaley protolith, but pelitic material is also common in the Merrimack Trough. At this stage it is impossible to identify the protolith for the Tadmuck Brook Schist and in fact, it may vary along strike. Certainly, it would not be unreasonable for the main movements along the fault, particularly if it was a thrust, to start as a bedding plane thrust in a rock of pelitic composition before transgressing other units. In any case, we feel there is sufficient evidence on every scale, from the regional to thin section, to assign a largely cataclastic origin to the present Tadmuck Brook Schist.

Summary--age, metamorphism and origin of the Nashoba Block

The age and metamorphism of the rocks in the Nashoba Block are not well constrained. No fossil-bearing rocks occur here, nor have any definitive correlations been made to well-dated rocks elsewhere. Olszewski (1980) dated zircons from the Fish Brook Gneiss (see Hill et al., this volume) at 730 +/- 26 m.y., which he interprets as the volcanic age for this unit. However, does this age for the Fish Brook Gneiss apply generally to the period of deposition for all the rocks in the block, or is the Fish Brook just an inlier of basement rock? DiNitto et al. (this volume), Hill et al. (1984), Hill et al. (this volume) and work still in progress attempt to date amphibolites from the Marlboro Fm. using Nd-Sm techniques. While a definitive age has not yet been established, model ages suggest the Marlboro is 450-550 m.y. old. Although older ages cannot be ruled out, this data suggests the Marlboro (and other metasediments?) are younger than the Fish Brook.

The metamorphism in the Nashoba Block has generally been assigned dates in the range of 430 to 450 m.y. (Olszewski, 1980; Zartman and Naylor, 1984), based on the Andover Granite and reset Rb-Sr ages. The pervasive metamorphism is that of the sillimanite or sillimanite - K-feldspar zones, and the peraluminous nature of the Andover suggests that it formed by anatexis of a sedimentary source. Spacially, the Andover is related to the Nashoba and migmatites generally increase within the Nashoba as the Andover is approached. Castle and Theodore (1972) indicate a P,T estimate for a pegmatitic phase of the Andover as 650°C. at 5 kb, close to our estimates of the regional metamorphism in parts of the Nashoba. Since the Andover has unfoliated varieties that have escaped most of the deformation, the date on the Andover is thought to put a maximum age on the metamorphism and regional deformation. Although Hill et al. (this volume) suggest there may in fact be two Andovers, and the relationship between the Andover and the Nashoba metamorphism has not been conclusively established, an age of 430-450 m.y. seems the current best estimate for the regional metamorphism. Interestingly, this age falls between what is normally thought of as Acadian and Taconic in the northern Appalachians.

Major faulting in eastern Massachusetts has been estimated to range from Precambrian to Alleghanian (see for example Castle et al., 1976; O'Hara and Gromet, 1983). The deformation within the Nashoba Block ranges from deep-seated ductile deformation to later, cross-cutting brittle faulting, as we shall see on the trip.

In what type of an environment did the Nashoba Block form? Geochemical evidence from amphibolites of the Marlboro Fm. indicate that they were originally mildly alkalic to high alumina basalts (Hepburn et al., in press; DiNitto et al., this volume) and contain trace element abundances consistent with formation at a convergent plate margin or in a marginal basin. The calc-alkaline dioritic intrusions in the Nashoba Block (Hepburn et al., in press; Hill et al., this volume) also indicate a convergent margin-type environment existing as late as 430 m.y. ago.

The sequence of events in the Nashoba Block is distinctive and very different from that in the Boston Platform to the east. Arguments recently summarized by Zartman and Naylor (1984) indicate that it is unlikely that these two blocks were joined together prior to the mid- to Late Paleozoic. The Late Precambrian (600-650 m.y.) granitic plutonism characteristic of the Boston Platform is not present in the Nashoba Block. At the time the Nashoba Block was undergoing high-grade metamorphism, calc-alkaline intermediate and peraluminous granite plutonism, the Boston Platform was experiencing only peralkaline plutonism. These different magmatic types occur in close proximity across the Bloody Bluff fault zone, but do not cross it. Thus it appears unlikely that these blocks could have been juxtaposed during the mid-Paleozoic plutonism. However, Permian K-Ar resetting of mineral ages (Zartman et al., 1970) is common to both, suggesting they were joined by the end of the Paleozoic (see also O'Hara and Gromet, 1983)

The time of juxtaposition of the Nashoba Block to the eastern Merrimack Trough is less well documented. Metamorphic isograds of apparently Acadian age are truncated along the Clinton-Newbury fault, indicating at least some post-Devonian movement.

The Nashoba Block is clearly an exotic block, with a geological history distinctly different from terranes surrounding it. As we come to better understand its rocks, it is likely that we will be in a better position to recognize its presence elsewhere in the Appalachians.

Acknowledgements

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A GEOLOGIC TRAVERSE ACROSS THE NASHOBA BLOCK, EASTERN MASSACHUSETTS

Road Log for Field Trip, Saturday, Oct. 13, 1984

Assemble at STOP 1 in Boylston, MA. at 9:00 A.M. BRING LUNCH. Stop 1 is approximately one hour from the Danvers NEIGC headquarters via I-95 South, Rt. 2 West, I-495 South to I-290 West. The trip will be in the following 7-1/2 minute quadrangles: Worchester North, Shrewsbury, Clinton, Hudson and Marlborough.

MILEAGE

Cum.	S/S	
		Exit from I-290 at Rt. 140. Exit 23 North toward West Boylston.
	0.0	Top of exit ramp, proceed north on Rt. 140. (Note gas stations here: Mobil and Gulf are the last we pass on the trip).
1.7	1.7	Junction with Rt. 70; continue north on 140 for 0.3 miles.
2.0	0.3	Park with caution along east side of the road by large road cuts. Meeting place. Stop 1.

STOP 1. Oakdale Formation. Typical light gray to purplish-brown weathering calcareous meta-siltstone and interlayered gray to gray-green phyllite of the eastern Merrimack Trough. Siltstone layers range in thickness from 1 to 20 cm and are separated by thin partings of micaceous phyllite or may be interlaminated with paper thin phyllitic partings on a scale of a few millimeters. Ankerite causes the characteristic purplish-brown weathering spots in the siltstones. The metamorphic grade here is biotite zone or lower. Note the small, tight to isoclinal folds in these outcrops. The prominent layering here is not bedding but a transposed bedding parallel to the axial surfaces of these isoclinal folds. Late kink bands with steeply dipping axial surfaces are conspicuous.

Continue north on Rt. 140 until U-turn can be made by Wachusett Reservoir. Reverse direction and proceed south on 140.

Cum.	S/S	
2.3	0.3	Return to junction with Rt. 70, turn left, and proceed northeast on 70.
3.3	1.0	Turn left at Boylston Town House. Proceed behind buildings and park. Walk up embankment to west. Stop 2.

STOP 2. Tower Hill Quartzite. Grew (1970) informally named this unit for exposures in Boylston, and it was formally adopted by Goldsmith et al. (1982a). Orthoquartzite is the most distinctive rock in the Tower Hill with sparse occurrences of

quartz-pebble conglomerate. Interbedded gray phyllites are found in the quartzite and as a mappable member between the quartzite and the Clinton-Newbury fault (or a branch of it), which passes not far below us here. The grade of metamorphism is that of the biotite zone or lower.

3.4 0.1 Return to Rt. 70. Turn right onto 70 West and retrace route back to junction with 140.

4.4 1.0 Turn left onto Rt. 140 South.

6.1 1.7 Turn left (east) onto Colonial Dr. and park. Walk across Rt. 140 to outcrops on the right side of the entrance ramp to I-290 West. Stop 3.

STOP 3. Nashoba Formation (western). The Nashoba Formation west of the Tadmuck Brook Schist is lithologically similar to the main Nashoba belt to the east, despite the fact that it is in a separate fault slice (Figure 2; Zen, 1983b). It is largely composed of various biotite gneisses and biotite schists metamorphosed to the upper amphibolite facies (sillimanite or sillimanite - K-feldspar zones). Muscovite is still present throughout the area, although much of it is commonly in large flakes and may be retrograde. Migmatites are locally present, although not to the extent seen in the northeastern part of the main outcrop belt. Folding is complex, with two or three generations of folds commonly present in an outcrop. Two or more periods of granitic or pegmatitic intrusions are also common.

An approximate log of features found in this heterogeneous series of outcrops as you progress from Rt. 140 along the north side of the entrance ramp includes: (1) a late silicified shear zone, (2) large muscovite flakes in the gneiss (retrograde?), (3) sheared granitic dikes, (4) a younger slightly foliated granodiorite/diorite intrusion, (5) cross-cutting aplite, (6) Mesozoic spheroidally weathered lamprophyre dike -- please do not remove, (7) garnet-bearing granite pegmatite, possibly related to the Andover, and (8) evidence for two generations of granite and several episodes of folding. At the end of the outcrop on the north side of the ramp, cross to the south side and proceed around the cloverleaf along 495. Excellent examples of the mylonite gneisses and blastomylonites common in the "western Nashoba slice" are found here. Note the garnet-bearing muscovite granite has also been affected by the ductile deformation. Continue along 495 to 140 overpass, climb the bank and low fence, or return to cars via original route.

Return to cars, continue a very short distance on Colonial Dr.

6.2 0.1 Take a left (north) onto Cross St.

6.4	0.2	At Y keep to the right, on Cross St.
7.3	0.9	Rocky Pond Granite outcrops on right. Continue on Cross St.
8.0	0.7	5-way road junction. Turn left onto Central St. and proceed west.
8.2	0.2	Rusty-weathering exposures of Boylston Schist.
8.4	0.2	Turn <u>sharp</u> left (150 degrees) onto old dirt road and park. <u>Stop 4.</u> Exposures along the road. PLEASE BE CAREFUL OF TRAFFIC ON THE TURN IN THE ROAD.

<u>STOP 4.</u>

Boylston Schist. The Boylston Schist (Goldsmith et al., 1982a) is only found in the "western Nashoba" slice (Figure 2) at high grades of metamorphism. Whether it is associated with rocks of the Merrimack Trough or the Nashoba Block remains a problem. The Boylston is a rusty-weathering, graphitic, sillimanite-garnet mica schist or granofels. Bedding and layering are rather indistinct and the rock is commonly massive. In this locality, large flakes of muscovite (retrograde?) are common. Abundant sillimanite is easily observed in hand sample, and two generations of sillimanite can be observed in thin section.

Return east on Central St. to 5-way road junction.

8.9	0.5	At 5-way road junction, turn left onto Cross St. toward Mt. Pleasant Country Club.
8.95	0.05	Take an <u>immediate right</u> onto Green St. across from the Mt. Pleasant Country Club.
10.2	1.25	Junction with Warren St. Keep to the left, continuing on Green St.
10.7	0.5	Turn left onto Mile Hill Rd. and park immediately on either side of road. Outcrops of Rocky Pond Granite in the woods on both sides of Mile Hill Rd. <u>Stop 5.</u>

<u>STOP 5.</u>

Rocky Pond Granite. This granite (mgr of Zen, 1983b) is named (Munn, in progress) for exposures near Rocky Pond in Boylston, 1 mile south of this stop. Zartman and Naylor (1984) attempted to date the granite of "West Berlin" with samples from the Wachusett-Marlborough Tunnel but were unable tt accurately determine its age. Because of possible correlation problems between the tunnel samples and surface exposures, a new name is given to this granite. The rock is a foliated 2-mica granite cut by muscovite bearing aplite dikes.

Sedimentary xenoliths and inclusions are common, with some resembling the Vaughn Hills Formation, which lies along strike south of the granite (Figure 2).

Continue north on Mile Hill Rd.

12.1 1.4 Turn right onto Linden St. Proceed east.

13.3 1.2 Turn left onto Barnes Hill Rd. Proceed north.

14.2 0.9 5-way road junction in West Berlin. Turn sharply left onto Willow Rd.

14.3 0.1 Park on either side. Outcrops along the road. Stop 6.

STOP 6. Tadmuck Brook Schist. Exposed here, and around the corner on the street heading north, are outcrops of a rusty-weathering, gray, graphitic, staurolite-andalusite-muscovite-biotite phyllitic schist with thin quartzose interlayers. Megascopic porphyroclasts of andalusite are readily identified. Note the late, ovoid-shaped biotite flakes that have grown across the foliation. Also note the lenticular augen or lenses of coarse grained polycrystalline material (some include andalusite and staurolite) enveloped by the finer grained micaceous matrix. This rock is interpreted to be a phyllonite or mylonitic schist.

Reverse direction and return to the 5-way junction in West Berlin.

14.4 0.1 At 5-way junction, turn right onto Derby Rd. Proceed south.

15.1 0.7 At T-junction with Linden St., turn left and proceed east.

15.6 0.5 Junction with Lyman Rd. Stay straight on Linden St.

15.8 0.2 Cross railroad tracks; park immediately on left. Walk along road and across the Corp of Engineers Flood Control Dam to outcrops at east end of dam. Stop 7 and lunch by the lake.

STOP 7. Nashoba Formation; calc-silicate granulite and marble, possibly correlative with the Beaver Brook or Fort Pond Members of Bell and Alvord (1976). Excavations for the dam spillway have produced the most complete section through one of the calc-silicate/marble units in the Nashoba in this area. The calc-silicate granulites here contain diopside, actinolite, phlogopitic biotite, plagioclase, tourmaline and opaques. The deformation has been largely by flow in the more calcareous layers, and tectonic "fish" of originally more dolomitic or shaley layers are readily observed. Schistose rocks and amphibolites are present at the north end of the cut, and garnet bearing amphibolites are present in outcrops to the south.

Return to cars and continue east on Linden St.

16.3 0.5 Junction in Berlin at blinking light. Turn right on Rt. 62; proceed east.

16.5 0.2 Junction with Pleasant St. Continue east on Rt. 62 to junction with I-495.

17.3 0.8 Note typically orchard-covered drumlin on right.

18.2 0.9 Junction with I-495. Proceed under I-495 and park adjacent to entrance ramp to I-495 North. Outcrops along entrance. Stop 8.

STOP 8. Nashoba Formation; biotite gneisses. The excellent exposures here show some of the diversity in the biotitic gneisses of the Nashoba. Generally, the gneisses are gray to dark gray, medium grained, unevenly layered to well layered plagioclase-quartz-biotite gneisses or schists with widely varying percentages of these minerals in individual layers. Garnet, sillimanite and K-feldspar are locally present, as are large flakes of muscovite (retrograde?). Minor amphibolite layers are also present. A relatively small amount of migmatite is seen here in comparison to the next stop. Note the presence of ductile shear zones and younger post-deformational granitic dikes.

Proceed along entrance ramp onto I-495 North. The cuts here are biotite gneisses typical of the Nashoba Fm. Proceed north on I-495.

20.5 2.3 Note typical Nashoba Fm. in roadcuts.

23.2 2.7 Rusty-weathering sillimanite schist of the Nashoba Fm. in roadcuts.

25.9 2.7 Take Exit 28, Rt. 111 - Boxborough, Harvard.

26.2 0.3 Top of exit ramp. Turn right (west) on Rt. 111 across I-495.

26.4 0.2 Park on right by large outcrops on Rt. 111, opposite entrance to I-495 South. Stop 9.

STOP 9. Nashoba Formation; gneisses and migmatitic gneisses. The large outcrops along the north side of Rt. 111 contain an excellent example of the migmatitic Nashoba. Here biotite gneisses are interlayered with migmatitic gneisses and pegmatites. Sillimanite is commonly present with biotite in the selvages along the rim of the melted material. Muscovite is again present in large, retrograded? flakes. Two generations of pegmatites are present here, the earlier having been deformed. Several generations of folding are also visible. Note the late brittle faults with gouge cutting the outcrop. It is believed that most of the pegmatite and granite in this outcrop is more or less locally generated by anatexis of layers with the appropriate composition. Note how the percentage of melt changes with the composition of the original layer. It is widely believed, but not yet firmly established, that where the melting in the Nashoba reached sufficient proportions for the magma to coalesce and move, it formed the Andover Granite.

Turn left onto entrance to I-495 South and proceed south on I-495.

34.9	8.5	Outcrops of Nashoba Fm.
35.1	0.2	The valley in front of you coincides approximately with the Assabet River fault zone.
36.3	1.2	Junction I-290 and I-495; leave I-495 at Exit 25A "to 85, Marlborough".
36.6	0.3	Follow signs "to 85, Marlborough and Hudson".
36.8	0.2	Note carbonate-filled brittle shear zone on right.
37.0	0.2	Pull over and park with care on the right shoulder, beyond merging traffic. <u>Stop 10.</u>

<u>STOP 10</u>

Straw Hill Diorite. The exposures at this cloverleaf are complex and show a wide variety of features. Our principal purpose in stopping here is to see the Straw Hollow Diorite as a representative of the Silurian (Zartman and Naylor, 1984) intermediate plutonic intrusions in the Nashoba Block (Straw Hollow, Assabet, Sharpners Pond; for more detail on these, see trip by Hill et al., this volume). Both a foliated and a non- or more weakly foliated variety of the hornblende-quartz diorites are present here. They have been intruded by garnet bearing pegmatites possibly related to the nearby Andover Granite. Blastomylonites believed to be associated with the Assabet River fault zone are present here and will be seen as an example of the type of deformation common along some of the larger shear zones in the Nashoba Block. If time permits, an example of a relatively late brittle shear zone, filled with carbonate mineralization, will be visited and provide contrast with the earlier ductile deformation.

Continue straight on "to 85" east.

38.0 1.0 Turn right (south) at traffic light onto Fitchburg St.

38.1 0.1 Turn left at entrance to Assabet Valley Regional Vocational High School.

38.3 0.2 Park in parking lot. Proceed to exposure by the building. Stop 11. NO HAMMERS PLEASE.

STOP 11 Andover Granite. The Andover Granite is widely exposed in the northern Nashoba Block (Zen, 1983b) and contains strongly to weakly foliated varieties (Castle, 1964; Hill et al., this volume). This brief stop is to illustrate the foliated phase of this important plutonic rock of the Nashoba Block. Here the Andover is a strongly foliated biotite-muscovite granite to granodiorite with somewhat rounded feldspars. The foliation is at least partially caused by shearing. The granite is intruded by muscovite bearing pegmatites.

Return to Fitchburg St.

38.5 0.2 Turn right (north) onto Fitchburg St.

38.7 0.2 At traffic light, turn left and return to I-495.

39.0 0.3 Passing outcrops of Straw Hollow Diorite on right.

39.6 0.6 Continue toward Exit 26A for I-495 South.

40.1 0.5 Take Exit 26A to I-495 South, toward Milford, and proceed south on I-495.

41.7 1.6 Passing outcrops of garnet-bearing Andover Granite on left.

41.8 0.1 Leave I-495 at Exit 24B, for Rt. 20 West toward Northboro.

42.2 0.4 Top of exit ramp. Turn right onto Rt. 20 West.

42.3 0.1 At first traffic light, make a U-turn and proceed east on Rt. 20.

42.5 0.2 Turn right onto entrance ramp for I-495 South and park with care along right shoulder. Stop 12.

STOP 12. Marlboro Formation. These outcrops are typical of amphibolites in the Sandy Pond Member of the Marlboro (Bell and Alvord, 1976; DiNitto et al., this volume). Characteristically they are fine to medium grained, massive to foliated hornblende-plagioclase +/- quartz, biotite and epidote amphibolites and layered amphibolites.

Continue along entrance ramp and proceed south on I-495.

43.9	1.4	Passing outcrops of the Indian Head Hill pluton, granite and diorite (see DiNitto et al. and Hill et al., this volume).
44.3	0.4	Valley trending NE-SW in front of you is the approximate trace of the Bloody Bluff fault zone.
44.9	0.6	Crossing the Bloody Bluff fault zone.
45.4	0.5	Passing outcrops of the Westboro Fm., quartzite and interlayered schists. The northern end of the exposure has some tectonically interleaved amphibolite and biotite schist.
46.2	0.8	Leave I-495 at Exit 23A for Rt. 9 East, toward Framingham. You are now entering the northern end of the "Milford Terrane".
46.5	0.3	Merge onto Rt. 9 East (stay in right lane).
46.9	0.4	Take a sharp right onto Washington St.
47.0	0.1	Turn left into private road for Data General.
47.2	0.2	Park in western side of parking lot by exposures. Stop 13.

STOP 13. Milford Granite Terrane. These exposures contain a foliated white to pink biotite granite with smokey to bluish, somewhat granulated quartz and prominent allanite crystals. The biotite occurs in "clots" and makes up 5-10% of the rock. A full discussion of the granitic rocks of the "Milford Terrane" is beyond the scope of this trip. For details see Wones and Goldsmith (in press). The rocks at this stop are assigned to the Scituate Granite Gneiss by Zen (1983b). Thus, we have crossed out of the Nashoba Block and onto rocks of the 600 m.y. old Avalonian granitic suite of the Boston Platform.

Probable end of the trip. Return to Rt. 9, proceed east to Rt. 95, and follow to NEIGC headquarters in Danvers.

Optional additional stop in the Bloody Bluff fault zone if time permits. Return to Washington St.

47.4	0.2	Turn right onto Washington St.
47.5	0.1	Re-enter Rt. 9 East.
48.5	1.0	Get into left lane.
48.8	0.3	Turn left onto Middle Rd.
49.0	0.2	Turn left onto John Matthews Rd.

49.3 0.3 Turn right on Parkerville Rd.

50.1 0.8 Crossing Wachusett Aquaduct.

50.2 0.1 Junction Main St., Rt. 30. Turn left and proceed west.

50.8 0.6 Pull into yard by new house. Flat outcrop in the yard and outcrops on the south side of Main St. Stop 14.

STOP 14. Bloody Bluff fault zone. Highly sheared and mylonitized granitic rocks are present at this site near the Bloody Bluff fault. The rocks were subjected to a late brittle deformation in addition to earlier episodes, as evidenced by fractures now filled with specular hematite. The outcrop on the south side of Rt. 30 had a more mafic original protolith.

End of Trip. Turn around and return east on Main St., Rt. 30.

51.4 0.6 Turn right onto Parkerville Rd. and follow to Rt. 9. Take Rt. 9 East to the Massachusetts Turnpike or to Rt. 95 to Danvers.

BOSTON BASIN RESTUDIED

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Abstract

Recent mapping of the Boston Basin has shown that the sedimentary and rhyolitic and andesitic volcanic rocks are interbedded and that all lithic types interfinger, reflecting a wide range of depositional environments, including: alluvial, fluviatile, lacustrine, lagoonal, and marine-shelf. In addition to the well-known sedimentary rocks, such as argillite and conglomerate, we now recognize calcareous argillites, gypsiferous argillites of hypersaline origin, black argillite, red beds, turbidity current deposits, and alluvial fan deposits. The depositional setting seems to have been a tectonically active, block-faulted terrane in a coastal area. The granites that underlie these rocks are approximately the same age, some of them intruding the lower part of the sedimentary and volcanic section and feeding the rhyolitic volcanics within the section. All of this took place in Late Proterozoic Z-Cambrian time. No unconformity is evident at the base of the Cambrian. The Basin has a well-defined N. 80° E. regional strike. Folds are few and large, and axes are overturned to the south. The Basin is broken into at least 10 longitutinal fault-slices paralleling the regional strike. It is also broken by many smaller faults, mostly younger in age, including some that date from Triassic-Jurassic time. The Boston Basin is terminated on the north by the Walden Pond fault. There is reason to think that the much faulted, wide zone of mafic rocks to the north of the Basin is an interplate-collision suture.

Introduction

Although Boston was the first city of the United States to be mapped geologically (Dana and Dana, 1818; LaForge, 1932, pl. 3), it probably will be one of the last to be mapped satisfactorily. The reason is simple: bedrock for much of the area is deeply buried by Quaternary deposits (Kaye, 1982). LaForge (1932) faced the mapping problem created by this situation squarely and honestly in his geological mapping: he left blank those areas lacking outcrops. Today's geologists are not as easily satisfied, mainly because there are other types of data besides natural outcrops, and, in consequence, they feel obliged to color the entire map area. For example, there are rock cores brought up by diamond drilling, and several hundred rock cores were utilized in recent mapping. Deep excavation to bedrock for construction purposes and tunneling in bedrock yielded invaluable information. Use was made of lithologies of pebbles and boulders in Pleistocene glacial till, because once the directions of glacial transport are understood, these show the azimuths of the bedrock sources of the pebbles. In places, such as the Boston Harbor area, these data were of great help in the recent mapping.

The writer spent well over a decade mapping the bedrock of the Boston area. A preliminary map of the Newton, Boston North, and Boston South quadrangles has been published (Kaye, 1980). The final map of the entire block of eight quandrangles covering the Boston Basin with an accompanying text will be forthcoming. The following notes briefly summarize some of the more important, or interesting findings of this work. The conclusions in some areas differ from those shown on the recently published geologic map of Massachusetts (Zen and others, 1983).

Boston Basin

The term Boston Basin has been used variously: (1) as a topographic lowland, (2) as a structural basin, and (3) as the outcrop of a group of rocks of similar age and origin. Generally, the three basins were thought to coincide. The faulted northern margin of the lowland was called the Northern Boundary Fault and was thought to be the north limit of the rocks of the Basin. The rocks were named the Boston Bay Group, and were described by LaForge (1932) as consisting of Roxbury conglomerate below and Cambridge slate above, with the Brighton melaphyr intruded into and interbedded with the lower part of the Roxbury conglomerate. In LaForge's opinion, these rocks were not older than Devonian and not younger than Carboniferous.

The recent mapping has demonstrated a more complicated stratigraphy. The sedimentary rocks vary texturally from coarse-grained to fine-grained, grading and interdigitating, both laterally and vertically, in a lithofacies complex. The volcanics are of two contrasting types, rhyolites and soda andesites (both spilites and keratophyres are present), which are interbedded with the sediments and occur in at least six horizons, not all of which are in the lower part of the section (Kaye,1980, fig. 2). In addition, granite, of the type formerly thought to form only the basement beneath the basin, is now seen to intrude the lower part of the basin deposits and to form the parent magma of the rhyolitic volcanics. The definition of the Boston Bay Group as given by LaForge is clearly in need of revision. The writer is avoiding it and informally using, instead, the term "Boston Basin rocks" to denote rocks that were deposited on the surface (both volcanic and sedimentary) in Proterozoic Z-Cambrian time.

Finally, the basin structure of the Boston Basin is true only in a general sense. If we were to draw a perimeter around the known outcrops of the sedimentary and volcanic rocks — the Boston Basin — we would find that for only about half its length are the lowermost, or basal, rocks in contact with older rocks. The rest is faulted. Moreover, within the basin itself, the centripetal structure is lost, and instead we find a series of long fault blocks, some of which are structurally high and some structurally low in a seemingly disorganized way.

Age

Until several years ago the age of the Boston Basin rocks was thought to be Devonian-Carboniferous. It now is believed to be Late Proterozoic Z-Cambrian (Kaye, 1978; Kaye and Zartman, 1980; Lenk and others, 1982). This new point of view is based on radiometric dates and microfossils, but these are the slow outgrowth of the geologic remapping of the area, as the following sequence of findings and ideas will show:

1. Mapping of the southern part of the Boston Basin (Hingham Bay, Quincy, Milton, North Weymouth, and Braintree) showed that the fossiliferous beds of Cambrian age in the Fore River estuary have the same strike and dip as the Boston Basin rocks adjoining to the north (N. 80° E., steep S. dip). Clearly, there was no angular unconformity between the two groups of rocks. If we than assumed that all of these rocks were part of the same stratigraphic sequence, then the fossilferous Cambrian rocks would be on top and the nonfossiliferous Boston Basin rocks on the bottom and, therefore, Precambrian in age. This would then explain the absence of fossils in the latter rocks — a situation that had long proved puzzling.

2. Rocks lithologically similar to the fossiliferous Cambrian of the Fore River estuary were found in several places within the Boston Basin, indicating that there was nothing particularly distinctive, lithologically, about Cambrian rocks other than the presence of fossils.

3. Mapping of the rocks along the northern margin of the Quincy Granite pluton failed to confirm the unconformity separating Cambrian rocks from Late Paleozoic Boston Basin rocks that W. O. Crosby (1900) recognized. The coarse conglomerate that he considered to be the basal conglomerate of the Boston Basin, to the writer seemed similar to other conglomerate zones in the Basin section and therefore probably of no fundamental stratigraphic importance.

4. The "fossil tree trunk" of Burr and Burke (1900), which provided the clinching argument for the Carboniferous age of the Boston Basin, was restudied and found to lack any internal cellular structure that would prove its organic origin. The cylinder consists of medium-grained arkosic sandstone. If it had been derived from a plant stem, it is clear that all tissue had disappeared prior to the deposition of the sand now occupying its space. It seems more likely to represent a nonorganic structure, such as a cyclindrical sandstone dike, for example. The transverse ridges on the surface of the cylinder are probably the result of soft sediment consolidation during diagenesis.

5. Lacking fossil proof of the Precambrian age of the Boston Basin rocks, radiometric dating was used. Zircons from the rhyolite in the basal part of the section were analyzed for U-Th-Pb and an age of 602 ± 3 m.y. was obtained (Kaye and Zartman, 1980). This is Late Proterozoic Z, or about 30 m.y. before the onset of the Cambrian.

6. The above age was confirmed a few years later by the accidental finding by the writer of a microflora in argillite from high in the sedimentary section (Lenk and others, 1982). Thin-sections of this rock, taken from the new subway tunnel under Massachusetts Avenue, Cambridge, had been made in order to determine whether the well-developed graded-bedding there was true or spurious (if true, the beds were completely overturned, which would be a unique occurrence for the Boston Basin). Abundant minute, opaque spheres were seen and recognized as possibile acritarchs of Precambrian age. The thin-section was shown to Professor Elso Barghoorn, Harvard University, and his students, Strother and Lenk. After acid digestion of the rock, they uncovered the organism beneath a very thin coating of pyrite (Lenk and others, 1982). The microflora agreed with Late Proterozoic Z (Vendian) fossils from localities in North America, Greenland, Norway, and Central U.S.S.R.

The Avalonian Zone

The rocks of the Boston Basin and those of the Avalon Peninsula, eastern Newfoundland, have much in common. They are the same age; are of low metamorphic grade; have both mafic and silicic volcanics concentrated in the lower part of the section; and consist of about the same variety of sedimentary rocks, including such distinctive types as conglomerate, argillite, red beds, and black argilllite. In addition, rocks in both areas are intruded by granite (King, 1980). The two areas are not identical, however. In Newfoundland, the lithofacies seem to be more broadly developed than in Boston and, therefore, are of lithostratigraphic value for regional mapping. In Boston, the grain of lithologic development is too local for a very usable lithostratigraphy. In addition, the statigraphic positions of such distinctive horizons as red beds and black argillite seem to be different in the two places.

Within the tectonic framework of the Caledonian-Appalachian orogen (Williams, 1978; Wones, 1980), the Boston Basin and all the southeastern corner of New England (southeastern Massachusetts and Rhode Island) fall in the Avalonian zone.

Rocks of the same age having somewhat the same molassic-flysch composition are also found in the eastern part of the southern Appalachians, as well, and probably form part of the same tectonic belt.

Walden Pond Fault

In the Boston North quadrangle, the Boston Basin, as it is now understood, extends about 7 km north of the old Northern Boundry Fault to where the volcanics and intruding Dedham granites are sharply cut off by the Walden Pond Fault (Kaye, 1980; incidentally, the fault is named after Walden Pond in Lynn Woods, Lynn, not after the pond of the same name in Concord). In the Lexington quadrangle, however, the Walden Pond fault coincides with the old Northern Boundary Fault, passing along the base of the topographic scarp. Traced to the east into the Boston North quadrangle and beyond, from the mouth of the valley of the Mystic Lakes, it follows an arcuate course, passing through Stoneham, Wakefield, Walden Pond, Sluice Pond, and Flax Pond to a point south of Phillips Point, Swamscott (Lynn quadrangle). From there it bends sharply to the northeast, passes down the middle of Marblehead Harbor to a point off Bakers Island where it bends sharply northeast, following along the south side of Cape Ann (Fig. 1).

Rocks on the north side of the fault are everywhere mafic amphibolitic rocks soaked by red granites and cut by later gray granite plutons. These will be discussed below under the heading "Suture(?) zone".

Basement

The basement beneath the Boston Basin rocks was long thought to be the Dedham granites, for in many places the rhyolites, soda andesites, and sedimentary rocks of the Boston Basin rest directly on these pink and light gray granites (in the Boston area, it should be noted, these rocks are normal granite, not granodiorite). At that time, the age of the Boston Basin was thought to be Late Paleozoic, the granites probably Precambrian (LaForge, 1932). Today, volcanics and associated sediments are known to be Late Proterozoic Z in age, as are the granites (Zartman and Naylor, 1984). Indeed, in several places we can see that the granite intruded the lower part of volcanic and sedimentary section (Kaye, 1980). The picture now conveyed by our data is that the extensive outcrop of Dedham-type granite making up most of southeastern New England (Zen and others, 1983; Zartman and Naylor, 1984) was built up by many separate surges of granitic magmas into a tectonically active crust. Contemporary with this, erosion and deposition of sediment and the extrusion of lavas were taking place in the rugged terrain. Thus while in some active fault blocks recently emplaced and freshly crystallized granite was already exposed to erosion, in adjoining blocks granitic magmas were still being intruded.

As for the quartzites and interbedded argillites of the Westboro Formation, which are engulfed by these granites and therefore were formerly thought of as being much older than the Boston Basin (LaForge, 1932), these are now seen as being part of the Boston Basin sedimentary section.

What, then, is the basement beneath the Boston Basin rocks? The answer to this — once so certain — now eludes us. We must probe deeper.

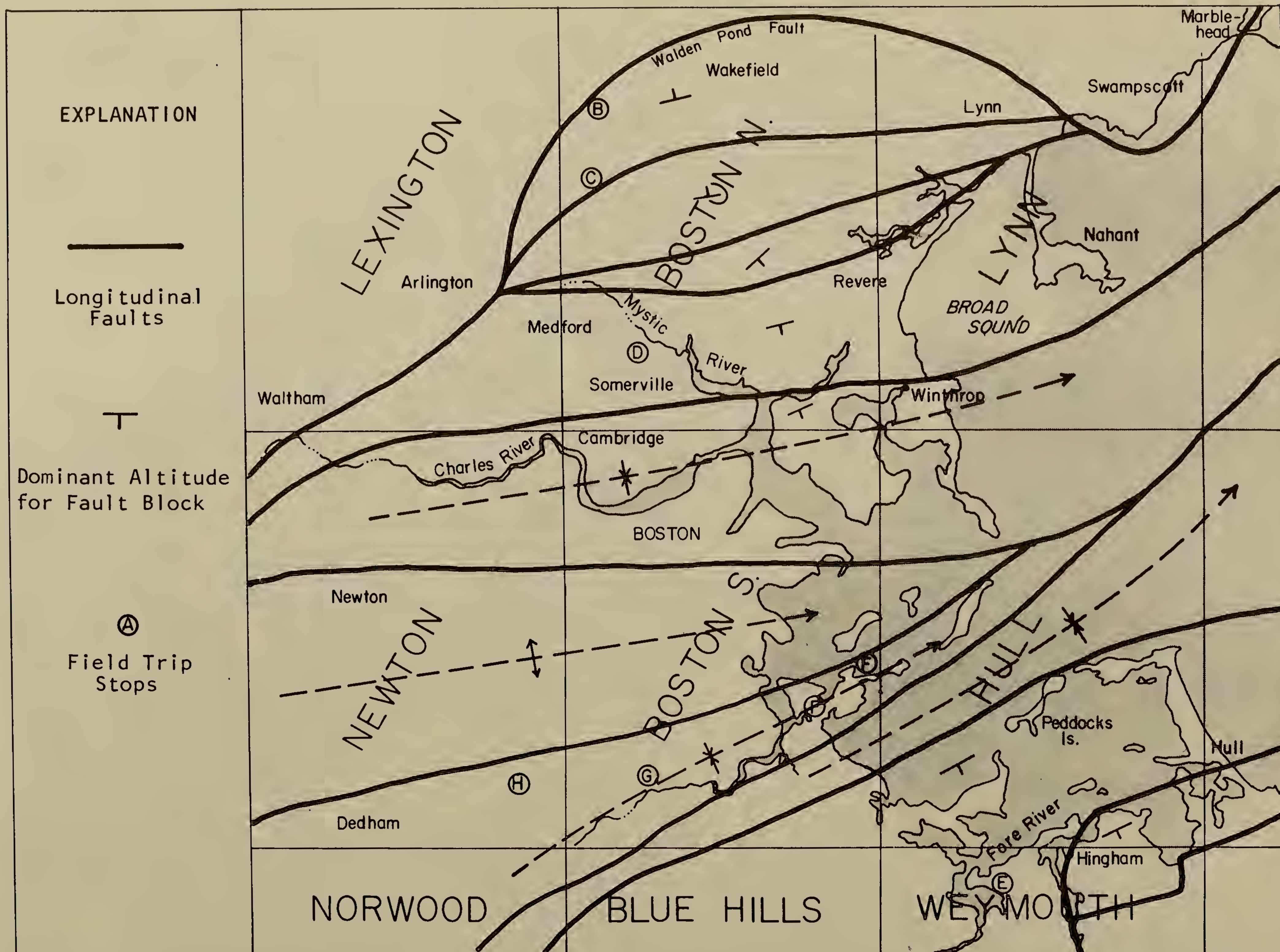

Figure 1.--Major longitudinal faults of Boston Basin and field trip stops.

Dedham Granite

The granites, called Dedham Granite by W. O. Crosby (1880) and Dedham Granodiorite by Emerson (1917) and LaForge (1932), are normal leucocratic granites consisting dominantly of K-spar, lesser albite-oligoclase, quartz (everwhere highly strained) and pyroboles and primary biotites that show advanced alteration to chlorite. These granites enclose large xenoliths, some of which are foliated and metamorphosed (Zen and others, 1983) while others only show slight alteration. Among the latter in the Boston area is a metagabbro cropping out in the town of Dedham (Kaye, 1980) and the large masses of argillite and interbedded fine-grained quartzite called the Westboro Formation.

The transformation of Dedham Granite into vitrophyric rhyolite can be studies in a number of places (viz. West Roxbury-Hyde Park-Stony Brook Reservation; Needham; Medford-Middlesex Fells Reservation; Malden, Maplewood; Malden, Linden). Typically, the grain-size of the granite diminishes progressively towards the rhyolite until the rock becomes aphanitic. In thin-section, the transformation is seen as movinge from granitic to granophyric or aplitic texture, to vitrophyric — the typical texture of the rhyolites.

Volcanic rocks

Volcanic rocks are important in the lower part of the Boston Basin section where they are interbedded with sedimentary rocks. This relationship is clearly visible around the perimeter of the Basin: in the Hull-Nantasket area (Crosby, 1893), in Hingham (Crosby, 1894), in the Neponset River basin (Milton, Mattapan, Hyde Park, West Roxbury), and in the Medford-Malden-Wakefield-Lynn area on the north side of the Basin (Kaye, 1980). The volcanic rocks are both felsic and femic, or more precisely, rhyolitic and soda andesitic. There seems to have been one major and one minor rhyolitic eruptive interval involving a number of vents and producing rocks ranging from coarse explosion-breccias to dense welded tuffs. We see remnants of rhyolitic domes, crater plugs, and spines. These are best viewed in the Mattapan area. The rhyolite is local in distribution and is absent in some places. It is clearly thickest on the north, in the Medford-Malden-Wakefield-Lynn zone, and thinnest in Hingham, where it occurs only in a few explosion pipes. There seems to be little reason for separating the rhyolites on the north side of the Basin from those further south, as was done by LaForge (1932), who called the former the Lynn Volcanic Complex and the others the Mattapan Volcanic Complex.

A second rhyolitic eruptive event took place shortly after the onset of andesitic eruption. This produced small bodies, that in some places seem to be embedded in the andesites. Several of these crop out just east of Mattapan Square (Field Trip Stop G). They probably represent rhyolitic crater spines and are of particular interest because they appear to have collapsed into an andesitic lava field. Similar structures can be seen on Mt. Hood, Saugus (Boston North quadrangle).

The soda andesites, formerly called melaphyres, are more widespread than the rhyolites, though nowhere are they as thick as some of the rhyolite domes. There were four or more separate eruptive intervals, ranging well up into the basin section. Flows grade laterally into pyroclastics and tuffaceous sediments. These volcanic horizons are separated by sediments of all textural types but dominantly by conglomerate. This can be seen in Hull, Hingham, the basin of the Neponset River, Newton-Brookline (where there are three keratophyric horizons within the conglomerate of the Central Anticline), and the east-west zone that extends from Newton Lower Falls, on the west, to Allston-Brighton on the east.

The soda andesites are of two types: keratophyres and spilites. Both are albite-oligoclase rocks, which, incidentally, is a characteristic of almost all igneous rocks — intrusive and extrusive — of the Boston Basin. The keratophyres are typically dense black, maroon, rarely greenish, massive to layered flows and flow breccias. The spilitic lavas contained more volatiles and give evidence of subaqeous eruption. They are greenish to blue-green, locally purple to maroon, and coarsely amygdaloidal. In three places pillow structures have been seen (Hull, Wakefield, Rte 128 in Newton Lower Falls).

Sedimentary Rocks

Lithofocies and depositional environments

Sedimentary rocks clearly vary markedly in many essential parameters when traced along their strike. Conglomerates, for example, pass into argillite without an intermediate sandstone facies; red beds interfinger with, and grade into, gray beds; volcanics grade into pyroclastics, volcanoclastics, or even into nonvolcanic sediments. We see evidence of marine deposits grading laterally into lagoonal, then into lacustrine deposits, and these grading into alluvial and possibly colluvial deposits. This heterogenity of sediment types points to a highly unstable depositional environment, one that fits a tectonically active fault-block, basin-range model.

The variability of lithofacies compounds the problem of geologic mapping. In a terrane of sparse or interrrupted outcrops, a change of lithology along strike can be interpreted as either the result of faulting or of facies change (folding is generally eliminated as a choice because of the scarcity of tight folds). Which one of these the mapper makes in a given situation is obviously a matter of value judgement.

Conglomerates

Pebble- to cobble-size conglomerate is important in the lower part of the sedimentary section, occuring more sparsely and only as thin, lenticular sheets at higher stratigraphic levels. The conglomerates are almost everywhere interbedded with volcanic rocks, both rhyolites and soda andesites.

Pebbles in the conglomerate typically are of three main lithic types: volcanics (soda andesite and rhyolite); fine-grained quartzite of the nonglassy type; and Dedham-type granites. The volcanics are identical to those that are part of the Boston Basin, the quartzites are like those in the Westboro Formation, and the granites are clearly of the Dedham granite complex, which, as explained above, is associated with the basin. These sediments, therefore, are endogenous, that is, derive from the same rock terrane as they are being deposited into. Rocks making up the lower part of the basin clearly had been uplifted along faults, eroded, and deposited, and redeposited—the sort of thing that happens in tectonically active areas today.

Bedding of the conglomerate is visible in places, but in others, the rock is massive. Bedding is generally expressed by parallelism of the long dimension of the pebbles and by occasional sandstone layers. Strikes and dips of the conglomerates are generally somewhat at variance with those of the fine-grained sedimentary rocks flanking them; this is attributed to the conglomerate having been deposited with considerable initial dip, either as in alluvial fans on steep slopes or as in deltaic foreset bedding.

Squantum Diamictite ("tillite")

For many years the badly sorted, lenticular conglomerates of Squantum (Boston South quadrangle) have been classified as tillite (LaForge, 1932). Several decades ago this interpretation was challenged by Dott (1961), who recognized that they also had the characteristics of subaqueous slumps, or turbidity current deposits. The writer's work supports Dott's interpretation mainly for the following reasons: (1) pebbles with soled, or faceted shapes that are characteristic of clasts of Pleistocene tills are lacking; also lacking are pebbles with glacial scratches, or striations; (2) with the exception of rare, angular blocks of argillite, lithologies of the clasts are the same as these in normal conglomerate except that the content of soda andesites pebbles is higher; (3) the "tillite" occurs in the stratigraphic position of a known spilite horizon and the red and green coloration of the associated graded-beds is characteristic of volcanogenic sediments, and (4) the pronounced graded-bedding of these sediments is especially characteristic of turbidity-current deposits, as pointed out by Dott (1961).

The writer sees the "tillite" as the result of a volcanic eruption, or eruptions, shaking down unstable subaqueous slopes of clay, sand, and gravel.

Argillite

Argillaceous rocks of the Boston Basin, formerly called "slate", are now termed argillite. Argillite is a more appropriate term, for several reasons. The first is that true slate is very rare in this terrane. Secondly, the rocks are metamorphic but of very low grade; they are harder and stronger than normal shales, mudstones, claystone, etc.; they lack fissility; and parting on bedding-planes is a property that is poorly developed. Included in the classification argillite are fine-grained argillaceous sandstones, tuffaceous argillites, calcareous argillites, gypsiferous argillite, and other types of fine-grained sedimentary rocks. These rocks range in color from cream through light to dark shades of gray, often with greenish, bluish, and red chroma. They range from very well-stratified to massive. There are many zones of penecontemporaneous deformation. Small depositional unconformities are common.

The rocks are of low metamorphic grade. Sericite, chlorite and quartz are the principal minerals. In thin-section, the flaky minerals are seen to be arranged in a triaxial decussate structure. In a few places one finds schistose foliation, probably produced by shearing associated with nearby faulting.

Argillite underlies most of the lowlying part of the Boston Basin, particularly those places where the top of bedrock falls below sea level.

Calcareous Argillite

Individual beds of calcareous argillite are interbedded with normal argillite in many places, and range in thickness from several centimeters to 1.5 m. The unweathered color and appearance of this rock is very like that of normal gray argillite except that it is slightly lighter gray than noncalcareous rock with which it is interbedded. The fresh rock effervesces readily with dilute hydrochloric acid; on weathered surfaces, where it has been leached, it has a characteristic dull brown appearance. Calcareous argillite is found under Boston Peninsula, to the east under Boston Harbor and Logan Airport, in Somerville, Dorcester, and West Roxbury.

Sideritic argillite

Well-stratified, light-colored, even textured argillite ("claystone"), sprinkled with small siderite crystals, underlies the area around Porter Square, Somerville. Pebbles of this rock are found widely in the drumlin till of the islands in Central Boston Harbor.

Gypsiferous and dolomitic argillite

Much of the argillite under downtown Boston, Cambridge, and Inner Boston Harbor is densely sprinkled with small gypsum crystals, about 0.5 mm or smaller in length. In places this rock also contains clusters of larger crystals of dolomite. There seems little doubt that these are hypersaline sediments and therefore indicate the existance of saline depositional basins in the Boston Basin.

Red beds

Sandstone, arkose, and argillite that is pink and hematite-red in color, not unlike the Newark Series rocks of Late Triassic-Early Jurassic age, are found as an integral part of the Boston Basin. These crop out in three areas: (1) Broad Sound (Lynn quadrangle), (2) Milton-Quincy-Houghs Neck (Boston South and Hull quadrangles), and (3) North Weymouth (Weymouth quadrangle). within the Boston Basin. In the first and third of these areas, they comprise the Lower Cambrian Weymouth Formation. The second outcrop belt may also be of the same age, although this is not certain, and on structural grounds there is evidence it is older. In the Broad Sound area, pebbles of red bed were first noticed in glaial till in some of the drumlins in Winthrop and Deer Island. Pebbles then were found in Revere Beach, and diamond drill cores from Winthrop, Revere, and part of Lynn also showed the presence of these rocks in the underlying bedrock. Some of the beach pebbles of red argillite contain small, irregular masses of limestone with Early Cambrian fossils (Clark, 1923).

Red beds, exposed in road cuts in Milton during construction of the Southeast Expressway, were described by Billings (1976). These rocks extend east of that outcrop to the western part of Peddocks Island. Further east on Peddocks as well as to the west of the Southeast Expressway, the red coloration fades and the rocks grade into interbedded gray sandstone and argillite. This same transformation of red to gray possibly also occurs in Broad Sound area when we trace the red beds to the west under Revere, Everett, Somerville, and Medford.

The red beds under North Weymouth also grade into gray. On the east they are cut off by a fault but on the west, in Quincy, our drill cores indicate that they are altered to black hornfels by the intrusion of the Quincy Granite pluton (Field Trip; Stop E). A similar transformation is seen in the fossiliferous black hornfels in East Point, Nahant.

"Black" argillite

Very dark gray argillite occurs in several stratigraphic or structural zones. One runs through Davis Square, Somerville; another runs along the axis of the Charles River sysncline; another through Nut Island just north of Houghs Neck, Quincy; and still another under the northeast end of Long Island, Boston Harbor. It is interesting to note that this lithology also occurs in the Avalonian terrane of eastern Newfoundland (King, 1980)

Fossils

Microfossils

Identifiable microfossils were found in argillite taken from the subway tunnel under Massachusetts Avenue, Cambridge (Lenk and others, 1982). These minute plant organisms were thinly encrusted with pyrite; and, when first spotted by the writer in thin sections of argillite, they appeared simply as minute opaque spheres. Treatment with hydroflouric and then nitric acids dissolved rock and sulphide and revealed the organisms beneath (Lenk and others, 1982).

Three basic types of acritarchs were found: (1) spherical cells, (2) multicellular filaments of cylindrical form, and (3) dense colonies of minute spherical cells (Bavlinella cf. faveolata). This assemblage is similar to that found in European rocks of Vendian age (Late Proterozoic Z) which agrees with the zircon-age obtained for the rhyolites (Kaye & Zartman, 1980).

Trace fossils

Small spiral burrows are found widely in laminated gray and white argillite and fine sandstone, cropping out in an east-west belt that extends from Tufts University, Medford, on the west to the town of Winthrop, on the east. These fossil structures are localized mostly within the dark argillaceous laminae and are infilled with light-colored sand from overlying sand laminae, thereby making them clearly visible to the naked eye. Individual burrows are generally 2-4 mm in diameter. The nature of the organism responsible for them is not known. They are informally referred to as trace fossil "Spirochete".

In Winthrop, borings indicate that a zone of this burrowed rock is interbedded with Lower Cambrian red beds, thereby indicating a similar age for the burrows. Are they diagnostic of a Cambrian Age?

Base of the Cambrian

We know from fossils, outcrops, borings, and drift pebbles that the Lower Cambrian crops out in two, possibly three, areas in the Boston Basin: in the Fore River estuary (North Weymouth, Quincy), the Broad Sound area (Nahant, Lynn, Revere, Winthrop), and possibly in a belt of red beds that as yet has yielded no fossils and which extends from Milton (Billings, 1976) to western Peddocks Island (Hull quadrangle). In the Fore River Estuary, the contact of the lower Cambrian with underlying nonfossiliferous beds is not exposed but is spanned by several lines of drill holes. The structure appears to be conformable. The Broad Sound area is less clearcut. Our data come from outcrops on Nahant; from drill holes from Lynn, Revere, and Winthrop, and from the towns lying to the west; from drift pebbles in Nahant Beach and from the many drumlins in the Winthrop area. These data allow for several structural interpretations. While it is possible that there is an angular unconformity at the base of the Cambrian here, the data can also be construed as a conformable contact, as in the Fore River estuary. If it is the latter, then the redbeds with fossiliferous limestone nodules pass laterally, into gray argillites under the existing and former marshlands of the Mystic River Estuary in Everett, Malden and Medford. This last is the preferred interpretation.

It is not clear, then, where in the stratigraphic section the Cambrian begins and the Proterozoic ends. We find gray argillites and sandstones in the Proterozoic-Z section that resemble rocks, also lacking fossils, in the Lower and Middle Cambrian. There is no clean-cut lithologic distinction between Cambrian and Precambrian where red coloration and fossiliferous limestones are lacking. It is quite possible, therefore, that a complete depositional transition from Precambrian to Cambrian occurs. The coming-in of fossil evidence may or may not mark the chronologic time signal for the outset of the Paleozoic. This is clearly a subject needing study.

Stratigraphy

It is difficult, if not impossible, at this state of our knowledge to draw up a detailed stratigraphic section making use of the many lithologic markers for the simple reason that our knowledge of the specifics of faulting and other structural deformations, as well as our incomplete grasp of the details of facies variability, is too tenuous. It is very probable, for example, that at least some of the lithologic markers are repeated in the section. It is for these reasons that the long established lithologic-stratigraphic names (viz. Roxbury Conglomerate, Cambridge Argillite, etc.) were avoided on the preliminary geologic map of the three quadrangles (Kaye, 1980). It seems to the writer, that in discussing specific rocks, be it argillite or any other lithologic type of the Boston Basin, that it adds nothing to the discussion or to our knowledge to append the formational name when these names imply a stratigraphic position about which we are uncertain. At this point, the lithologic names seem sufficient.

Later granites

A uniformly fine-grained tonalite rims the south side of the Walden Pond fault in the Boston North quadrangle and extends irregular branches to the south into the Boston Basin rocks. In the Lexington quadrangle this rock occurs on the north side of the fault as well. It is found in bodies of similar ragged shape in many places north of the Boston area to as far as southern New Hampshire (Newburyport Quartz Diorite).

Another post-Boston Basin granite is the peralkalic Quincy Granite and the closely related Cape Ann and Peabody granites. The last two crop out just north of the Walden Pond fault, while the Quincy Granite pluton forms the south edge of the Boston Basin in Quincy and Milton. The Quincy Granite body has been recently discussed by Billing (1983).

Structure

Regional strike

A rigid regional strike of N. 80° E. ±5° characterize almost all the onshore and Boston Harbor portions of the Boston Basin. Offshore, to the east, under Massachusetts Bay, our data indicate that the strike changes to about N. 45°E. in the northeastern sector of the basin. Most substantial deviations from the regional strike can be attributed to one or more of the following: closure of folds, high initial dips (as in the conglomerate), drag on faults, and skewing of crustal blocks bounded by faults.

Longitudinal faults and fault slices

The basin is broken by nine long faults that are parallel or subparallel to the regional strike and which effectively divide the basin into ten parallel fault blocks, or slices (Fig. 1), each of which seems to be structurally independent (Table 1). In the few places where these faults have been seen in tunnels (Richardson, 1977) and in outcrop, they show little if any cataclasm. This may mean that either they are ductile faults or they have been "healed " by subsequent relithificaton of cataclastic debris.

The longitudinal faults may be Ordovician(Taconic) in age, generated by crustal plate closure. It is also conceivable that the faults originated in, or reflect, the position of the original block-faults of the proto-Boston Basin.

Other faults

There is good evidence of many faults in the Boston Basin, although it must be acknowledged that geologic mappers will differ on the details. Because they are generally weathered and etched by erosion, glacial or otherwise, the traces of faults on the surface are rarely visible. In tunnels, their visibility is highly variable, depending on such factors as: lighting conditions; washed vs. unwashed walls; obliquity of fault to tunnel axis; ductile vs. brittle faults; compound vs. single fractures; presence vs. absence of quartz and calcite veins; amount of cataclasis; amount of "healing"; dike infilling; etc. As can be seen from the maps of the several bedrock tunnels under Boston (see bibliography given in Kaye, 1980, Sheet 2), many mafic dikes give evidence of having been intruded along earlier faults. Beds on both sides of the dikes show considerable drag (certainly in excess of that required by wall dilation); in addition, some dikes mark lithologic breaks. The walls of these fault dikes are commonly well-striated, which is best explained by faulting, whether pre-dike or post-dike in age.

While the age of most of these faults is probably Permian, recent work has shown the importance of Middle Mesozoic faulting in the Boston Basin (Kaye, 1983). Two types of faults that were found cutting the Newark Series rocks of the Middleton Basin, north of Boston, are also found in abundance in the Boston Basin. These are steeply dipping N-S faults with strike-slip displacement, and NE-striking faults with well-developed cataclastic zones.

Folds

There are 4 large folds in the Boston Basin, each of which seems to be isolated within a single fault-slice (table 1, fig. 1). The axes of all these folds parallel the regional strike. All are asymmetric, and indicate compressional stresses from north to south. All plunge to the east, possibly an expression of downward crustal tilting to the east on a regional scale.

Several areas of smaller folds exist. One of these is the stretch lying between Harvard Square, on the south, and the campus of Tufts University, Medford, on the north, and extending an unknown distance to the west into the Lexington quadrangle. These folds measure 0.5 km or less in width (N-S) and probably do not exceed 50 m in height (closure). Three folds of about this size are also seen in Hingham in the Squirrel Hill-Crows Point area.

Table 1.—Major longitudal fault slices of the Boston Basin (see Figure 1)
(In sequential order, north on top, south on bottom.)

Name of slice	Relative age of rocks exposed, (No. 1 oldest; No. 10 Cambrian)	Dominant structure
Wakefield	1	Homocline, N. dip
Oak Grove	1	Do.
Malden	?	Homocline, SE. dip
Chelsea	10	Homocline, S. dip
Boston	5-8	Charles River syncline, plunges east
Brookline	1-4	Central anticline, plunges east
Neponset River	1-6	Syncline, plunges east
Nantasket Roads	4-8	Do.
Hingham Bay	6-10	Homocline, S. dip
Hull	1	Homocline, N. dip, some folding

Suture(?) Zone

The Walden Pond fault separates two very different rock terranes. To the south are the sedimentary and volcanic rocks of the Boston Basin and associated Dedham granites, all of which are Late Proteozoic Z-Cambrian in age. To the north we find a broad belt of mafic rock of hybrid origin into which several distinctive petrologic types of granite have been intruded. All these rocks are thought to date from the Ordovician and perhaps Early Silurian. This terrane extends about 25 km north of the Walden Pond fault to the Clinton-Newbury fault, where it is sharply terminated.

The mafic rocks are both massive and foliated and range from fine-grained to coarse-grained. All of the many thin-sections studied show evidence of metamorphism. The dark minerals generally include amphibole (mostly green hornblende), biotite, pyroxenes, and ore, in that order of importance. The feldspars range from orthoclase and albite to rare relic fragments of andesine. Apatite is always abundant. The rocks have evolved from several types of igneous rocks and, indeed, here and there we find fairly intact small bodies of gabbro.

Clapp, (1910, 1921) named these rocks the Salem Gabbro-Diorite, but I would modify this to the Salem Gabbro-Diorite Complex. This general family of mafic amphibolitic rocks has been stratigraphically subdivided (viz. Marlboro, Nashoba, Sharpners Pond Tonolite, etc.); but I think these mapping refinements mask the essential common origin of the terrane. The cross-cutting granites are of 4 main types. The most pervasive is a granite, generally pink to red in the Boston area but light gray further west. This occurs mostly as dikes (some of which are coarse permatites) and as bodies of ill-defined shape with long gradational contacts with the surrounding mafic rock. Indeed, the dense peppering of felsic minerals of this granite within the Salem-Gabbro-Diorite Complex rock suggests that the magma of the red granite had literally soaked the mafic rocks but nowhere digested it entirely and made a pluton of its own.

The other granites form well-defined plutons. One of these granites is that of the Cape Ann and Peabody plutons. The other is the two-mica Andover granite. Another is the Newburyport quartz-diorite (tonalite) which is found cutting the Salem Complex but which also extends beyond the faulted borders of this belt.

There are three other features of the mafic zone that are worthy of note, particularly if we are concerned about the origin of this zone. One is that there are broad bands of mylonite, measuring many hundred of meters in width and trending more-or-less parallel to the zone (Castle and others, 1976; Kaye, 1983). Another notable feature is the presence of several large grabens. One of these contains the large block of the Newburg Volcanic Complex of Late Silurian-Early Devonian age (Shride, 1976); another down-faulted block is the Middleton Basin (Kaye, 1983), a small basin of Newark Series rock (Triassic-Jurassic). Also there are a number of sizeable xenoliths of high-grade metamorphic sedimentary rocks (mica schists, quartzites, marble) embedded in the Salem complex. Lastly, there are at least two bodies of serpentinite in this zone, a rock that is not found elsewhere in southeastern New England.

This fault-bounded rock terrane stretches for over 180 km in an accuate course into south central Massachusetts and southeastern Connecticut (Harwood and Zietz, 1976; Lundgren and Eblin, 1972; Castle and others, 1976; Zen and others, 1983; Kaye, 1983). Wilson (1966) and then Skehan (1968, 1969, 1973a, 1973b, 1980) suggested that it marked the suture formed at the closing of Iapetus in Ordovician time by the collison of the Avalonian and North American crustal plates. This mechanism, in my opinion, offers the best explaination. For example, the mafic rocks themselves may have originated as oceanic crustal basalts, caught up and pinched in the suture, magmatized and/or matamorphased by the collision stresses. Large blocks of continental crust could have broken off the margins of the continental plates as they collided and subsequently were magmatized by the frictional heat generated by the closure stresses. Differences in the several granites may reflect differences in the compositions of the several crustal blocks. Similarly, the metamorphic xenoliths represent blocks that survived collision without fusion. The mylonites resulted from post-collision shearing along the suture, movement that might have persisted long after the initial collision. Lastly — if this model is correct — the field facts point to subduction to the south—i.e. the Avalonian plate overrode the North American plate.

In all objectivity, however it must be said that this important suture has also been postulated for structural zones in western New England. This interpretation can be reconciled, however, with the above if there were several sutures, one behind the other, resulting perhaps, from the compound nature of the continental margins. Such a margin would occur if island-arc accretionary zones lay in front of either, or both, continental plates.

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One-day field trip of
Boston Basin

Clifford A. Kaye, leader

Information

Maps

Standard U.S.G.S. 7 1/2' quadrangles, scale 1:25,000: Salem, Reading, Boston North, Boston South, Weymouth, Lexington, Newton (all in Massachusetts); U.S.G.S. Miscellaneous Field Studies Map MF-1241 (2 sheets). It is advised to bring these along.

Point locations given below for each stop (viz. 37,800E/14,750N) are in the 1,000m-grid of the UTM system, which is printed on the faces of all 1:25,000-scale topographic maps. The coordinates can be readily measured off with the use of dividers, employing the grid-lines on the map face and the meter-bar scale printed at the bottom of each map.

Caution: We will adhere to time alloted for each stop. A whistle will be blown once, 4 minutes before departure time; twice, two minutes before departure. We will leave promptly. If you miss the bus, you are in big trouble.

Stop A (20 minutes)
Salem quadrangle, 37,800E/14,750N;
west side of Newburyport Turnpike-Rte 1; just N of
B & M single track crossing, parking lot next to building

Typical Salem Gabbro-Diorite Complex, the rock making up the major part of the suture zone north of the Walden Pond fault. It is a dark gray to black, fine-grained rock with igneous texture but in places has a distinct foliation. Petrographically, these are unusual rocks, probably hybrid in origin. The felsic minerals are those of peralkalic granites in this area (orthoclase, albite, quartz) but combined with an excess of femics (green hornblende, biotite, and, in places diopside). Almost everywhere these rocks are complexly intruded by red granite as multiple dikes and dikelets, and as isolated crystals of feldspar and quartz within the matrix of the mafic rock. The impression is conveyed that the mafic host rocks were either completely soaked in the granite magma or that they underwent a transformation from an amphibolitic parent rock to a granite (granitization?).

This cut exposes an interesting multiple fault zone of a type that is common in the area. It parallels the cut and consists of a number of individual slip surfaces, all highly polished, grooved, and striated. The deep curvature of some of these surfaces and the lack of paralllism of the striations, even where slip surfaces merge on crosscut, indicates the complexity of fault movement.

Stop B (10 minutes, we will stay on bus)
Boston North quadrangle; 26,175E/04,825N;
Highway I-93, where it crosses the valley of Sweetwater Brook

We are crossing the Walden Pond fault. The surface trace of this major fault is marked in places by valleys and trenches (viz. Walden Pond in Lynn Woods; Marblehead Harbor, Mystic Lakes Valley, etc.) Rocks under this valley are known to be deeply decomposed.

From here on our trip will be in the Boston Basin.

Stop C (20 minutes)
Boston North quadrangle; 26,850E/02,100N
Middlesex Fells Reservation, Stoneham

We will see soda andesite of the Boston Basin intruded by one of the Dedham granite plutons and in consequence contact metamorphosed. These rocks were named the Middlesec Fells Volcanics by Bell and Alvord, 1976. We can trace this volcanic horizon 5 km to the northeast where the granite and hornfelsed nature of the rock disappear, and the flanking rhyolite is clearly present. The rock then resembles in every way the soda andesite that we shall see later in the trip in Mattapan Square (Stop G).

Stop D (30 minutes)
Boston North quadrangle; 27,600E/95,600N

This is the old Mystic quarry (LaForge, 1932, pl.11), a 19th century source of freestone blocks used for building walls and house foundations before the age of concrete. The backwall of the quarry today is mostly a NW-striking metadiabase dike of a type that is found widely in the Boston Basin. This dike is cut by two small N-striking lamprophyr dikes, both of which give evidence of occupying earlier faults and one of which is faulted in turn. These latter dikes are probably Triassic in age (Kaye, 1983; LaForge, 1932). They are specifically referred to by LaForge, in the caption of the above cited photograph as containing xenoliths of minerals and rocks brought up from "considerable depths".

In the central part of the backwall of the quarry, the metadiabase dike has been faulted out and in its place there is an excellent exposure of argillite. We will note the broad range of argillite lithologies and depositional structures present in this small thickness of sediment, paying particular attention to lenticular beds of calcareous argillite, thin zones of black argillite, graded bedding, and the presence of the trace-fossil "Spirochete".

Stop E (30 minutes)
Weymouth quadrangle; 38,920E/77,800N

This is the type locality of the Weymouth Formation of Early Cambrian age and the only place where the fossiliferous facies crops out in its unmetamorphosed state. We will see in several small outcrops in the yards of the houses that lie between Gilmore Street and Brewster Road somewhat shaly argillite with flattened fist-size limestone nodules and thin limy beds. The color of the argillite here is red, but in places close by the red changes to gray within a very short distance. Along the shore of Mill Cove, at the foot of Brewster Road, we find that the argillite has changed to a very dark gray flinty rock and that a 0.3 m-thick limestone bed is marblized. This is a good example of the baking effect of the nearby Quincy Granite, whose contact is only about 100 m off-shore here.

The outcrops of Weymouth formation here have an almost E-W strike and steep south dip. We know from excavations, borings, and from other data that the limy fossiliferous argillite is underlain by fine to coarse-grained, reddish to gray sandstone. These beds crop out under the higher ground between Evans Street and Bridge Street (Rte 3A) just to the north. Beneath the sandstones are gray argillites of the type that is widespread in the Boston Basin. Time permitting, we will discuss the problem of the base of the Cambrian.

West, across Fore River, the higher ground on the horizon marks the outcrop of the Quincy Granite pluton. At the foot of the low scarp forming the contact of the pluton with the argillites into which it is intruded — on the low shelf that rims the river — we can see the site of the Haywards quarry, famous for the many fine fossils of the large Middle Cambrian trilobite, Paradoxides harlani, that were found in it in the 19th century. The quarry was situated where we see the large black, shed-like, building at the left(south) end of the General Dynamics Corporation shipyard. The Middle Cambrian argillites (the Brainstree Slate, but they are slate only locally, due to faulting) extend, therefore, under the river as a broad E-W band about 0.3 km in front of us.

* Stop F (30 + 10 minutes)
Boston South quadrangle; 34,340E/85,350N
and 32,650E/83,475N, time permitting

This is the locality of the Squantum "tillite" — a diamictite that is probably of turbidity-current origin. The rock occurs as a number of elongated, lenticular masses completely embedded in a thick sequence of tuffaceous fine-grained sediment with conspicuous graded bedding. The diamictite is a massive sediment similar to many glacial tills in grain-size distribution. It differs from till (as, for example, the Pleistocene tills of the Boston area), however, by the absence of glacial striations on the surfaces of pebbles of fine-grained softer rocks and by the lack of pebbles with "soled" shapes, a characteristic of many glacial tills.

With one exception, clasts of the diamictite are of the same rocks types as those that make up the conglomerates of the Boston Basin — that is, rhyolitic and andesitic volcanics, quartzite, and Dedham-type granite. A feature of importance is that pebbles of spilite are in greater relative abundance than in conglomerate and are locally conspicuous by their very large size. In addition, there are sparse angular blocks up to 0.6 m across, of argillite of the Boson Basin type. The latter have not been noted in the conglomerate.

If we make the most probable structural interpretation, the Squantum beds are the stratigraphic equivalent of the Brighton spilite horizon (Brighton Melaphyre). Structurally, the Squantum beds are folded into a syncline whose axis strikes ENE and passes down the center of the island, and is probably the same fold as the Neponset River syncline to the west.

The most logical explanation for the diamictite is that it represents subaqueous sliding, or slumping, of slopes cut in gravel, sand and clay, and that this debris was carried along the bottom as a number of turbidity currents These slumps were probably triggered by explosive volcanic eruptions, the same eruptions that produced the large spilite fragments. The blocks of argillite probably were carried as unconsolidated or semiconsolidated clay,for it is known that turbidity currents have the capablity of transportating particles with a minimum of abrasion. On the other hand, if these deposits were true tillites, then we would have to conclude that Boston Basin clays had already hardened to argillite at the time of the glaciation that was responsible for the deposit — an unlikely situation.

Stop G (50 minutes)
Boston South quadrangle 27,510E/82,000N
Mattapan Square and nearby Tileston School quarry

The close relationship of rhyolite, soda andesite, and sedimentary interbeds are well-exposed here. We will walk from the area on the west side of Blue Hills Avenue to the old quarry on the east side.

On the west side of Mattapan Square we can see thin-bedded argillite and volcanic conglomerate overlying and in depositional contact with a stratified sequence of rhyolite that includes coarse breccias and well-stratified welded ash. Above the sedimentary rocks are flows of black keratophyric lavas. Across Blue Hills Avenue, on the east side of Mattapan Square, there are two curious large elliptical bodies of well stratified welded tuff embedded in mafic flow rocks. The long axes of the two ellipses are about 250 m across each. The fine stratification of these structures conforms to the margin of the structure but is discordant with the flow layering of the surrounding andesites. Taking all factors into account, the best explanation for the twin ellipticla bodies of rhyolite is that they are fragments of a large crater spine that originally emanated from the volcanic crater that produced the rhyolite cropping out on the other side of Blue Hills Avenue, and that this spine had collapsed as two pieces, falling into the molten andesite that flowed about its base.

The rhyolite seen on this stop represents the second rhyolitic eruptive interval (Rhyolite II). Rhyolite I crops out in the broad area extending on both sides of Cummins Highway, and which is confined for the most part between the two sets of Conrail tracks lying west of Mattapan Square (Boston South topographic sheet, Kaye, 1980).

Stop H (30 minutes)
Newton quadrangle; 23,470E/82,325N
We will examine the tranformation of Dedham granite to rhyolite.

The upland area of Stony Brook Reservation, on both sides of Washington Streeet, shows the gradation in grainsize between aphanitic rhyolite and medium-grained granite. Although there are places where rhyolite, clearly of volcanic origin, can be seen unconformably overlying granite, there are others where there is obviously a textural gradation between the end members. The typical transformation of one rock to the other represents a reduction in average grain size by a factor of about 50. The fine-grained phase of the granite is aplitic in appearance. The rhyolite represents much the same mineralogy but a reduction of grain size by about an order of magnitude.

We are probably seeing in much of the fine-grained rock the chilled margins of a rising granite cupola. This is essentially hypabyssal rhyolite rather than extrusive rhyolite. In some places, however, the aphanitic rock is probably extrusive. The field relationships of the different grain sizes are difficult to determine because this is a very much faulted terrane, as a visit to the nearby West Roxbury Crushed Stone quarry (22,000E/81,500N) would show (if we had the time).

End of field trip

A PRECAMBRIAN CONTINENT MARGIN SEQUENCE (SLOPE DEPOSITS AND OLISTOSTROMES): BOSTON NORTH QUADRANGLE, MA

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Introduction

The oldest rocks of the Boston Platform of eastern Massachusetts and western Rhode Island are present in a southwesterly trending belt between the Boston and Narragansett Basins to the east and southeast and the Lake Char and Bloody Bluff Faults to the west and northwest (Figure 1). The metasedimentary and metavolcanic rocks within this belt in northeastern Massachusetts comprise the Weston Group (Skehan and Murray, 1980). On this field trip we will examine a small portion of this metasedimentary sequence that lies within the Boston North quadrangle. These metasedimentary rocks are the only ones I have found in the Weston Group that preserve primary sedimentary structures and elements of original textures.

Stratigraphy and Age

Precambrian stratified rocks in the Boston North quadrangle were assigned to the Westboro Formation by Bell and Alvord (1976). They recognized that much of the unit was not a quartzite as it is near the town of Westboro at the type locality, hence they modified the name and generalized the description to include same slates, schists, and gneisses in the quadrangles between Marlboro and Boston North (Figure 1). Nelson (1974) refined the stratigraphy of the Westboro in the Natick and Framingham quadrangles and correlated the middle part of the unit with the Plainfield Formation of Connecticut and the Quinnville Quartzite of the Blackstone Group in Rhode Island. Most of these correlations are rather tenuous in that they involve amphibolite or higher grade rocks in structurally isolated and distantly separated blocks. I favor a separate unit of formational rank for the metasedimentary rocks within the Boston North quadrangle as they are lithically distinct and I accept the tentative correlation of these rocks with some portion of the Blackstone Group farther to the south.

Figure 2 shows the outcrop area of metasedimentary rocks in the Boston North quadrangle (as delimited by Kaye, 1980) and the locations of measured sections in structural blocks labeled I-IV. The thickest single section is about 600 m and Bell and Alvord (1976) report a total thickness of 1,100 m. Bedding generally dips steeply to the north but dip reversal and variability is present. Structures useful for determining bed tops

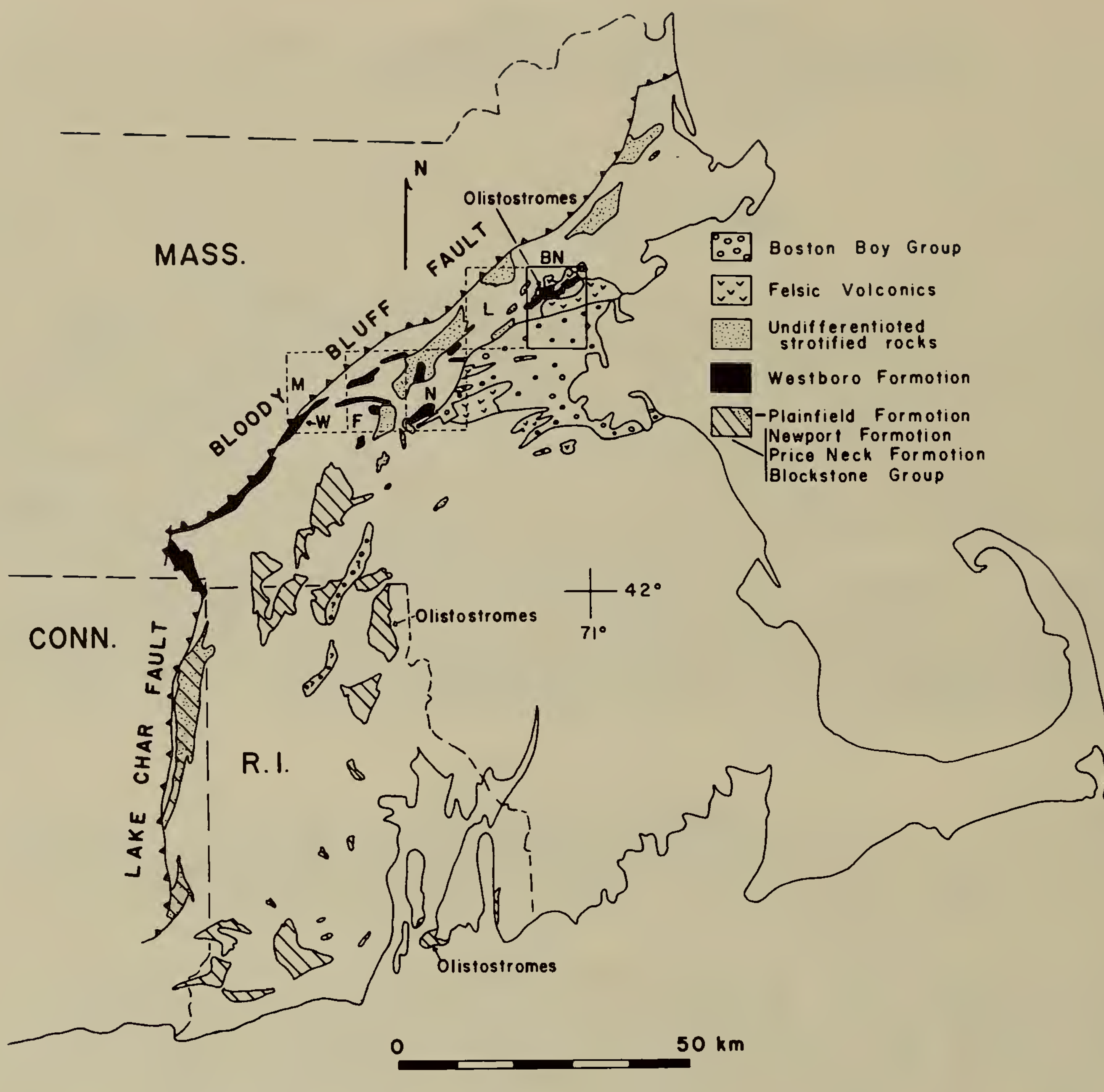


Figure 1. Map of Boston Platform showing distribution of stratified Precambrian rocks. Quadrangles shown; BN = Boston North, L = Lexington, N = Natick, F = Framingham, M = Marlboro, W is type locality of Westboro Fm. Areas where olistostromes have been reported are shown.

Figure 2. Outcrop area of Precambrian metasedimentary rocks (Westboro Fm., stippled) in Boston North quadrangle. Map boundaries are from Kaye, 1980. This outcrop belt is shown in black in Fig. 1. Field trip stops are numbers in circles. Structural blocks labeled I-IV. Heavy black bars are measured sections shown in Fig.3. Heavily stippled pattern represents massive or thickly bedded metaquartzarenite.

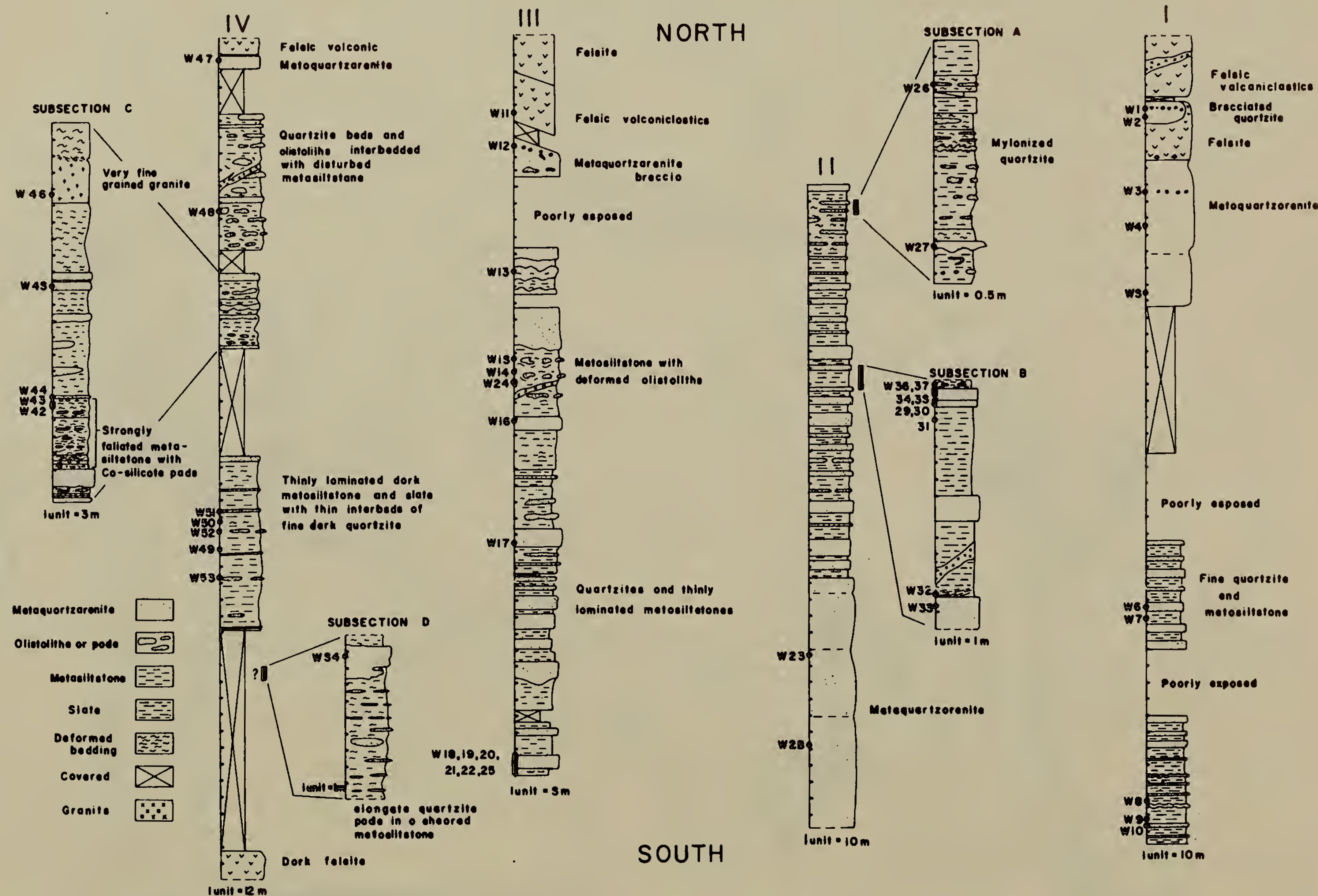


Figure 3. Composite and generalized stratigraphic sections of metasedimentary rocks in Boston North quadrangle. Note the variable scales given at the base of each section. Sections have been presented from north (top) to south (bottom) in the event that undetected inversion or fold duplication of section exists.

and bottoms are very rare, but in two cases an erosional truncation and a possible graded bed indicate that the sections in blocks II and III are probably right-side up. Bell and Alvord (1976) divided the Westboro in Boston North into an upper quartzite and geniss member and a lower member of somewhat less than half massive quartzite. These members seem to be present in blocks II and III (Figures 2 and 3) but the two member scheme is unworkable for the entire outcrop area. The southern or lower contact of the stratigraphic sequence is not exposed; however, based on regional relationships, it is probably an intrusive or fault contact with the Dedham Granite and a fault contact with felsic volcanic rocks. The upper contact has been described as gradational into overlying felsites and volcanic rocks of the Middlesex Fells Volcanic Complex (Bell and Alvord, 1976), but evidence will be presented at stops 2 and 5 that implies a profound structural and/or unconformable break at the top of the section. Correlations between stratigraphic sections (Figure 3) are not convincing. In a few instances it is possible to trace massive quartzite beds over distances of about 1 km. A very important observation made by Bell and Alvord (1976) is that many of the pure metaquartzarenite beds form lenticular bodies up to 60 m thick and 1.5 km in length. At the scale of a single outcrop the stratigraphy is often highly disturbed and discontinuous.

The age of the sedimentary sequence is imperfectly known, but U-Pb ages of detrital zircons from the quartzarenite beds establish a maximum of about 1500 Ma (Olszewski, 1980) and Rb-Sr whole rock ages from the cross-cutting Dedham Granite provide a minimum of about 600 Ma (Kovach and others, 1977). Recently, Zartman and Naylor (1984) suggest that an age of about 620 to 630 Ma is more appropriate for the Dedham.

Facies Characteristics

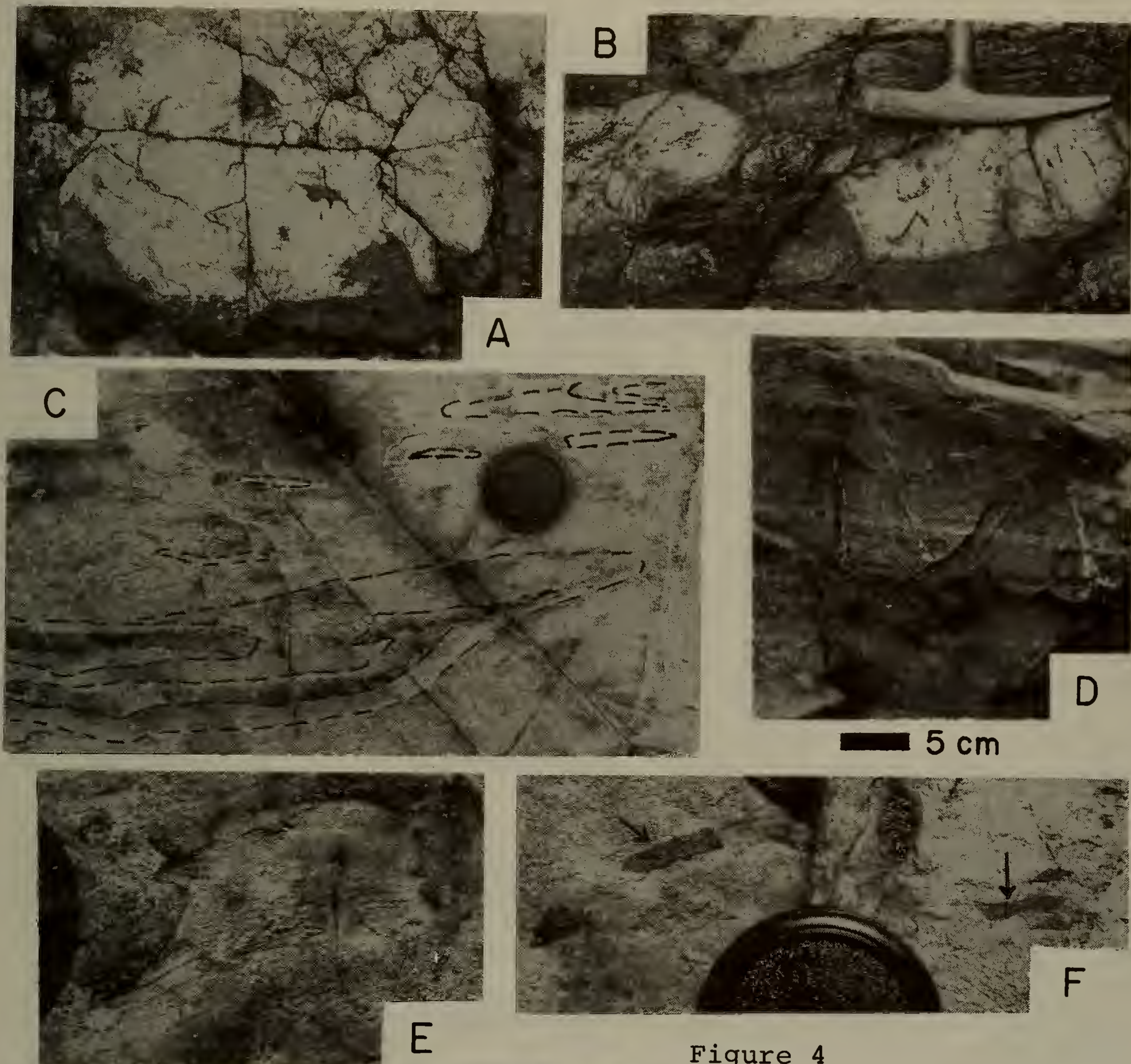
Characteristics of the 5 basic lithofacies that comprise stratigraphic sections in the Boston North quadrangle are given in Table 1.

Sedimentary Structures

Soft sediment deformation structures and intraclasts are common in some outcrops, (Figure 4) especially where beds of metaquartzarenite and metasiltstone are intercalated. Some of the soft sediment folds resemble tectonic structures and it is important to distinguish between the two (Figure 4C). Because of their complexity, often in association with thinly laminated and undisturbed strata, I interpret most of the small scale folds to be seen on this trip as penecontemporaneous with slumping or debris flows. Some units, up to several tens of meters thick, have a very chaotic appearance. Irregular masses or blocks of metaquartzarenite with very intricate margins occur in a fine,

Table 1. Characteristics of Lithofacies

	Lithology	Bedding	Sedimentary Structures	Petrography
1.	Metaquartzarenite	0.5 - 50 m	Intraclasts - little internal structure	70-95% fine to coarse quartz grains; interpenetrating; muscovite and hornblende.
2.	Laminated dark meta-arenite	1mm -.50cm	Delicate parallel laminations; very rare graded beds	50-80% very fine to fine grained quartz; 20-50% muscovite, chlorite, hornblende, actinolite;
3.	Metasiltstone	0.5 - 10mm	Very thinly laminated	Quartz silt and thin, very fine quartz stringers with mixture of epidote, chlorite, amphibole and muscovite.
4.	Slate or Argillite	0.5 - 5mm	Very thinly laminated	Mixture of muscovite, chlorite, amphibole, with some quartz silt.
5.	Olistostromes (?)	Chaotic	Irregular pods, masses, and blocks of metaquartzarenite or meta-arenite usually in a highly deformed metasiltstone; Siltstones are occasionally mixed with quartz-rich beds; intraclasts, folds.	Depends on portion sampled; coarser quartz grains sometimes admixed with finer clastics.



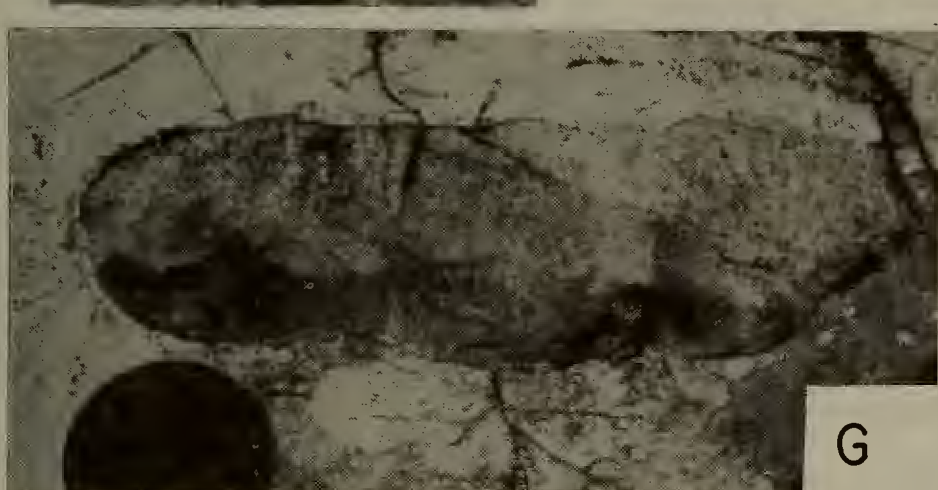


Figure 4
Primary or soft sediment structures: A. metaquartzarenite mass in fine dark quartzarenite; B. metaquartzarenite blocks in metasiltstone; C. soft sediment folds and metasiltstone intraclasts; D. thinly laminated metasiltstone, scale bar is 5 cm; E. laminated intraclast in metaquartzarenite bed, note ragged and distorted edges, edge of lens cap on left for scale; F. metasiltstone intraclasts in metaquartzarenite bed; G. rounded fine metaquartzarenite intraclast in metaquartzarenite bed.

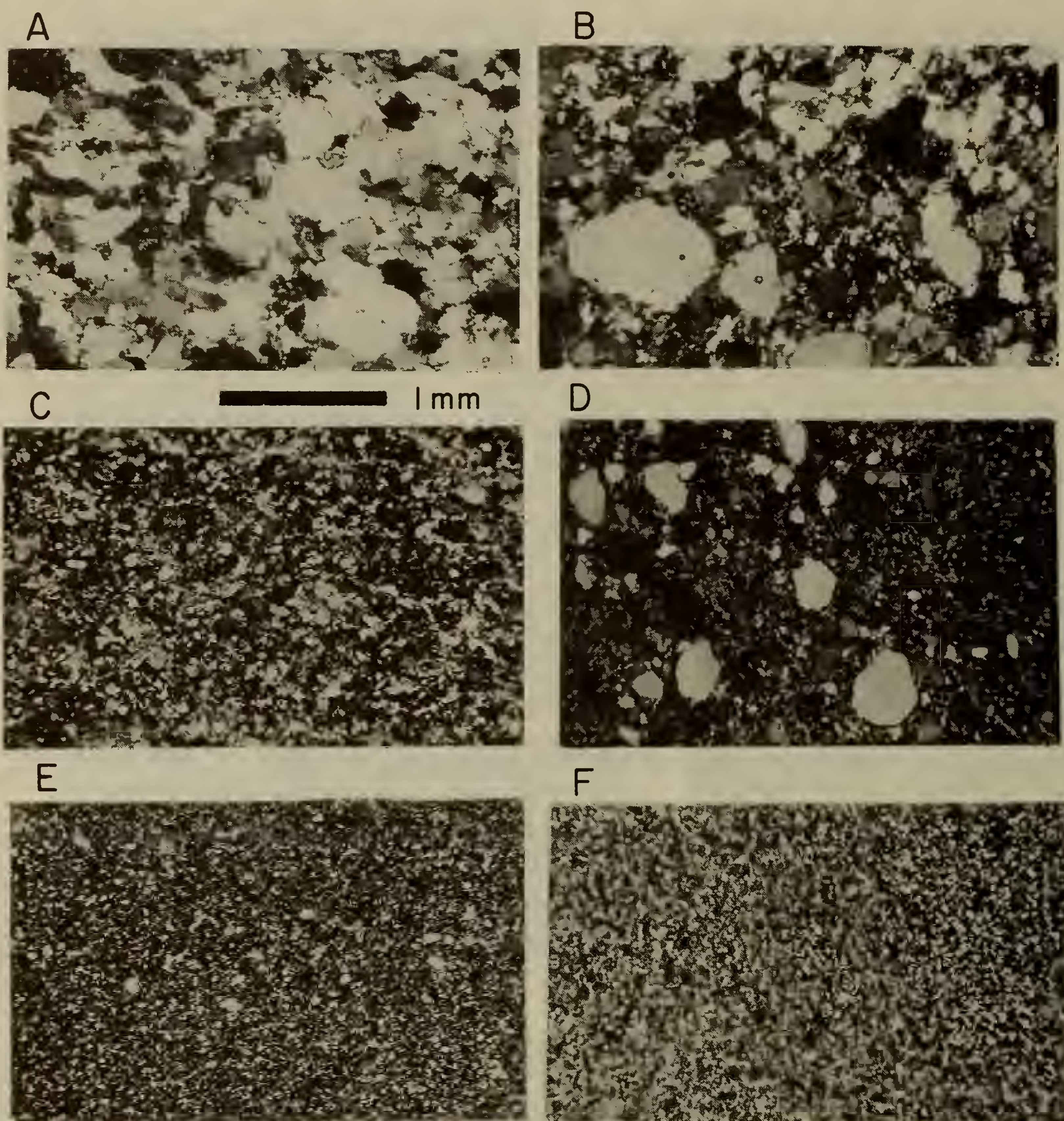


Figure 5. Photomicrographs of typical lithologies. Scale bar below photos applies to all photos. All with doubly polarized light. A. well sorted medium grained metaquartzarenite, grains show moderate interpenetration; B. moderately sorted fine to medium grained metaquartzarenite, grains only sightly interpentrating; C. very fine grained metaquartzarenite; D. diffuse intraclast with grains "escaping" into metasiltstone matrix; E. metasiltstone with thin stringer of very fine quartz grains across middle of photo; E. metasiltstone.

dark metaquartzarenite or metasiltstone matrix (Figure 4). In some cases the blocks have a texture nearly identical with the matrix except that the block, blob, or patch has been more thoroughly cemented and in other cases the quartzarenite is clearly distinguishable from the matrix that surrounds it. Possible load and injection structures are present along the bases of some metaquartzarenite beds.

Petrography

Petrographic characteristics are summarized in Table 1. The metaquartzarenite protolith is a mature to supermature sandstone. Quartzitic textures of an interlocking and interpenetrating mosaic of quartz grains with sparse muscovite are characteristic of thicker metaquartzarenite beds (Figure 5A). Occasionally a portion of the original detrital texture is preserved (Figure 5B). Quartzarenite intraclast boundaries are often diffuse and quartz grains seem to have been released from the intraclast or are floating in the surrounding matrix (Figure 5D). This is good evidence that the intraclasts were only weakly consolidated when they were transported. Metasiltstones and fine metaquartzarenites are very homogeneous mixtures of quartz grains with variable amounts of chlorite, muscovite, actinolite, epidote, and hornblende. Bedding or any layering is very obscure in thin section (Figure 5C, F). Thin stringers of fine quartz grains are found in metasiltstones (Figure 5E) but grading is extremely rare.

A quartz-feldspar-lithic fragment plot is given in Figure 6. All fine to coarse grained meta-arenites studied in thin section are included in the plot. Most of these rocks contain from 70-90% quartz, 5-20% potassium feldspar, 5% plagioclase, and 5-15% muscovite and amphibole. The quartz content may have been relatively higher in the protolith as some quantity of the feldspar is probably secondary. Chert and arenite sized lithic fragments are extremely rare and presumably were not present in abundance in the sandstone protolith.

Structure and Metamorphism

The metamorphic grade and degree of ductile deformation within the Weston Group increase to the west. In the Boston North area chlorite grade metamorphism predominates and locally higher temperature assemblages are present adjacent to plutons. A weak foliation or cleavage parallel to bedding is present. Other cleavages are absent or poorly developed. No definitely tectonic folds of outcrop scale were observed but larger scale folding cannot be ruled out. Rocks with a strongly tectonic fabric are present at stops 6, 7, and 8 and are discussed in the stop descriptions. Faults, fractures, and joints are evidence of extensive brittle deformation. Most outcrops show evidence of small offsets on faults.

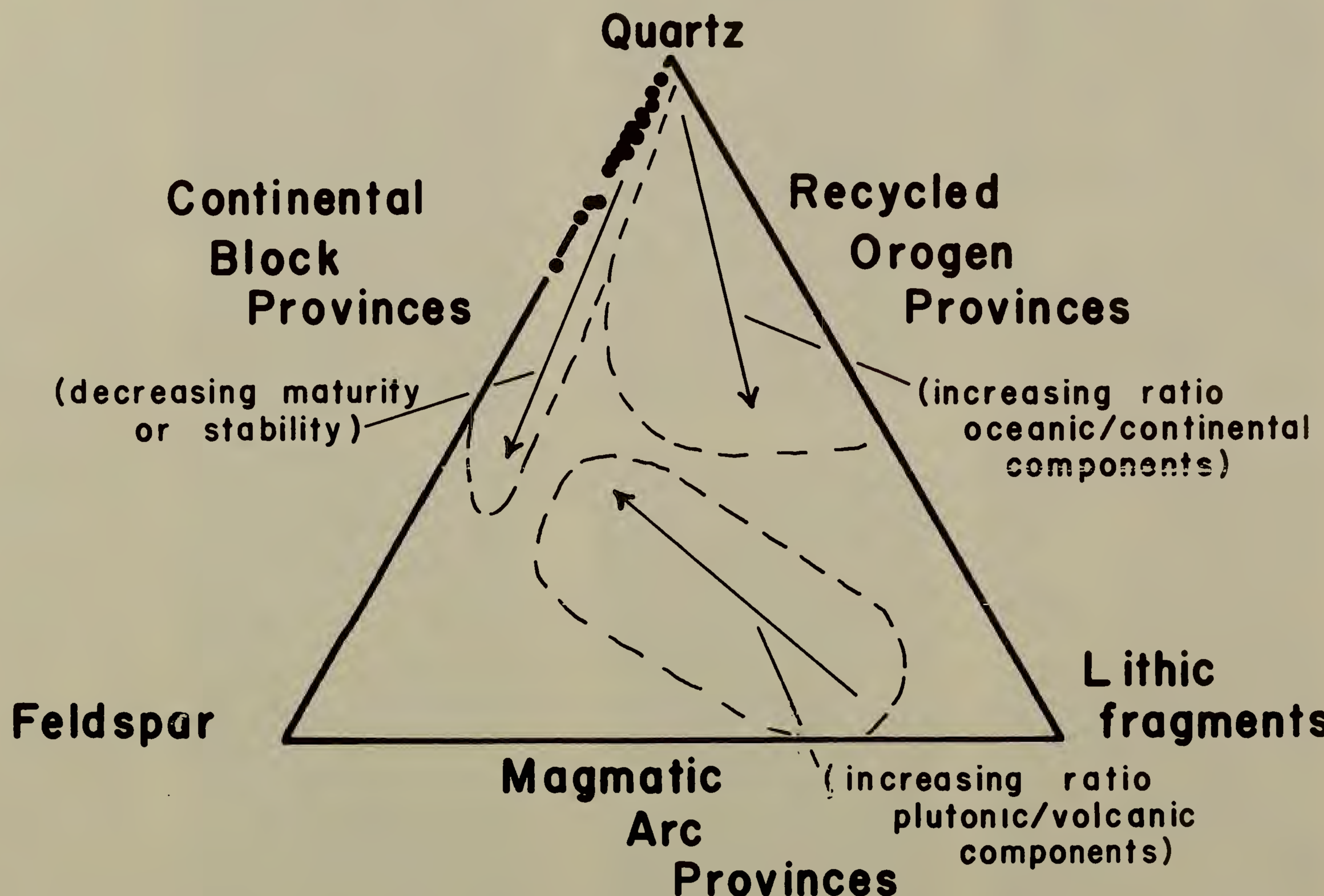


Figure 6 Quartz, feldspar, lithic fragment plot of 17 fine to to coarse grained, metaquartzarenites. Fields for sandstone composition from Dickinson and Suczek (1979).

Depositional Environment and Regional Significance

The mature character of the sediment (Figure 6) is indicative of multicyclic sands associated with a cratonal or stable platform setting (Dickinson and Suczek, 1979). Olszewski (1980) demonstrated the need for a cratonal source for the 1500 Ma detrital zircons from the Westboro. He suggested the West African Shield would be the closest source if Africa and North America are arranged in a standard pre-drift configuration. More recently O'Brien and Wardle (1983) concluded that the Avalon zone in Newfoundland bears striking similarities to Pan-African belts where supracrustal and cratonal margin sequences not unlike the Weston and Blackstone Groups were involved in 800-1000 Ma rifting. Dreier and Mosher (1981) proposed rifting, syn-depositional mixing, and development of olistostromes as an explanation for the chaotic character of the Blackstone Group and Rast and Skehan (1981) have described rocks of possibly similar age and origin at Newport, Rhode Island. The metasedimentary rocks of the Westboro Formation thus appear to be a part of this disrupted cratonal sequence widely represented on the Boston Platform.

Deposition on a passive basin margin or upper slope provides a good explanation of the stratigraphy and sedimentary structures observed in the Boston North metasedimentary sequence (Keith and Friedman, 1977). The soft sediment folds, intraclasts, quartzarenite blocks and irregular masses resulted from a slump or debris flow mechanism. The large lenticular metaquartzarenite masses may represent olistoliths that moved downslope as coherent but uncemented blocks (Hsu, 1974). There are no exotic (ie. chert or basalt) clasts in the chaotic beds interpreted as possible olistostromes. Most of these beds seem to have been distorted but not transported significantly. Massive structureless sand beds with isolated intraclasts may have originated as grainflow deposits and the thin parallel-laminated very fine sandstones and siltstones as suspension deposits or very low energy turbidites (Keith and Friedman, 1977).

Acknowledgments

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Road Log

Mileage

0.0 Inn at Danvers parking lot

0.05 Exit onto Dayton St. on north side of Inn: 180° right on Dayton onto Armory Road

0.1 Left at sign for "Boston-Newburyport Rte 95"

0.2 Right at intersection; follow sign for "Rtes. 95 and 1 Boston and Topsfield"

0.3 Left at intersection; follow sign for "95 south and Rte. 128 Burlington and Gloucester"; enter onto Rt. 95; Continue south 95/Rt 1

4.0 Cross over Rt. 128; continue south

6.8 Right at Lynn Fells Pkwy exit; Melrose and Stoneham

7.0 Left for Rt. 1 north to Lynnfield; cross over Rt. 1

7.3 Merge onto Rt. 1 N.

7.4 Right turn into parking lot immediately beyond Exxon Station. Lock cars and walk back south along Rt. 1 up onramp to high cut along Saugus River.

Stop 1. Across the Saugus River you can see (from left to right) the Dedham Granite, a thick dolerite dike, a large metaquartzarenite xenolith or roof pendant (?), more granite, a smaller xenolith followed finally by strange looking dolerite dike, and more granite. The Dedham is clearly younger than the metasedimentary blocks it encloses. Rb-Sr ages of the Dedham at this locality are about 600 Ma (Kovach and others, 1977). Minor folding and injection parallel to foliation along margins of large xenolith are evident. Lithologies and contacts can be closely examined along west bank of river.

7.5 Exit parking lot and turn north (right) onto Rt. 1.

8.0 Go under overpass and immediately bear right onto Rt. 129 west. Cross over Rt. 1

8.2 Bear right onto Rt. 1 south

8.8 Take Lynn Fells Pkwy exit (east) toward Stoneham

9.1 Right turn into Breakheart Reservation

9.5 Entrance to Breakheart Reservation. Park in lot across from maintenance building. Walk north up loop road to left of building about 800 ft. Take left at paved road blocked by steel gate and walk about 200 ft. to dirt road (Ash Path) entering from left. Take left onto Ash Path (at gate) and walk about 0.25 mile along path to only dirt road that enters from left. Go left about 50 ft. down the dirt road and climb outcrop immediately to left in woods to Stop 2.

Stop 2. Many of the metasedimentary outcrops in the Boston North quad. are like this one. Rusty brown metasiltstone and fine meta-arenite with irregular blobs and patches of metaquartzarenite. Many of the blobs are in strings as if they represented disjointed beds. Is this chaotic appearance the result of tectonism or a sedimentary process? I think the latter, but it certainly is debatable. After examining glaciated surface, cross the small ravine about 20 ft. to the east and observe the structures in cross-section and on a bedding surface.

Return to the intersection of the paved road with the loop road. Walk north about 200 ft. and climb steep rock face to the right to the top of the hill to Stop 3.

Stop 3. The sketch map below (Figure 7) shows the contacts and distribution of rock types in this important area and the stratigraphy is summarized in Figure 3. When you are standing on the top of the highest bedrock knob you are on flow banded and agglomeritic crystal-vitric felsic tuffs. Moving toward the edge of the cliff you can find several large pods of Dedham Granite that cut and include the volcanics. Moving to the south you cross from volcanics into a large brecciated quartzite block. The contact between the volcanics and quartzite is not visible. The quartzite block has been engulfed by a presumably intrusive felsite. The two lithologies are very similar in the field. The southern contact of the felsite with the main sequence is exposed. Careful searching will reveal small xenoliths of quartzite in the felsite at the contact. Please do not hammer on contact. The import of this locality is to cast serious doubt on the idea that the sedimentary sequence grades upward without serious complications into the volanics. This is an important point in environmental reconstructions. If the contact were conformable and gradational the supermature metaquartzarenite depositional enviroment and tectonic setting would have to be compatible with a thick section of felsic volcanics. I interprete the contact here as a probable fault along which felsite was intruded. Walk to the south down the hill in the ravine. All the way back to the road you are walking in a massive, apparently unbedded, metaquartzarenite about 90 m thick.

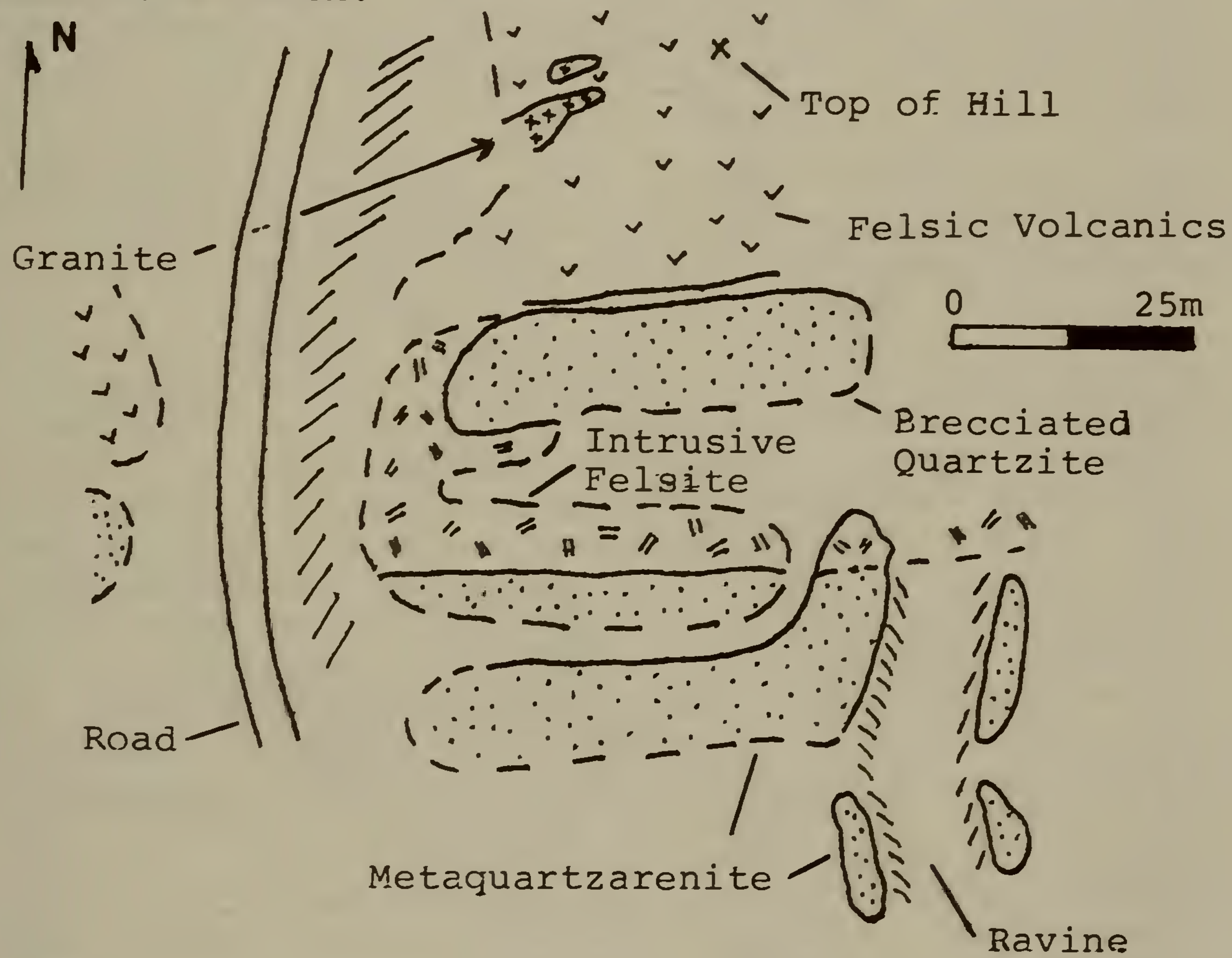


Figure 7
Outcrop sketch map of crest of westernmost hill within southern end of loop road showing contact relationships of quartzites, volcanics, felsites, and granite.

9.8 Leave parking lot and return to Lynn Fells Pkwy. Take right turn onto Pkwy

10.5 Take right turn onto Main St.

10.7 Outcrop is on right; reverse direction and park on opposite side of road across from outcrop

Stop 4. Massive metaquartzarenite. This rock is typical of the thickly bedded mature quartzarenites. A sample of the gray quartzite contained 77% quartz, 12% potassium and plagioclase feldspar and 11% muscovite. Several beds of fine quartzite and granitic intrusions are also present.

Head south on Main St.

10.8 Right turn onto Howard St.

11.4 Right turn onto Windsor St.

11.6 Park on right underneath powerlines. Lock cars. Take right onto small trail into woods just north of powerlines. Outcrop here is massive metaquartzarenite. Follow trail and road under or parallel to powerlines for 0.5 mile to a point where the road ascends a steep hill and makes a sharp 90° turn to left and back right again.

Stop 5. A sketch map of this section is given in Figure 8 and the stratigraphy is summarized in Figure 3. The southerly dipping rocks in these outcrops form a section about 180 m thick. This is the best area to observe the various sedimentary structures illustrated in Figure 4. Please do not hammer on structures. Intraclasts, quartzite masses and blocks, and chaotic horizons are present. Massive metaquartzarenite beds are intercalated with fine metaquartzarenite and metasiltstone strata. The contact of metasedimentary with volcanic rocks cuts very obliquely across the strike of the sedimentary sequence. An irregular basaltic dike or mass has been injected approximately parallel to the contact. All of these features plus the presence of a brecciated quartzite suggest that the contact is a fault. North of the contact are outcrops of lithic-crystal-vitric felsic tuff. This tuff contains flattened "pumice" blocks suggestive of an ash-flow or welded tuff. If time permits you can follow the road to the crest of the hill where a small body of diorite intrudes a dark, fine grained, felsite (?).

Reverse direction and walk about 0.3 of a mile back along powerline to a point where a road climbs steeply up a hill to the left. Climb to the top of the hill near the base of the powerline tower.

Stop 6. The highest knob on the top of the hill exposes a blastomylonite zone at least 10 m wide. This is the only such zone thus far discovered. On the weathered outcrop surface parallel to sub-parallel laminae of quartz give the rocks an irregular laminated fabric. Greatly elongated quartz grains are visible on some handspecimens. Figure 9 illustrates textures of 2 mylonite samples from this outcrop. The localized strain may be due to shearing associated with a fault mapped by Kaye (1980). The fault is approximately parallel to the powerline and no more than 100 m west of this locality.

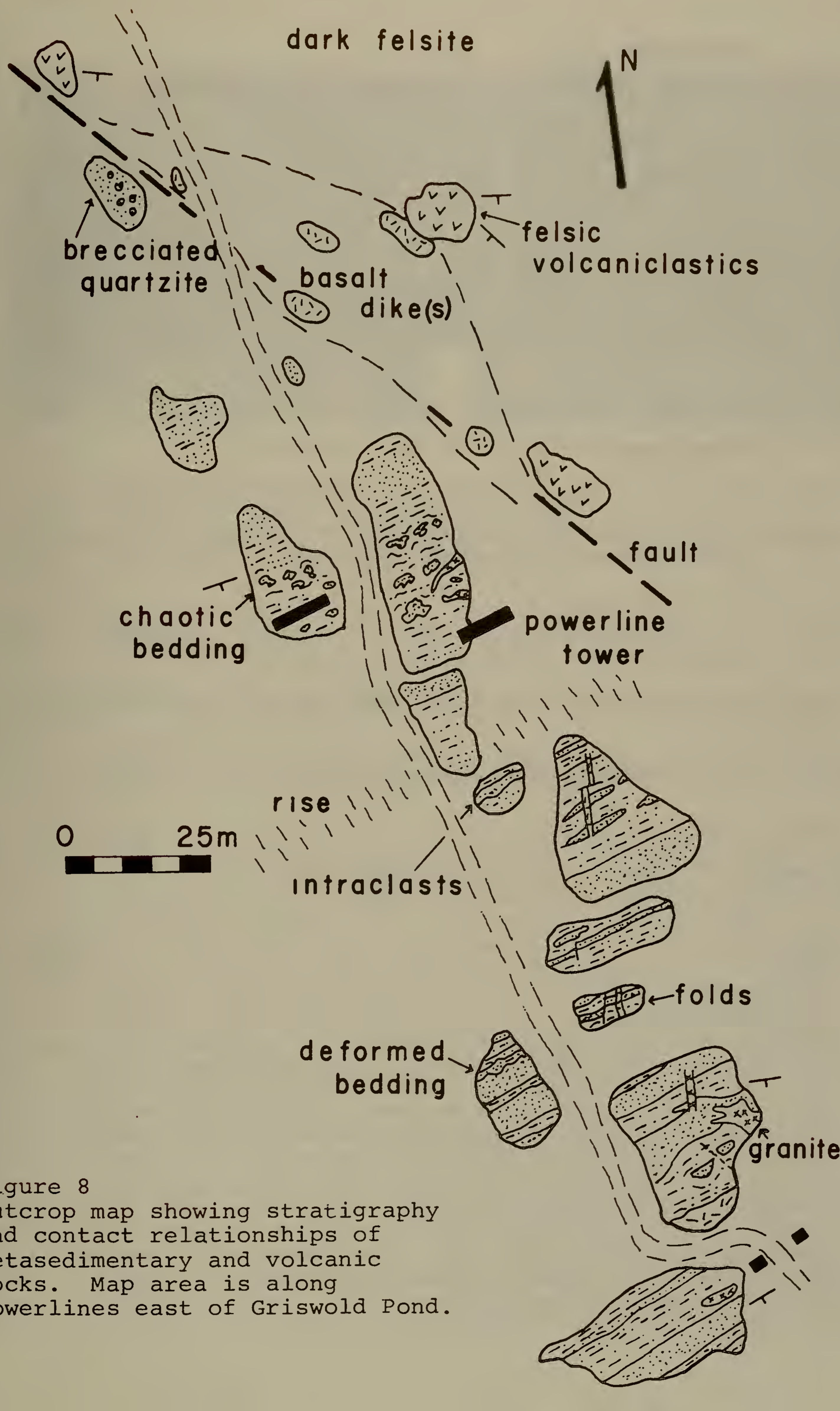


Figure 8
Outcrop map showing stratigraphy and contact relationships of metasedimentary and volcanic rocks. Map area is along powerlines east of Griswold Pond.

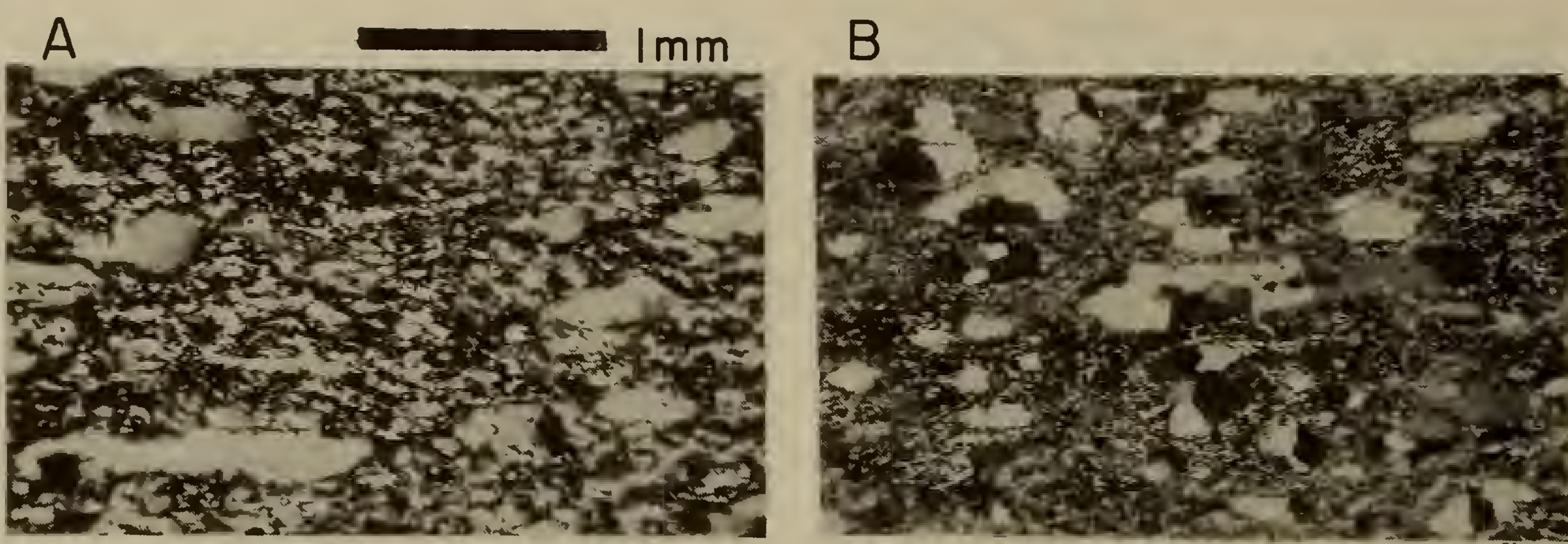


Figure 9
Photomicrographs of blastomylonites at stop 6. Scale bar at bottom applies to both photos. a. meta-quartzarenite, B. poorly sorted meta-arenite.

Walk down hill and follow trail south back to cars.

11.8 Reverse direction and return to intersection with Howard St. Right turn on Howard St.

12.4 Bear left onto Elm St.

12.6 Rt. onto Lynn Fells Pkwy.

13.3 Go under railroad bridge and immediately turn left onto Vinton St.

13.5 Stop sign; continue straight on Vinton St.

13.7 Right turn onto Maple St.

13.75 Stop 7 is steep outcrop to right. Turn left onto Cleveland Street and park on right.

Stop 7. This is a difficult stop to visit with a large group as the outcrop is heavily vegetated and very steep. The entire cliff is on private property and you should obtain permission before trespassing. Hammering is discouraged. Depending on the size of the group we may or may not study this outcrop, but I encourage those interested to come back in smaller numbers especially when the leaves are off the trees. Lenticular masses and blocks of highly recrystallized metaquartzarenite are found all along the cliff. The quartzite pods are boudin-like although randomly distributed throughout a foliated metasilstone (Figure 10A). The rock has a clearly tectonic fabric and I interpret it as a tectonized olistostrome. In thin section the quartzites have a crystalloblastic texture produced by extreme recrystallization (Figure 10B).

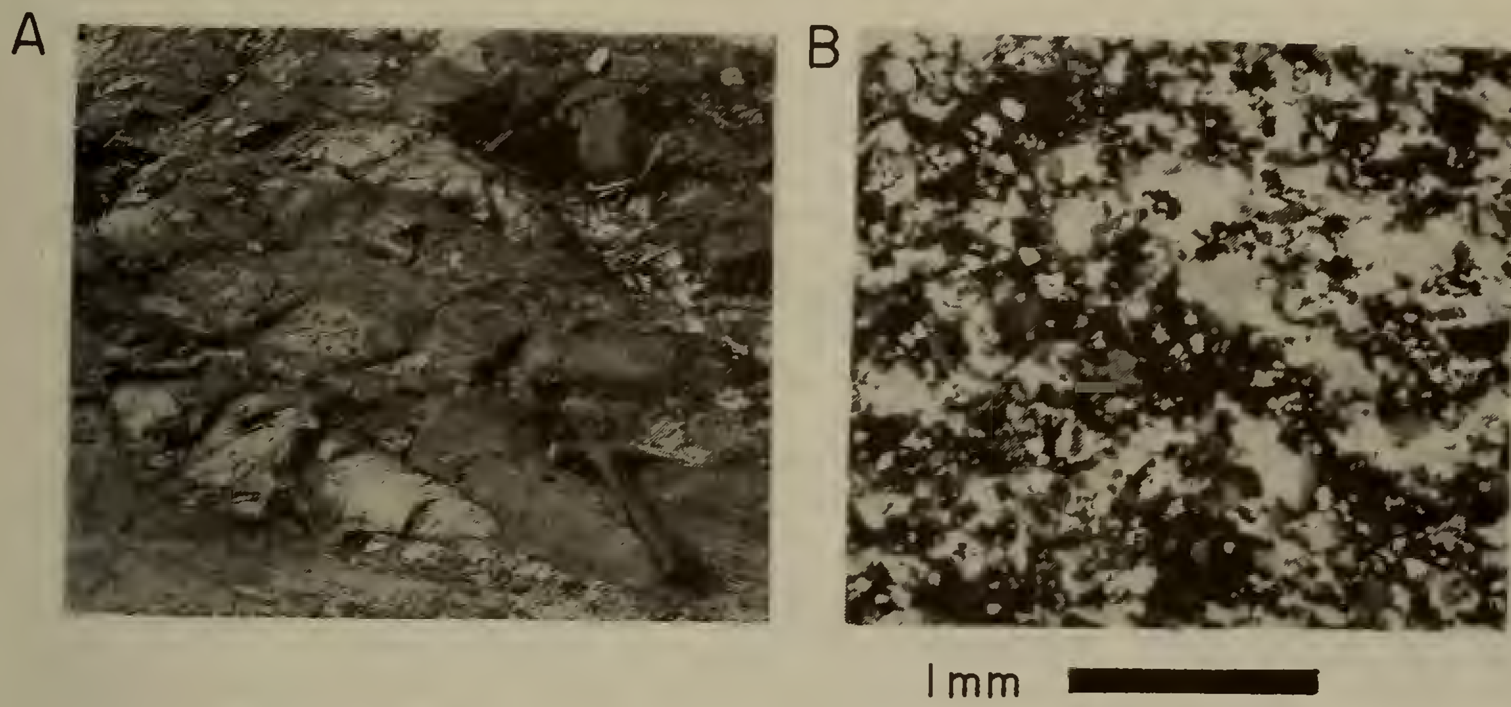


Figure 10
A. Outcrop at stop 7 showing elongated metaquartz-arenite pods; B. photomicrograph of metaquartzarenite from pod, note crystalloblastic texture.

14.0 Continue straight (south) on Cleveland St. to intersection with Wyoming Ave; right turn on Wyoming Ave.

14.5 Cross Lynn Fells Pkwy; continue straight on Pond St.

14.8 Right turn and continue on Pond St., outcrops are on right but we will continue up to parking areas at Stone Zoo.

15.2 Right into large unpaved parking lot (or left into zoo lot if gate is closed). Park in the southern most area of either lot and walk back down Pond St. to beginning of series of road cuts on east side of Pond Street. Cross carefully as traffic moves quickly along road.

Stop 8. The beds in this outcrop are illustrated in Figure 3 as subsection C. The metasedimentary rocks at this locality are strongly sheared and injected in lit par lit fashion with intrusive felsites. Some of the beds contain pods of calc-silicates, tremolite and quartz, in a strongly foliated metasiltstone. Some of the intrusive rocks are very similar to quartzites in the field and one must look carefully to distinguish between the two. About 0.2 mile up the road there is a cut in the side of a small hill. The rock is rather impressive in that it contains abundant xenoliths of quartzite and other lithologies in a gray felsite. This felsite is probably an offshoot of the Peabody or related granites exposed nearby. Note how difficult it is to tell what is xenolith and what is matrix. Walk back to unpaved lot east of Pond St.

Climb to the crest of Whip Hill, to the east of the unpaved lot, by following the dirt road that parallels the southern edge of the unpaved lot. A small trail leads from parking lot to the dirt road. Do not follow the road up the steep hill but bear left. Walk about 0.25 miles (swamp on left) and take right turn at intersection. Walk about 300 ft. and take road to left. After 100ft take road to left for about 50 ft. and climb right through woods to top of Whip Hill. There are several excellent exposures of the chaotic units examined earlier in the day. One metaquartzarenite block, about 5 m by 2m, is completely surrounded by chaotic metasiltstone with smaller irregular quartzite masses. Follow trails back to parking lot.

Exit from parking lot and head south on Pond St.

15.5 Turn left and continue on Pond St.

15.6 Turn left into small roadside parking area with steel gate. Lock cars. Take path labeled Virginia Wood to stone bridge over small creek. Follow trail to right (just before bridge). Trail climbs for about 200 ft. to crest of low ridge. Follow ridge left through woods to outcrop along stream.

Stop 9. This small natural outcrop is characteristic of the fine clastics that are usually very poorly exposed. Metasiltstone and fine dark metaquartzarenites are present along stream. A fault marked by a breccia zone has disjointed and folded a felsite (?) dike. Metaquartzite occurs at top of ridge.

Backtrack to cars; continue straight on Pond St. to intersection with Lynn Fells Pkwy.

15.7 Left onto Lynn Fells Pkwy. Continue on Pkwy for about 4.5 miles to intersection with Rt. 1. Cross over Rt. 1 and follow signs for Rt. 1 north. Continue north on Rt. 1 and/or I-95 to Inn at Danvers. End of trip.

DUCTILE AND BRITTLE STRUCTURES WITHIN THE RYE FORMATION OF SOUTHERN COASTAL MAINE AND NEW HAMPSHIRE

Mark T. Swanson, University of Southern Maine
John A. Carrigan, University of New Hampshire

INTRODUCTION:

The Late Precambrian (?) Rye Formation of Southern Maine and New Hampshire was originally named and described by Katz (1917) and Wandke (1922). The formation was further subdivided by Billings (1956) into an upper metavolcanic and a lower metasedimentary unit (Figure 1). Detailed mapping was carried out by Hussey (1962) and Novotny (1969) to determine the extent and structure of the Rye Formation in southern Maine and southeastern New Hampshire, respectively. These studies verified the overall structural configuration of the Rye Formation as a SW plunging antiform overturned to the SE. Mapping at Gerrish Island by Hussey & Pankiwskyj (1976) located minor NE-plunging antiformal structures suggesting a doubly-plunging antiformal structure for the Rye Anticline. Continued detailed mapping by Hussey (1980) has led to a major reinterpretation of the Rye Formation as a variably sheared and injected metasedimentary sequence. Studies by Carrigan (1984a;b) have resulted in a re-evaluation of the Rye Formation in terms of poly-phase metamorphic events and in the along-strike correlation of two major ductile shear zones and several zones of distinctive lithologic units from Gerrish Island, Maine to Newcastle Island, New Hampshire. The basic lithologic and structural complexities within the Rye Formation have also been superimposed by a mutually complex array of brittle structures that includes brittle faulting and fracturing of Late Paleozoic to Mesozoic age (Swanson, 1982). The Mesozoic structures were coupled with the emplacement of dikes and intrusions related to the development of a rifted continental margin. It is possible, therefore, to examine, within a single formation, ductile structures related to Precambrian-Paleozoic compressional tectonics and brittle structures related to Mesozoic extensional tectonics in the geologic evolution of eastern North America.

LITHOLOGY:

The original sedimentary sequence is hypothesized (Hussey, 1980) to be a series of sandstones, siltstones, carbonaceous shales and limestones. These are now preserved as metamorphic equivalents of the pre-existing lithologies where they have

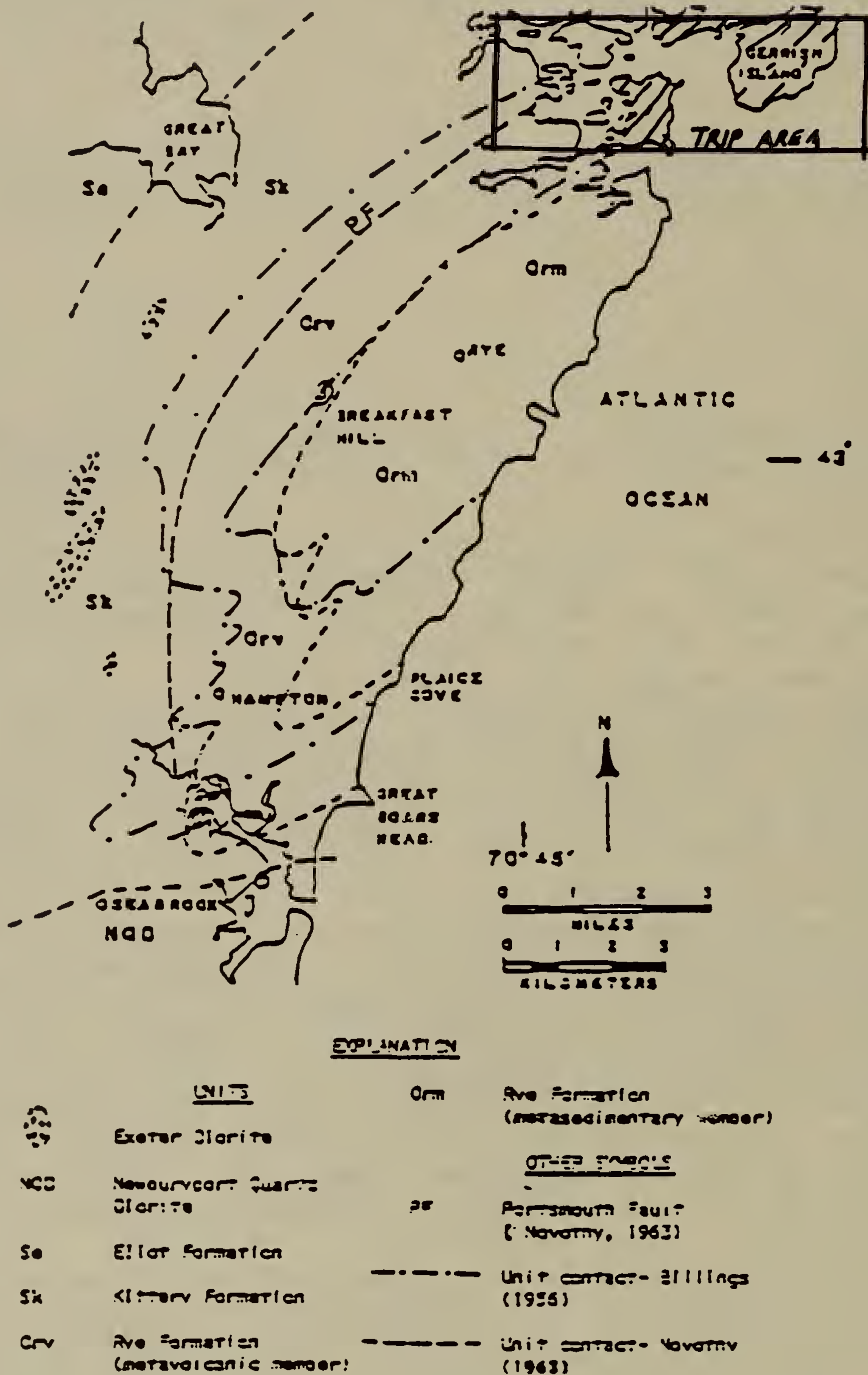


Figure 1. Generalized Geologic Map of southeastern New Hampshire (after Novotny, 1969 and Billings, 1956)

survived a phase of felsic injection and a later phase of mylonitization. The resulting rock types within the Rye Formation consist of variably-sheared calcareous quartzite, metapelite, amphibolite, calc-silicate-bearing amphibolite, graphitic-sulfidic pelitic schists and impure marble, representing this preserved metasedimentary sequence (Figure 2). The injection of quartzo-feldspathic materials as dominantly concordant sheets and a phase of mylonitization during deformation has resulted in the formation of numerous blastomylonitic schists granitic augen gneisses and ultramylonites from these injected metasedimentary lithologies.

STRUCTURE:

The outcrop pattern and minor fold structures found within the Rye Formation indicate a NE-trending, doubly-plunging antiformal structure (Novotny, 1969; Hussey & Pankiwskyj, 1976). The Rye Formation has been multiply-deformed with the last fold-phase (Carrigan, 1984a; Rickerich, 1983) being synchronous with the NE-trending upright folds in the adjoining Kittery Formation. It is this last NE-trending, tight to isoclinal, fold phase that is responsible for the present-day outcrop pattern and overall regional structural configuration. Both gently- and steeply-plunging isoclinal and asymmetric minor fold structures can be found indicating several phases of folding in the structural evolution of these distinctive lithologies. The deformation and metamorhism of the Rye Formation must also predate the Late Ordivician Newburyport Quartz Diorite suggesting an Early Ordivician or possibly Precambrian age (Zartman & Naylor, 1984; Carrigan, 1984a).

Two phases of ductile faulting have been recognized (Hussey, 1980; Carrigan, 1984a) which are represented by the earlier blastomylonitic fabric of the typical Rye lithologies and the later concentration of strain into discrete zones of ultramylonite. The ductile fault zones are marked by the Portsmouth Fault Zone at the Rye-Kittery contact and the Great Common Fault Zone containing the southern ultramylonite, as well as numerous, smaller, mylonitic zones throughout the Rye (Figure 3). Both phases of ductile deformation may represent a continuum of events in the progressive development of this complex ductile high strain zone. The development of pseudotachylyte within the ultramylonites is interpreted as a later, brittle structural development. The major ductile fault zones and distinctive lithologic assemblages found at Gerrish Island have been traced along strike into New Hampshire (Carrigan, 1984a) where the ultramylonite at the Rye-Kittery contact is found to be folded about the nose of the regional SW-plunging antiformal structure.

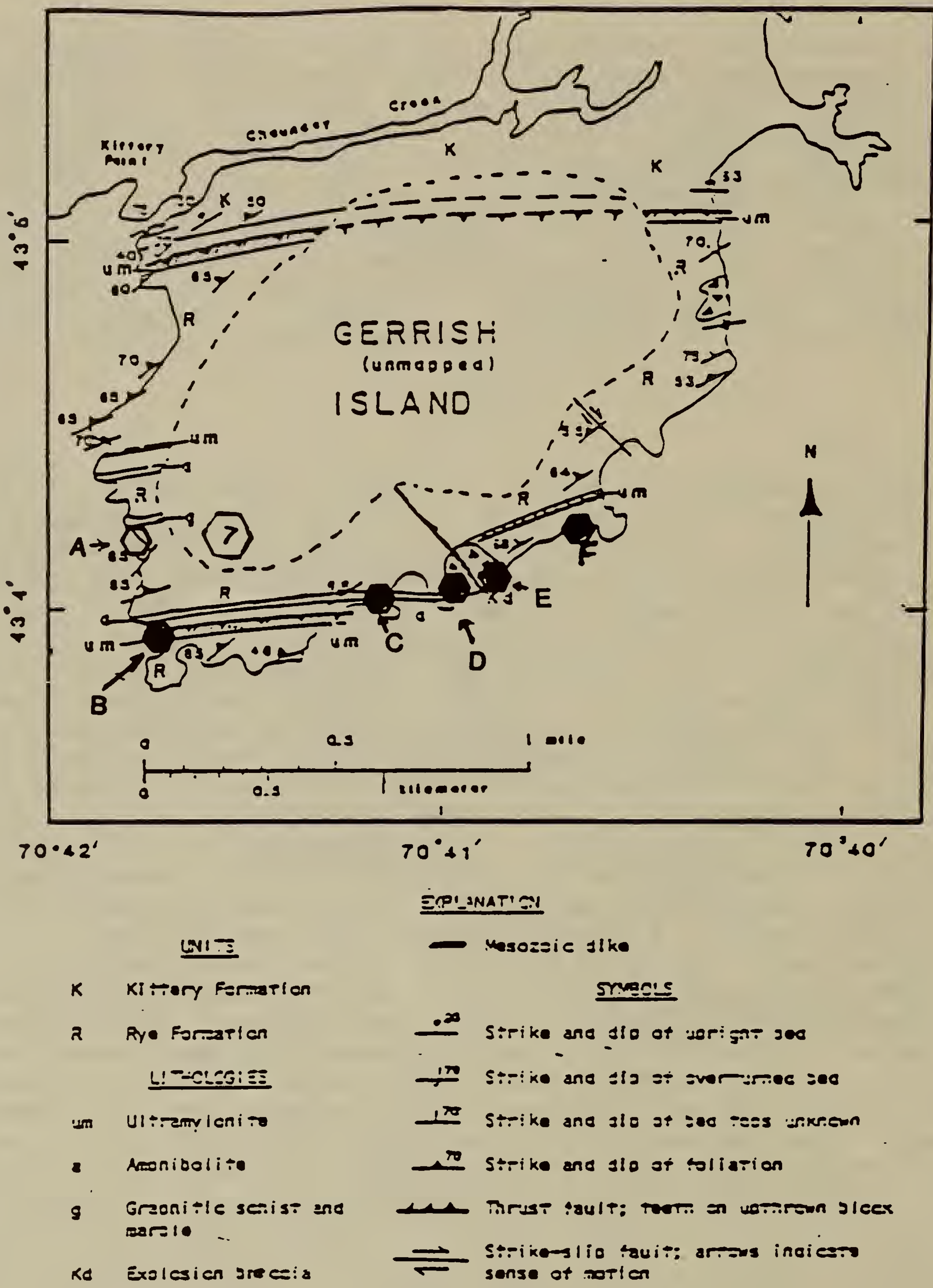
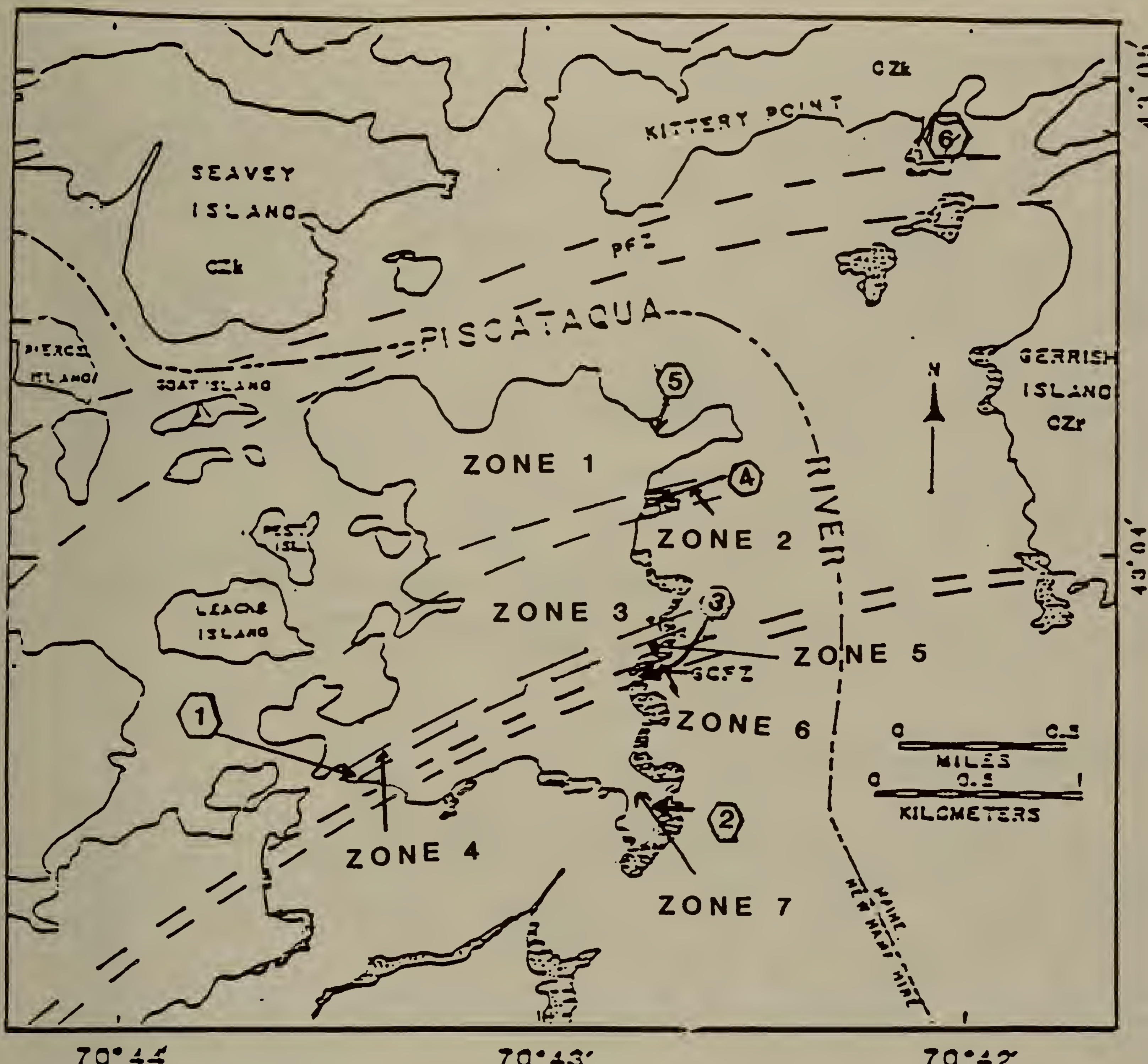


Figure 2. Geologic map of Gerrish Island, Maine (after Hussey, 1980).



Zone 1:- Predominatly mylonitized granitic rocks and calc-silicate bearing rocks with ultramylonite and metadiorite.

Zone 2:- Augen gneiss and calcsilicate bearing metasedimentary rock.

Zone 3:- Predominatly injected pelitic schist and mylonitized granitic rocks .

Zone 4:- Layered calsilicate bearing rock and amphibolite.

Zone 5:- Sulfidic schist and marble.

Zone 6:- Ultramylonite

Zone 7:- Pelitic schist, pegmatite, and quartzite.

Figure 3. Schematic Geologic Map of New Castle Island area New Hampshire.

METAMORPHISM:

The rocks of the Rye Formation have been subjected to poly-phase metamorphism as indicated by zoned amphiboles, garnet reaction rims and andalusite porphyroblasts with relict staurolite and sillimanite (Carrigan, 1984c). This suggests an initial sillimanite grade metamorphism followed by a lower andalusite grade event. In relation to the mylonitization, the early sillimanite grade metamorphism, with its crushed and rounded garnets and undulose extinction for sillimanite within the foliation appears to have been pre- to syn-tectonic. The later andalusite grade metamorphism appears to be post-tectonic as is evidenced by the undeformed andalusite porphyroblasts deflecting the foliation. The development of the ultramylonites of the Portsmouth and Great Common Fault Zones is interpreted as being post-metamorphic (Carrigan, 1984c) suggesting the occurrence of two phases of mylonitization (Carrigan, 1984a,c; Hussey, 1980). The metamorphic lithologies, structures and metamorphic isograds have all been folded by later multi-phase deformational events into the present-day doubly-plunging antiformal structure.

DUCTILE DEFORMATION FEATURES:

The most prominent effects of ductile deformation within the Rye Formation are the distinctive fold structures and the extensive development of a pervasive and locally intense blastomylonitic fabric, first recognized by Hussey (1980). The mylonitization of the nearly concordant felsic injections has produced a variety of granitic to dioritic blastomylonitic gneisses and schists. These rocks exhibit characteristic porphyroclastic textures which consist of larger, sheared and broken feldspar crystals in a more ductile recrystallized matrix of quartz and biotite.

Distinctive mesoscopic and microscopic textures are abundantly developed within the Rye Formation and can be used with care in determining the sense of shear. These textures include steeply-plunging asymmetric intrafolial folds, asymmetric feldspar augen, pressure shadow structures, oblique quartz shape fabrics, composite planar fabrics and displaced broken grains. The textural evidence examined, thus far, from these mylonitic rocks at Gerrish Island indicates a dominant dextral strike-slip component to this ductile deformation phase.

Two major zones of ductile faulting are developed within the deformed rocks of Gerrish Island (Figure 2) and are marked by the development of ultramylonite (Hussey, 1980). These ultramylonite zones occur at the Rye-Kittery contact and in the southernmost exposures at Gerrish Island and Newcastle Island where the ultramylonite appears to form a contact between the original metavolcanic and metasedimentary lithologic subunits. These units are now interpreted as representing different grades of metamorphism and intensities of granitic injection. Several other smaller zones of ultramylonite can be found throughout the perimeter exposures of Gerrish Island but are extremely difficult to correlate across the interior of the island. Both of the major ultramylonite zones have, however, been traced southwestward into New Hampshire as the Portsmouth and Great Common Fault Zones (Carrigan, 1984a,b). The major ultramylonite zone at the Rye-Kittery contact as the Portsmouth Fault Zone is found to be folded around the nose of the regional antiformal structure (Carrigan 1984a,b). This is an extremely important observation in making any regional interpretations of the nature and significance of these ductile fault zone structures.

REGIONAL STRUCTURAL INTERPRETATIONS:

The major ultramylonite zones have been interpreted as sites of significant deep-level ductile thrust faulting (Hussey, 1980). These faults have brought the higher-grade Rye Formation into contact with the lower-grade Kittery Formation at the Rye-Kittery contact and the lower-grade ,somewhat less-injected, lower Rye unit into contact with the slightly higher-grade, more intensely injected upper parts of the Rye Formation along the southern ultramylonite. This thrust interpretation is also supported by a recent structural and tectonic synthesis by Lyons et al (1982) for this and other fault-bounded basement blocks within this portion of New England as well as detailed studies of similar structures and blastomylonitic lithologies by Ratcliffe & Harwood (1975). However, one must reconcile the apparent age and metamorphic relationships between the Rye Formation on the SE and the Kittery Formation on the NW with the prominent NW dip of the foliation and the abundant evidence for a dominant dextral strike-slip component to the faulting with any model for a thrust-type deformation history.

The fact that the ultramylonite zones are found to be folded around the nose of the Rye anticline suggests the occurrence of a folded and possibly refolded, original thrust fault system. This relationship with the regional fold system eliminates any possible correlation with the Carboniferous Clinton-Newbury or Norumbega fault systems and suggests a pre-Acadian origin. Such a complex folded thrust fault structure could yield apparent strike-slip components for thrusting along the limbs, and dip-slip components on the nose of a regionally-plunging antiformal structure. The exact timing and age relations of the metamorphic events, the mylonitization and ductile thrust faulting, and the later multi-phase refolding are still somewhat speculative at this time. However, these rocks and structures may represent a Late Precambrian or Early Paleozoic compressional deformation expressed in the development of deep ductile fault structures followed by an later phase of refolding to give the structures observed today.

BRITTLE DEFORMATION FEATURES:

Pervasive, and locally intense, brittle deformational structures can be found superimposed on all of the ductile deformation features described in the previous section. These brittle structures can be grouped into an early, possibly Late Paleozoic phase and a later, Mesozoic phase of brittle structural development. The earlier phase of Paleozoic brittle deformation was dominated by dextral layer-parallel strike-slip and a subhorizontal layer-parallel extension accompanied by the production of pseudotachylyte in complex zones of intense brittle shear fracturing and brecciation. The later phase of Mesozoic brittle deformation was dominated by a subhorizontal layer-parallel compression resulting in conjugate shear fracture and kink-band structures associated with a variety of mafic and felsic intrusions.

LATE PALEOZOIC(?) BRITTLE FRACTURING:

The most conspicuous structures associated with the earlier brittle deformation are pairs of conjugate strike-slip shear fractures. These structures form easily recognizable "horst & graben" structures where cross-cutting the prominent near-vertical mylonitic foliation. Larger shear structures are also prominently developed at a small oblique angle to the foliation with dextral strike-slip offsets of distinctive lithologic layers. These larger displacements (up to approximately 10 meters) are sufficient to sharply truncate the along strike continuation of any mappable unit within these mylonitic gneisses, at least locally at Gerrish Island. Layer-parallel shear structures are also evident in these exposures but are sometimes difficult to recognize in the absence of any cross-cutting pre-existing geologic structures to determine offset relations. However, several complex brittle structural configurations and some offset cross-cutting pseudotachylyte veins suggest the need for significant layer-parallel dextral strike-slip faulting.

PSEUDOTACHYLYTE GENERATION ZONES:

Pseudotachylyte is found to be locally abundant within the brecciated outcrops of the southern ultramylonite zone at

Gerrish Island, Maine and its continuation into Newcastle Island, New Hampshire. The pseudotachylyte is found as rootless veinlets first recognized by Hussey (1980) within the brecciated ultramylonite and along the sheared contacts with the adjacent rock units. Discrete "pseudotachylyte generation zones" can be recognized within the brecciated outcrops at Gerrish Island and Newcastle Island (Swanson,1982). Each generation zone , as originally described by Grocott (1981) for brittle shear zones in Greenland, is defined by a pair of layer-parallel or near-layer-parallel slip surfaces. Individual pseudotachylyte generation zones may vary from a few millimeters or less to over a meter in width. It is along these nearly layer-parallel slip surfaces that pseudotachylyte is generated and subsequently injected into an internal geometric array of shear and extension fractures. All shear fracture surfaces within these Gerrish Island exposures are nearly vertical indicating a subhorizontal compression and a dominance of a strike-slip component to the deformation. These brittle pseudotachylyte-bearing zones are interpreted as characteristic structures related to stick-slip along brittle seismic fault zones (Grocott, 1981;Swanson, 1982). Any regional structural configuration for this early brittle deformation must account for the dominant dextral shear fracturing, both layer-parallel and slightly discordant, as well as the development of the conjugate layer-parallel extensional structures. An EW-trending dextral strike-slip shear couple is hypothesized to be responsible for this brittle structural assemblage. The near-vertical layer-anisotropy within the brittle ultramylonite zones would be in a favorable orientation for reactivated dextral slip and layer-parallel extension within the hypothesized shear couple. This zone of limited dextral strike-slip deformation may continue westward where it appears as a late brittle reactivation of the Rye-Kittery contact, described by Novotny (1969) as the Portsmouth Fault Zone or may correlate with a zone of offset magnetic anomalies described by Birch (1983).

BRITTLE MESOZOIC STRUCTURES:

Structures produced during the later brittle deformation phase consist of a prominent series of NE-trending basaltic dike intrusions. These dike intrusions are also intimately associated with a set of conjugate strike-slip fault and kink-band structures. This assemblage of brittle structures indicates a subhorizontal near layer-parallel compression for this phase of deformation. Similar relationships have been documented at the Seabrook Nuclear Power Plant foundation site by Bellini et al (1982). Cross-cutting relationships in conjunction with K-Ar age determinations for key dikes at this site indicate a Late Triassic age for this later brittle structural phase of deformation.

In addition to the basaltic intrusions developed throughout this region there is a series of explosive igneous breccias (Figure 2) locally developed along the SE shoreline of Gerrish Island, first described by Hussey (1962). These explosion breccias, mapped in detail by Swanson (1982), show a distinct association between the intrusion of rhyolitic melts and the brecciation. The rhyolitic melts often occur as sheath-like structures and peripheral dikes about the virtually matrix-free explosion vent breccias.

These vent breccias are exposed in several discrete structures clustering in northern and southern groups. Within the intervening kilometer or so of outcrop, later-stage hornblende lamprophyre dikes are found to carry an impressive assemblage of plutonic xenoliths that are mafic to felsic in composition. The intrusive rocks represented by these plutonic xenoliths include gabbros, hornblendites, diorites, syenites, alkaline syenites and syenite pegmatites. Quartz-bearing granitic plutonic types are extremely rare.

This has led to the hypothesis of a subsurface White Mountain intrusive complex beneath Gerrish Island which had not intruded through the present level of exposure. This inferred gabbro-syenite intrusion, the Gerrish Island Igneous Complex (Swanson, 1982), can only be seen in outcrop as roof structure breccias, upper-level felsic melts, basaltic intrusions and sampled subsurface plutonic rock types. The occurrence of basaltic xenoliths within the breccias and the development of cross-cutting basaltic dike intrusions suggests a possible Late Triassic age for this subsurface intrusive complex. This phase of crustal extension and intrusion would be more closely related to the Agamenticus intrusive activity. The lamprophyric dikes on the other hand appear to cross-cut most of the basaltic intrusions as well as the felsic melts and vent breccias.This late intrusive phase of hornblende lamprophyric dikes, hornblendite magmas and other associated intrusives would be more closely related to the intrusive activity of the Cape Neddick Gabbroic Complex.

The youngest structures to develop include a set of NNE-trending brittle extensional fractures that may be related to a late phase of dike intrusion. These fractures are associated with a silicic hydrothermal alteration that produces bleached, rusty-weathering alteration haloes about the open fractures. Their intimate association with the last phase of dike intrusion would suggest that these fractures may represent a continuing phase of crustal extension after the cessation of partial melting at depth. Circulating hydrothermal solutions during this waning stage of extension would be responsible for the conspicuous wall-rock alteration.

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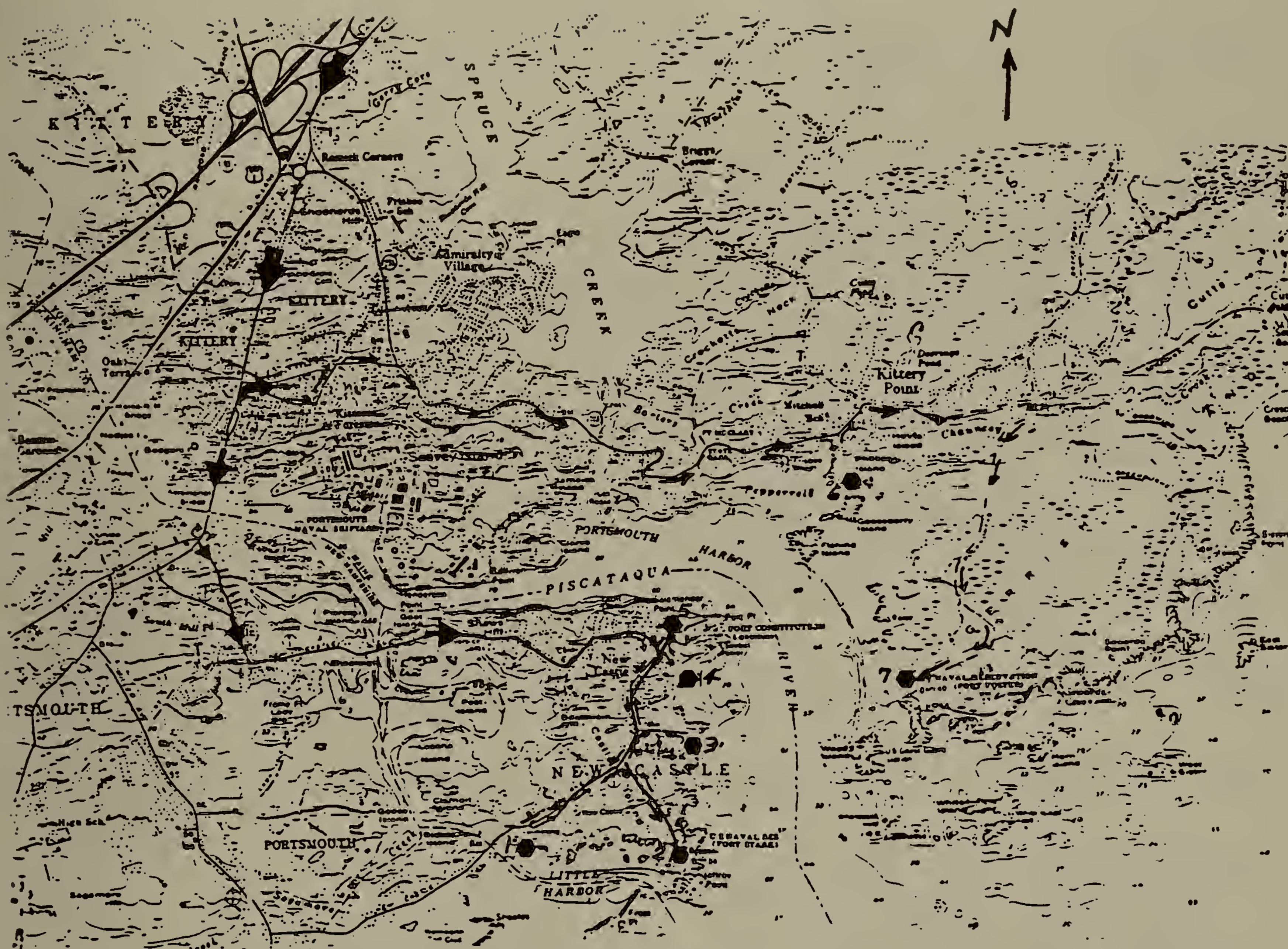

/ ⬣ Stop Location

FIGURE 4: Trip Map

ROAD LOG:

Mileage:

0.0 Assembly point at information and comfort station at Kittery, Maine Rest Area- 2.9 miles north of the Piscataqua River bridge on interstate 95 North. Exit from the rear of the Rest Area, turning right on to U.S. Rt. 1 South.

2.4 Take Kittery Rt. 1 exit, turn left at end of ramp on to U.S. Rt. 1 South.

2.7 Traffic circle, exit right continuing south on U.S. Rt. 1.

3.8 Straight through both stop lights continue south on U.S. Rt. 1.

4.4 Sharp right at south side (New Hampshire side) of Rt. 1 Memorial Bridge, proceed beneath bridge.

4.5 Left onto Darcy Street at park, continue straight on Darcy Street to the end.

4.8 Left onto Rt. 1B South, proceed along south shore of Piscataqua River.

6.4 Entering NW portion of Newcastle Island, proceed on Rt. 1B around island to the SW side.

8.0 STOP 1: Wentworth-by-the-sea, SW Newcastle Island: Stratigraphy and structure of the upper "metavolcanic" unit of the Rye Formation (Zones III, IV, & V) in approaching the Great Common Fault Zone (SW extension of the southern ultramylonite of Hussey 1980). Rock types and structures from NW to SE include variably injected, sheared & refolded amphibolite with gently NE-plunging fold structures; calc-silicate gneisses; sulfidic schists; marble with steeply SW-plunging dextral asymmetric folds, containing pyroxene and epidote consistent with amphibolite facies metamorphism; variably injected pelitic and quartzo-feldspathic metasediments; and pseudotachylyte-bearing ultramylonite with its characteristic creamy well-foliated texture.

8.0 Reverse direction and proceed north back on Rt. 1B.

8.6 Right on to Wild Rose Lane, proceed south to end of lane.

9.0 STOP 2: Fort Stark, SE Newcastle Island: The lithology and structure typical of the lower "metasedimentary" unit (Zone VII) of the Rye Formation SE of the Great Common Fault Zone. Variably injected schists of this zone contain abundant andalusite porphyroblasts with inclusions of staurolite, sillimanite and garnet as well as the foliation suggesting a second phase of metamorphism. Discrete pegmatite injection bodies have been intensely- sheared and preserve fractured tourmaline crystals floating in a more ductile pegmatitic quartzo-feldspathic matrix. These lithologies and ductile deformation features have been overprinted by later EW-trending oblique dextral shear zones, kink-band structures and Early Mesozoic dike injections.

9.2 Reverse direction, proceed back to Rt. 1B and turn right going north on Rt. 1B.

9.3 Turn right into Newcastle Common Park at southeast corner near baseball diamond.

9.3 STOP 3: Great Island Common, E New Castle Island: Variably injected, sheared and multiply-folded amphibolites (Zone V) and a gradual(?) transition into ultramylonite (Zone VI) of the Great Common Fault Zone (southern ultramylonite of Hussey, 1980). Both lithologies have been complexly folded with gently-plunging structures being the most prominent feature. within the ductile fault zone with The ultramylonite also exhibits the characteristic creamy, well foliated texture typical of these zones and style of brittle deformation that is associated with the development of pseudotachylyte generation zones. Near-layer-parallel slip surfaces can be observed some filled with dark fine-grained pseudotachylyte. These slip zones may develop an internal brecciation as well. The center of the ultramylonite within this fault zone develops a chaotic internal texture in contrast to the well foliated nature of most of the ultramylonites. The later development of altered joint systems can also be observed.

10.0 Leave New Castle Common turning right on to Rt. 1B North.

10.4 At right-angle bend in Rt. 1B proceed straight on to Main Street and turn right on to Ocean Street. Proceed to end of street.

10.4 STOP 4: Ocean Street Beach, E Newcastle Island: Augen gneisses (Zones II& III) of the upper "metavolcanic" unit of the Rye Formation can be seen as intensely injected and sheared amphibolitic metasediments. The feldspar augen are preserved as porphyroclasts of an originally coarse-grained pegmatitic injection. Numerous lenses of amphibolite are preserved within the augen gneisses as remnants of the original injected lithology. Late oblique dextral shears are also abundantly developed.

10.5 Proceed back to Main Street and turn right continue to right angle bend in Main Street.

10.5 STOP 5: Coast Guard Station Pier, NE Newcastle Island: This outcrop exhibits the lithologic and ductile-brittle deformational complexity characteristic of the Rye Formation in general. Rock types include variably injected amphibolites, sheared augen gneisses and mylonitized granitic gneiss with its irregular aplitic injections as well as basalt dikes of Mesozoic age. The multi-phase fold structures are complex. The overprinted brittle structures include a prominent and well developed EW-trending dextral shear structure with its own assemblage of minor fractures and sigmoidal en echelon gash veins. These structures are clearly cross-cut, and locally intruded by a Mesozoic basalt dike.

10.8 Proceed south back on Main Street and turn right on to Rt. 1B North.

11.7 Leaving Newcastle Island.

12.4 Take right off of Rt. 1B on to Darcy Street and proceed straight to end of Darcy Street.

13.2 Turn right at end of Darcy Street proceed under bridge.

13.3 Left at stop sign and proceed North on Rt. 1 across Memorial Bridge.

14.1 Take right at second stop light on to Rt. 103 East.

14.9 Right at intersection of Rt. 263 and Rt. 103.

14.9 Continue East on Rt. 103.

17.1 Right onto hidden drive marked by "Rutelidge" sign on tree. Continue to end of road parking on southeast corner of Philips Island. Please note we are on private property and here only at the courtesy of the owner. Please stay with the group leaders.

17.1 STOP 6: Philips Island, Maine, Midway between Newcastle and Gerrish Islands: These outcrops exhibit Kittery-type lithology as a purplish weathering, actinolite-bearing calcareous quartzite. These Kittery-type litholgies grade southeastward into the ultramylonites of the Rye-Kittery contact exposed at Gooseberry Island. These Kittery litholgies also develop their own internal ultramylonite zones toward the northwest suggesting that the Portsmouth Fault Zone developed within rock types of both the Rye and Kittery Formations generating a tectonic Rye-Kittery transition zone. The Rye-Kittery Contact, NW Gerrish Island, as originally described by Hussey (1980), can be seen from this point. The tectonic contact between the Rye and the Kittery Formations is marked by a 100' wide zone of ultramylonite. The transitional character of this contact zone is due to mylonitization of both lithologies resulting in a gradational change southeastward from the Kittery lithologies into the mylonitic and metasedimentary lithologies of the typical Rye Formation.

17.6 Proceed back to Rt. 103 and turn right on to Rt. 103 East.

17.9 At left curve in Rt. 103 continue straight off of Rt. 103 on to Chauncey Creek Road.

18.4 Turn right off of Chauncey Creek Road and cross bridge to Gerrish Island. Turn right at south end of bridge on to Pocahontas Road and continue to Fort Foster and Stops 7A-F.

19.6 Enter gate at Fort Foster and proceed to parking lot, if gate is open. Otherwise we will park at gate and walk into the Park.

STOP 7A: Pre-cataclastic metasedimentary lithologies, W Gerrish Island: Remnant metasedimentary lithologies include graphitic-sulfidic schists, marble, and amphibolite preserved in gently plunging antiformal structures. Also included in these exposures are examples of the injected felsic lithologies exhibiting their characteristic porphyroclastic textures as mylonitic augen gneisses and minor ultramylonite zones.

STOP 7B: Southern Ultramylonite zone, S Gerrish Island: The southern ultramylonite zone forms a complex structural boundary between the upper, more heavily-injected, "metavolcanic' unit and the lower, less-injected, metasedimentary unit. The lithologies present include amphibolite, pelitic-schists, mylonitic granitic gneisses and abundant ultramylonite that demonstrate the progressive although complex mylonitization of these rocks.

The ultramylonite exhibits internal fold structures of a gently plunging isoclinal nature. A localized high strain zone within the adjacent biotite schist unit contains a sheared lens of exotic, ductilly-deformed marble as a remant lithology similar to the marble observed in the last outcrop. The internal structure of this zone consists of moderately plunging Z-shaped asymmetric intrafolial folds.

Both the ultramylonite and adjacent cataclastic lithologies of this zone are cross-cut by assemblages of brittle structures representing subhorizontal layer-parallel extension and layer-parallel-compression. The layer-parallel extensional structures consist of conjugate near-vertical strike-slip faults which combine to form conspicuous "horst & graben" structures where cross-cutting the prominent lithologic layering in these rocks. Near layer-parallel dextral shear structures are also prominantly developed representing an intense brittle deformation phase that has resulted in the formation of pseudotachylyte. The pseudotachylyte is produced in distinctive generation zones defined by two near layer-parallel principal shear surfaces as the sites of generation, bounding an injected internal, geometric fracture array. Numerous pseudotachylyte generation zones are developed in these outcrops and range from a few centimeters to over a meter in width.

Early Mesozoic dominantly NE-trending structures related to a second phase of brittle layer-parallel compression have been superimposed on all previous structures and lithologies. These structures include conjugate strike-slip faults, kink-band structures and the injection of basaltic dikes.

The remaining outcrops are on private property. Please stay with the group leaders.

STOP 7C: Cedar Point, SE Gerrish Island: This outcrop represents an isolated block of unmylonitized calcsilicate laminated amphibolite preserved within ductile shear zone lithologies. This block essentially forms a "mega-porphyroclast" whose internal structure can indicate sense-of-shear within the zone as well as the nature of the pre-mylonitic geologic processes. The internal structure of the amphibolite exhibits components of an early phase of layer-parallel flattening in the form of discordant highly-contorted ptygmatic granitic veins and small-scale boudinage structures within the foliation. Thicker, nearly-concordant granitic injection layers remain virtually undeformed except where crosscut by sets of conjugate ductile-brittle fault zones and zones of en echelon vein arrays. The granitic injections are mylonitized where they are crosscut by these faults, particularly along the dominant NNE-trending sinistral shear structures of this conjugate set. The dominance of these internal, sinistral shear structures produces an effective clockwise rotation of the segmented amphibolite blocks as would be expected of thin section scale displaced broken grains in a more ductile recrystalling matrix within a ductile shear zone.

This amphibolite block is bounded by the mylonitic ductile shear zone lithologies that include augen gneisses and thinner but more intense zones of mylonitization. Both of these rock types exhibit textural characcteristics in the form of asymmetric feldspar augen structures which indicate a dominant component of dextral strike slip.

This outcrop is also located at the outer fringe of the roof-structure assemblage which represents the near-surface expression of a subsurface Mesozoic White Mountain intrusion. This roof-structure assemblage is represented by the beginning of rhyolitic dike intrusion as the outcrop is approached from the SW and by the development of the southernmost vent breccia at the NE end of the outcrop. Vent breccias are characteristically chaotic, virtually matrix-less and are often bordered by a sheath of rhyolitic melt. Abundant NNE-trending basaltic intrusions are also present within these outcrops, many of which appear to be reactivating the dominant sinistral shear structures within the host amphibolite block.

STOP 7D: Yellow Rocks, SE Gerrish Island: Excellent vent breccia exposure exhibiting the chaotic ,matrix-free characteristics, as well as, the nature of, and variation in the xenolith compositions. The metamorphic xenolith composition changes within the vent breccias, reflecting the abrupt change in lithology that is apparent within this complex ductile high strain zone. Basaltic dike rocks occur as fragments within the breccia and as cross-cutting intrusions. A sheath of rhyolitic melt can also be seen along the NE contact of this vent breccia exposure.

STOP 7E: Dike Point, SE Gerrish Island: This outcrop represents the largest, and one of the youngest, basalt dike intrusions in the area and exhibits an anomalous NW trend. This intrusion crosscuts all lithologies, structures and intrusions . Excellent vent breccia exposures on the NE margin of the dike contain abundant basaltic xenoliths and rhyolite is present on the SW margin.

STOP 7F: Robbins Rocks, E Gerrish Island: These exposures contain abundant ductile shear zone lithologies which range from mylonitic biotite schists and coarse-grained porphyroclastic augen gneisses to finer grained blastomylonites and thinner zones of ultramylonite toward the NE along this coastal section. The ductile shear zone rocks are cross-cut in these exposures by numerous NW-trending dextral shear zones that are correlated with the pseudotachylyte-bearing structures within the southern ultramylonite exposures.

These rocks and structures are, in turn, cross-cut by a series of intrusions representing, at least in part, the roof assemblage for the hypothesized subsurface intrusive complex. These intrusions begin with several NE-trending quartz-breccia veins that are crosscut by numerous NNE-trending basaltic dike intrusions. The basalts are then crosscut by several hornblende lamprophyre dikes that carry a complete assemblage of sampled plutonic xenoliths from this subsurface complex. The plutonic xenoliths range in composition from gabbro to syenite similar to the White Mountain Magma Series.

0.0 Return to Fort Foster and vehicles.

1.7 Retrace route from Fort Foster Gate back to Rt. 103 and proceed west.

4.4 Intersection of Rt. 103 and Rt. 263; continue on Rt. 103 to Rt. 1 or continue on Rt. 263 back to Interstate 95.

GEOLOGIC FRAMEWORK OF THE MASSABESIC ANTICLINORIUM AND THE MERRIMACK TROUGH, SOUTHEASTERN NEW HAMPSHIRE

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Introduction

The purpose of this trip is to examine the field relations between the gneisses of the Massabesic Anticlinorium and the uppermost units of the Merrimack Trough in southeastern New Hampshire and some of the igneous rocks that intrude them. Based on ongoing mapping, isotopic, structural, and petrologic studies, both the Massabesic gneisses and the Merrimack Group (Kittery, Eliot, and Berwick Formations) are interpreted as late Precambrian in age. The age of the Merrimack Group itself cannot be fixed with certainty, yet minimum ages are established by dates of the numerous plutonic rocks that intrude it. These rocks likely represent a separate lithotectonic belt between the Gander and Avalon zones of Williams (1978) (Lyons and others, 1982; Olszewski and others, 1984).

Geologic Setting and Previous Work

In southeastern New Hampshire northeast-trending, polydeformed, low to high grade metamorphic rocks crop out within the Massabesic Anticlinorium and the Merrimack Trough (figure 1). To the northwest a sequence of Siluro-Devonian rocks, in part dated by fossils, occupy the Kearsarge-Central Maine Synclinorium as recently redefined by Lyons and others (1982) (see Eusden and others, this volume). To the southeast and along the New Hampshire and southwesternmost Maine coast, variably mylonitized rocks of the high grade Rye Formation define the Rye Anticlinorium (see Swanson and Carrigan, this volume).

The Massabesic Anticlinorium is composed of high grade felsic gneisses, migmatite, and crosscutting felsic igneous rocks. The Merrimack Trough is composed of metaclastic rocks of the Merrimack Group (Kittery, Eliot, and Berwick Formations from oldest to youngest) which are intruded by the 473 Ma Exeter Diorite (Gaudette and others, 1984) and other mafic and felsic igneous rocks (figure 2). No fossils have ever been found in any of the metasedimentary rocks despite diligent search by many workers. Isotopic ages from the Massabesic gneisses and the Exeter Diorite provide the basis for our evolving reinterpretation of the geology of southeastern New Hampshire.

As the study of these rocks began more than a century ago and there has been no concensus of opinion about their age and correlation, we begin with a brief historical review. Billings (1956) elevated the stratigraphic name "Merrimack," first used by Hitchcock (1877), to the rank of Group. He included three formations, the Kittery, Eliot, and Berwick, as established by Katz (1917) and assigned an age of Middle Silurian based on correlation with similar rocks in central Maine (see Billings, 1956, p. 44, 104). This package

Figure 1. Geologic map of southeastern New Hampshire and southwestern Maine (after Lyons and others, 1982)

was interpreted to constitute the southeastern limit of the Merrimack Synclinorium and was observed by Billings (1956, p. 67) to be cross-cut by the Fitchburg Pluton, which he recognized then as a more complex igneous sequence than a simple pluton.

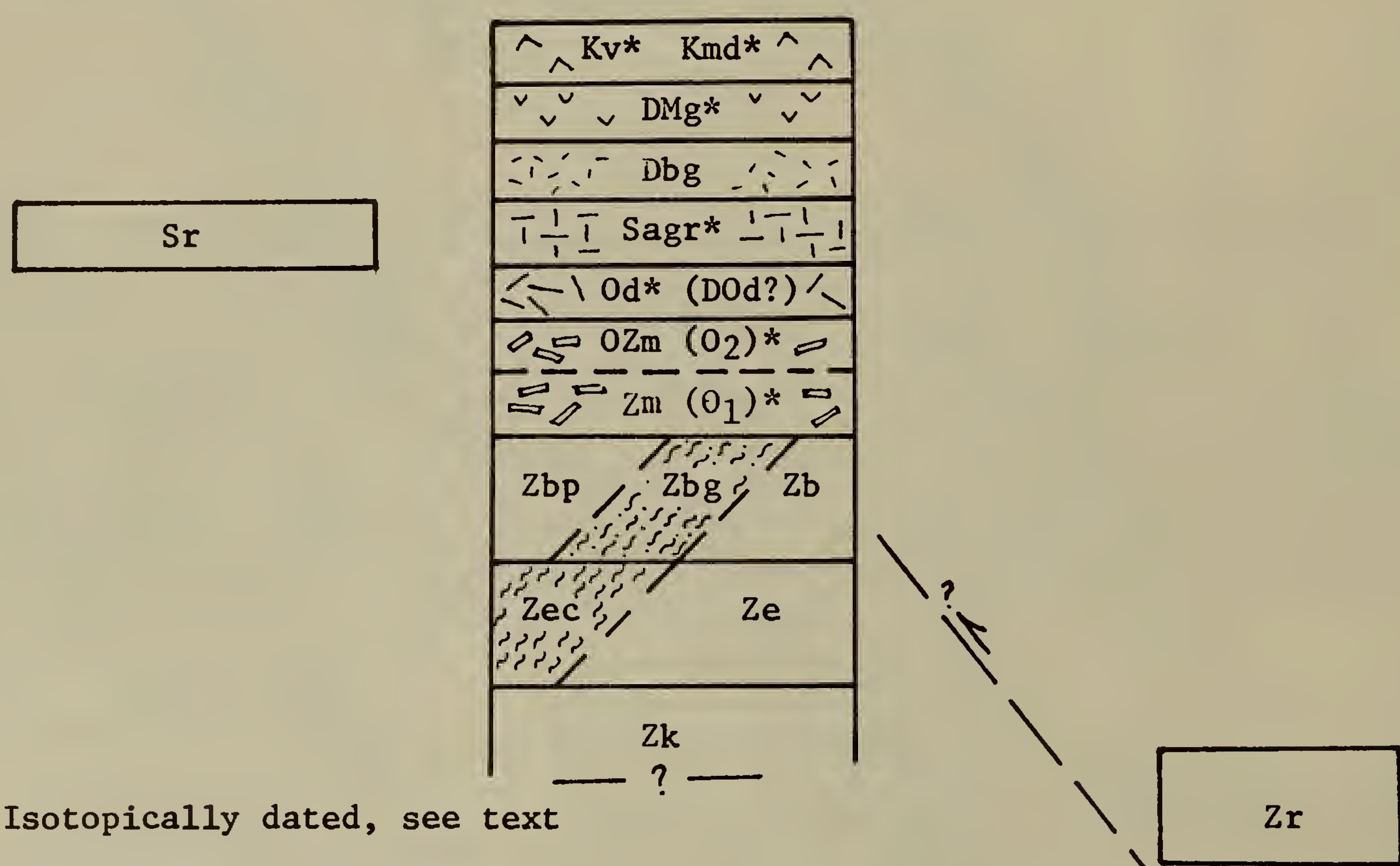
Isotopic dating within southeastern New Hampshire (Besancon and others, 1977; Aleinikoff, 1978; Aleinikoff and others, 1979; Kelly and others, 1980) suggested a late Precambrian age for some, if not all, of the Merrimack Group rocks. Lyons and others (1982) suggested that the Merrimack Group rocks should be separated from those of the Merrimack Synclinorium of Billings and reassigned to their own structural succession, the Merrimack Trough. This structure was originally and perceptively proposed by Emerson (1911). In order to avoid further confusion Lyons and others (1982) also recommended that the name Merrimack Synclinorium be discontinued and a new name, the Kearsarge-Central Maine Synclinorium (KCM) be adopted for the layered rocks northwest of the Fitchburg (Massabesic) "pluton." The metasedimentary rocks of the KCM are demonstrably of Silurian and Devonian age (Hatch and others, 1984; see also other references therein).

E X P L A N A T I O N

The geologic map is modified after Lyons and others (1983) and Zen (1983)

LITHOLOGIC KEY

KEARSARGE-CENTRAL MAINE SYNCLINORIUM	MASSABESIC ANTICLINORIUM MERRIMACK TROUGH	RYE ANTICLINORIUM



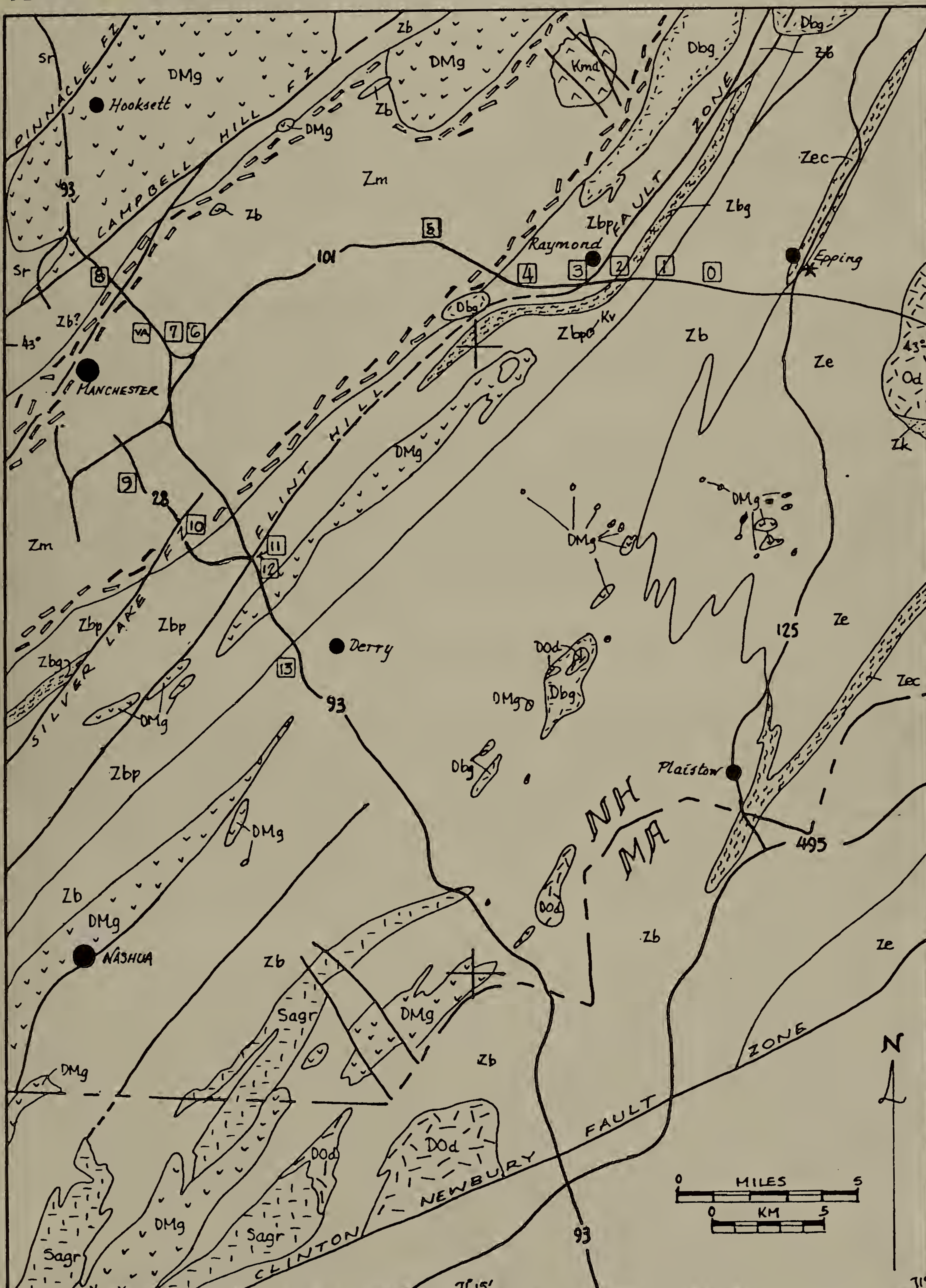
* Isotopically dated, see text

CRETACEOUS	Kv,Kmd	White Mountain Series (Little Rattlesnake Hills volcanics, Mt. Pawtuckaway Complex monzonites and diorites)
DEVONIAN-MISSISSIPPIAN	DMg	New Hampshire Plutonic Series - Two-mica granites
DEVONIAN	Dbg	Biotite granites
DEVONIAN to ORDOVICIAN?	DOd	Diorite (Dracut and Sweepstakes)
SILURIAN	Sagr	Ayer Granite and associated rocks
	Sr	Rangeley Formation
ORDOVICIAN	Od	Exeter Diorite and associated rocks
ORDOVICIAN to PRECAMBRIAN	OZm(O_2)	Massabesic orthogneiss
PRECAMBRIAN	Zm	Massabesic Gneiss Complex (ortho and para)
	Zb	Merrimack Group - Berwick Formation
	Zbp	Berwick Formation high pelite component
	Zbg	Gove Member
	Ze	Eliot Formation
	Zec	Calef Member
	Zk	Kittery Formation
	Zr	Rye Formation (not shown on this map)

✳ STARTING POINT

☐ FIELD TRIP STOPS

FIGURE 2 GEOLOGIC MAP OF SOUTHEASTERN NEW HAMPSHIRE AND ADJACENT MASSACHUSETTS



Lyons and others (1982) also found no basis for the Rockingham Anticlinorium in the region occupied by the Merrimack Group and redefined the Rye anticline of Billings as the Rye Anticlinorium. They believed that these rocks more closely resembled those of the Avalon zone of Williams (1978) than those of his Gander zone. Based on lithic sequence, timing of metamorphism and plutonism, and deformation, Olszewski and others (1984) have recommended that the Merrimack Group and bounding Massabesic and Rye complexes represent a separate exotic terrane between the Gander and Avalon terranes. Its relation to Zen's (1983a) Craton X is unknown.

The rocks along the northwest margin of the Merrimack Trough, previously included within the Fitchburg pluton, are paragneisses, orthogneisses, and migmatites cut by numerous two-mica granites that have been referred to as the Massabesic Gneiss (Srirarmidas, 1966; Aleinikoff and others, 1979), the Massabesic migmatite (Boudette, 1977), and the Massabesic Gneiss Complex (MGC; Kelly and others, 1980). Their distribution defines the Massabesic Anticlinorium of Lyons and others (1982), but they note a lack of structural data to support the interpretation of a broad structural arch (STOP 5). The migmatite and gneisses contain enclaves of high grade rocks that we recognize as the Berwick Formation and provide a basis for suggesting that the rocks of the Merrimack Group, specifically the Berwick Formation, are gradational with the paragneiss of the Massabesic. Alternative possibilities include an unconformity or a fault boundary, both obscured by late migmatization. These we hope will be prime points of discussion for the trip.

Stratigraphic Framework

The immediate concern of this trip are lithologic, petrologic, and structural data from the Berwick Formation and the Massabesic gneisses and the isotopically dated igneous rocks that cross cut them. The other units however deserve brief comment. The rocks of the Rye Anticlinorium will not be examined on this trip although they play an important part in the geologic history of southeastern New Hampshire (see Hussey, 1980; Carrigan, 1984a, 1984b; and Swanson and Carrigan, this volume). They represent a belt of high grade metasedimentary and very minor metavolcanic rocks cut by felsic metaigneous gneisses nearly parallel to the Massabesic Anticlinorium. All are strongly mylonitized.

The Kittery and Eliot Formations, the lower units in the Merrimack Group, similarly will not be examined. Sriramades (1966) and Sundeen (1971) consider the rocks at "STOP 0" and STOP 13 to be Eliot Formation; however, the lithologies are virtually identical to the Berwick Formation to be examined at STOPS 1 and 8. The interested reader is referred to Katz (1917), Billings (1956), Hussey (1968), Rickerich (1984), and Hussey and others (this volume) for details concerning the stratigraphic evidence supporting a conformable, northwest-facing homoclinal sequence for the Merrimack Group. Merrimack Group equivalents in Massachusetts have been treated by Peck (1976), Robinson (1979), Barosh and Pease (1981), and Zen (1983b) among others. In Maine Osberg (1968, 1978) considered the Merrimack Group equivalent to the Vassalboro sequence. Most previous workers have followed Billings and assigned the Merrimack Group to the Silurian, although the new geologic map of Maine (Osberg and others, 1984) allows the possibility of an Ordovician or older age.

Berwick Formation

The Berwick Formation lies stratigraphically above the Eliot Formation and in places, above the discontinuous, black, graphitic, phyllitic Calef Member of the Eliot Formation (fig. 2) as recognized by Freedman (1950) in the Mt. Pawtuckaway 15-minute quadrangle. Freedman (1950) also mapped rocks that he assigned to the Littleton Formation and the very pelitic Gove Member of the Littleton Formation in that area. Billings (1956) reassigned the Littleton there to the Berwick Formation which we retain, noting that a lithic distinction is present and probably traceable to the southwest.

The Berwick Formation within the field trip area is divisible into three subunits based on lithologic grounds and degree of partial melting. In areas least affected by partial melting the Berwick Formation is best described as a well bedded, laminated, purplish-gray biotite granofels with thin often discontinuous calc-silicate layers, stringers, pods, and occasional boudins (STOP 1). Bedding, some graded, in the granofels ranges from a few centimeters to 20 centimeters and contains thin, often rusty weathering pelitic interlayers. On average, biotite granofels occupies 70%; calc-silicate, 20% (max); and nonrusty to rusty pelitic schist or phyllite, 10% of an outcrop. The biotite granofels typically includes biotite + plagioclase + quartz with accessory pyrite, tourmaline, zircon, and apatite. Chlorite, white mica, calcite, sphene, epidote, and rutile also occur locally. The intercalated dark gray, sometimes rusty weathering biotite schist (or phyllite) consists of quartz + biotite + plagioclase + sphene + pyrite ± graphite ± white mica ± tourmaline at low grade. Calc-silicate assemblages include quartz + plagioclase + calcic amphibole + epidote ± biotite + accessory sphene, carbonate, and pyrite.

The Gove Member is exposed southeast of the Flint Hill Fault Zone (FHFZ) from Nottingham Square through Raymond, and again near Merrimack and South Merrimack, NH. It is magnificently exposed along the new section of Rt. 101 southwest of Raymond (STOP 2). It is best characterized as a strongly crenulated, silvery gray white mica + quartz + biotite + staurolite + garnet + sillimanite schist with accessory tourmaline and apatite. Chlorite occurs as an alteration product of biotite.

The Berwick Formation adjacent to the Gove Member both to the west and east is much like that already described in that it is dominantly biotite granofels with calc silicate interlayers. However, the proportion of pelitic intercalations is greater, overall grain size is coarser, and calc-silicate lenses are more irregular and pinstriped. West of the Gove Member near the MGC, pegmatite and foliated biotite granite are abundant (STOPS 4, 10). The increased pelitic component was probably the reason that Freedman originally assigned this to the Littleton Formation and A.M. Hussey (1983 personal communication) has indicated that this zone is a bit atypical for the Berwick Formation at the type locality of Berwick, ME. We indicate the change by Zbp on Figure 2 and suggest that the changes are a function of continuously increasing metamorphic grade as the Massabesic gneiss contact is approached, and perhaps facies change, within the Berwick Formation.

This part of the Berwick Formation is strongly brecciated in the FHFZ (STOP 3). Importantly, however, there does not appear to be a discontinuity in metamorphic grade across the fault either along Route 101 or along I-93 (see section on Metamorphism). Furthermore, biotite granofels and calc-silicate interlayers, typical of the Berwick Formation, occur on both sides of the fault.

Massabesic Gneiss Complex (MGC)

The Massabesic Gneiss Complex occupies the core of the Massabesic Anticlinorium and is approximately bounded by the Campbell Hill-Nonesuch River (CHNR) fault zone on the northwest and by a gradational contact with the Berwick Formation on the southeast (Fig. 2). To the northeast it is terminated by intrusive rocks of the New Hampshire Plutonic Series (Anderson, 1978) Roadcuts along Rts. 43 and 202 to the northeast in Northwood and Rochester, as well as in Manchester show that the Berwick Formation also crops out along the northwest margin of the Massabesic Anticlinorium southeast of the CHNR (Lyons and others, 1983). The MGC consists of two major lithologic phases: a paragneiss that we believe is the high grade equivalent of the Berwick Formation previously described (also noted by Sriramadas, 1966, and Carnein, 1976); and an orthogneiss. A clear distinction between the two gneisses is often very difficult (and occasionally generates hot debate). For the purposes of the trip, orthogneiss contains unequivocal xenoliths of felsic and calcsilicate rock, among others, and paragneiss may contain enclaves of calc-silicate rock. Both gneisses are cross-cut by a variety of foliated to unfoliated biotite granite and two-mica granite bodies of varying sizes, mineralogies, and ages.

The paragneiss, usually a "swirly" gray migmatite, is coarse-grained and compositionally banded. Quartzofeldspathic layers are 1 to 20 cm thick separated by 1-5 mm biotite-rich layers and occasional discontinuous green calc-silicate enclaves (STOPS 5, 7). The felsic layers consist of quartz + plagioclase + microcline (often pink and may reach sizes up to 7 cm) + lesser biotite. The biotite-rich layers are dominated by large oriented biotite plates interspersed with white mica, and fibrolitic and/or prismatic sillimanite. Magnetite is present as a ubiquitous accessory, often locally abundant (1 cm euhedra), particularly in or near cross-cutting pegmatites. Calc-silicate layers and pods typically consist of hornblende + diopside + quartz + plagioclase ± epidote. Detrital zircons from paragneiss in the Milford, NH, area have yielded a provenance age of 1237 Ma (Pb^{207}/Pb^{206}) which corresponds closely to a 1188 Ma (Pb^{207}/Pb^{206}) age from detrital zircons from the Berwick equivalent Oakdale Formation in Massachusetts (Aleinikoff, 1979). This implies that the protolith (Berwick) was deposited after 1200 Ma.

Orthogneiss crops out throughout the MGC. Cross-cutting relations can usually be found in any large outcrop between orthogneiss and "country rock." The usually strongly foliated orthogneiss is typically gray or pink, medium- to coarse-grained microcline (white or pink) + quartz + plagioclase (oligoclase-andesine) + biotite gneiss (Kelly, 1980). White mica and garnet may be additional phases. Accessory zircon is homogeneous and euhedral. The best evidence for an igneous origin for this rock is the character of the zircons (Poldervaart, 1956) and the presence of frequently rotated xenoliths of paragneiss, biotite + plagioclase + quartz granofels, sillimanite-bearing pelitic schist, aluminous calc-silicate, and massive amphibole and epidote rock (STOP 6). The xenoliths that we equate with the Berwick Formation represent at present our best evidence for the subdivision of the MGC. Furthermore the xenoliths plus the facts that: (1) the foliation in the Berwick Formation strikes parallel to that in the MGC and (2) a continuous increase in metamorphic grade toward the MGC that reaches temperatures near the granite minimum are the bases for the close genetic relationship between the MGC and the Berwick Formation that we are proposing.

Intrusive Rocks and Available Geochronological Data

Intrusive rocks within both the Massabesic Gneiss Complex and the Merrimack Trough (MT) span late Precambrian to Cretaceous (?) time. The earliest intrusion is the 650 Ma (Bescancon and others, 1977; Kelly and others, 1980; Rb/Sr whole rock and U/Pb zircon) Massabesic orthogneiss previously described near Raymond and Manchester, NH. This rock was apparently intruded as semi-conformable sheets into the paragneiss. Aleinkoff and others (1979) have dated the paragneiss itself by U/Pb zircon methods and obtained an age of 646 Ma (maximum $^{207}Pb/^{206}Pb$). They consider the zircons to be of volcanic origin which implies that the protolith of the paragneiss was deposited in the late Precambrian. They also report another orthogneiss within the MGC dated at 475 ± 48 Ma (U/Pb zircon) whose relationship with the older orthogneiss is still uncertain.

Early Ordovician igneous activity within the MGC and MT are now corroborated by a new 473 ± 37 Ma Rb/Sr whole rock age of the Exeter Pluton (Gaudette and others, 1984) and a similar but metamorphosed diorite on Appledore Island (Olszewski and others, 1984). Other diorites in the same strike belt (Sweepstakes Diorite, Dracut Diorite) may be of similar age as suggested on Figure 2. The Newburyport Quartz Diorite as dated by Zartman and Naylor (1984) is slightly younger (450 ± 15 Ma $^{207}Pb/^{206}Pb$, zircon). These intermediate intrusive rocks place a minimum early Ordovician age for the rocks of the Merrimack Group. The nearby Ayer Granite and associated gneisses and the Chelmsford Granite also intrude the Merrimack Group and its equivalents in Massachusetts and have yielded ages of 433 ± 5 Ma ($^{207}Pb/^{206}Pb$, zircon) and 389 ± 5 ($^{207}Pb/^{206}Pb$, zircon) respectively (Zartman and Naylor, 1984).

Two-mica granites in this region are of special interest because of their possible genetic relationship to the metamorphic and tectonic events that shaped southeastern New Hampshire. At least three types can be catalogued: (1) an early gray biotite-schlieren bearing two-mica granite (STOP 7); (2) a younger usually pegmatitic pink granite (STOPS 5,6,7); and (3) the "footballs" in biotite or amphibole-rich host rock that are thought to represent a partial melt/restite assemblage (between STOP 8 and VA). Types (1) and (2) are in general conformable to host rock and sheet-like but there are local discordant relationships that will be seen at some of the stops (e.g., 7, between 8 and VA,figure 2). Of the two-mica granites in the immediate vicinity of this trip, the Hooksett Granite and the "footballs" have been dated at 402 Ma (Rb/Sr whole rock; Hayward, 1983; Hayward and others, 1984) and 385 Ma (U/Pb zircon, J.B. Lyons, 1983, personal communication), respectively, and the Milford granite to the southwest has been dated at 275 Ma (U/Pb zircon, monazite, Aleinikoff and others, 1979). Two-mica granite generation and emplacement as semiconcordant sheets within the MGC and MT occurred over a fairly long time span. Most of the granite bodies are relatively undeformed indicating that deformation and metamorphism in the Massabesic Anticlinoriium was pre-Devonian. The Exeter Diorite is also undeformed and suggests that metamorphism and deformation of the Merrimack Group was pre-Middle Ordovician.

The latest intrusive activity in this region is represented by the Mesozoic White Mountain Volcanic-Plutonic Series and younger diabase and lamprophyric dikes. Foland and Faul (1977) report K/Ar ages of 121 and 111 Ma for the alkalic Mt. Pawtuckaway Complex (see Eby, this volume) and Little Rattlesnake Hills Complex, respectively. Diabase dikes throughout the region represent a range of ages between the Permo-Triassic and Cretaceous (McHone, 1978; Public Service Company of NH, 1981).

Metamorphism

Petrologic study of the aluminous calc-silicate rocks and the pelitic rocks confirms the earlier work by Freedman (1950) and Sriramadas (1966) that metamorphic grade increases from east to west and from south to north toward the MGC. Northeast of the area covered by this trip the assemblage quartz + plagioclase + carbonate + chlorite + white mica + biotite is stable in the Berwick Formation; amphibole is not found. The assemblage quartz + plagioclase + calcic amphibole + epidote + biotite composes the aluminous calc-silicate layers that crop out at STOP 0. Following the petrogenetic analysis of Ferry (1976; p. 850 and figs. 4 and 16), STOP 0 is higher grade than rocks to the northeast and the following reactions may be mappable east of STOP 0: chlorite + epidote + quartz = calcic amphibole + plagioclase + H_2O and chlorite + calcite + quartz = calcic amphibole + plagioclase + H_2O + CO_2.

The mineral assemblage in aluminous calc-silicate layers observed at STOP 0 also occurs at STOP 1. The grain size appears to be greater at STOP 1, suggesting perhaps that the metamorphic grade is somewhat higher there. The assemblage plagioclase + quartz + garnet at STOP 1 and 6.3 miles west of the beginning of the trip (see road log for description) indicates temperatures of metamorphism between 500 and 600° C at 2 Kbar and 600 to 700° C at 4 Kbar based on experimental data within the CaO - Al_2O_3 - SiO_2 - H_2O - CO_2 system (fig. 3a). "Dirtying" the system with Na_2O in the plagioclase would decrease the temperature range over which this assemblage is stable, while the presence of FeO + MnO + MgO in garnet would increase the temperature range.

STOP 2 is lithologically distinct from STOP 1 in that pelitic schist is much more abundant [at Stop 2]. However, metamorphic grade may not be appreciably different. The assemblage quartz + muscovite + biotite + garnet + staurolite + sillimanite at STOP 2 indicates temperatures within the sillimanite stability field below the melting of pelitic schist and between the reactions garnet + chlorite + muscovite = staurolite + biotite + H_2O and staurolite + muscovite = sillimanite + garnet + biotite + H_2O (fig. 3b, also see Eusden and others, fig. 6, this volume). (Petrographically staurolite does not appear to be breaking down to sillimanite + garnet + biotite.) As discussed by Eusden and others (this volume) with respect to their figure 6, it is difficult to determine the exact pressure-temperature relations of these assemblages. However, using the Al_2SiO_5 triple point of Holdaway (1971) and the experimental data of Kerrick (1972) for the system K_2O - Al_2O_3 - SiO_2 - H_2O gives the same temperature range as estimated for the calc-silicate rocks at STOP 1 is obtained, i.e., 500° C to no more than 700° C (fig. 3b). The temperature must be below the intersection of the reaction muscovite + quartz = sillimanite + K feldspar + H_2O with the granite melting curve as muscovite + quartz occurs here and the evidence for partial melting occurs farther west.

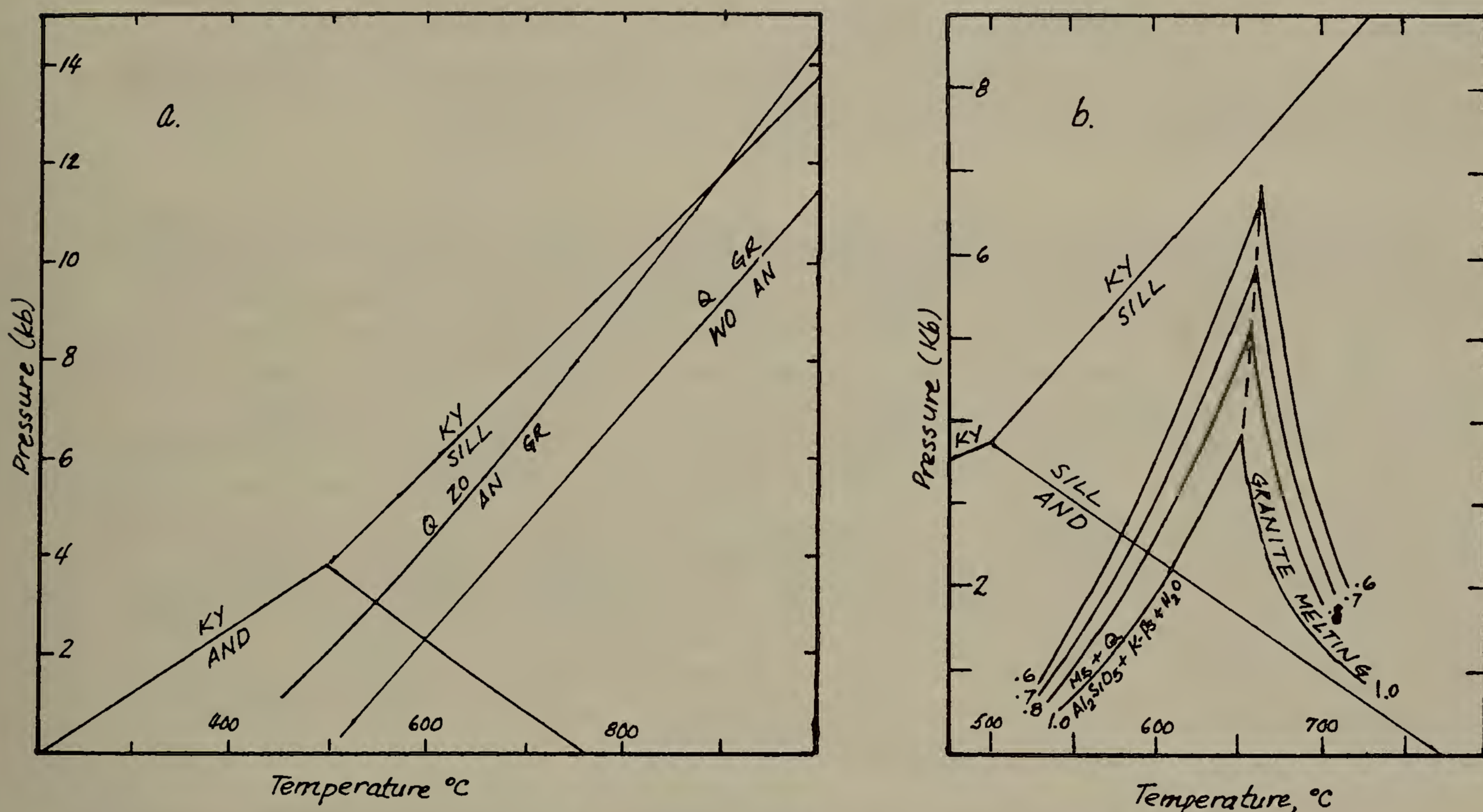


Figure 3. Simple system phase equilibria pertinent to calc-silicate and pelitic assemblages. a) Al_2SiO_5 phase diagram from Holdaway (1971) and P-T relations of the reactions quartz (Q) + zoisite (ZO) = anorthite (AN) + grossulartite (GR) + H_2O and quartz + grossularite = wollastonite (WO) + anorthite. The assemblage Q + AN + GR is stable in between these two reactions. After Perkins and others, 1980. b) Al_2SiO_5 phase diagram from Holdaway (1971) and P-T-X_{H_2O} relations for the reaction muscovite (Ms) + quartz = andalusite/sillimanite + K-feldspar (K-fs) + H_2O and the granite melting minimum. X_{H_2O} is indicated as is the intersections of both equilibria for equal values of X_{H_2O}. After Kerrick (1972). AND = andalusite, SILL = sillimanite, KY = kyanite.

Metamorphic grade increases westward toward the MGC based on the presence of quartz + white mica + biotite + sillimanite without staurolite in pelitic layers at STOP 4 and quartz + plagioclase + clinopyroxene + amphibole + carbonate in calc-silicate layers. The prograde reaction staurolite + muscovite = sillimanite + garnet + biotite + H_2O may be mappable in the pelitic layers between STOPS 2 and 4 (see Eusden and others, fig. 6, this volume). In the calc-silicate layers the diopside isograd (calcic amphibole + calcite + quartz = diopside +H_2O + CO_2, see Ferry, 1976, p. 857, among many others) occurs between STOPS 1 and 4. Further sampling is needed to determine if the diopside isograd occurs west or east of the Flint Hill Fault Zone (FHFZ, STOP 3) east of the MGC.

East of the MGC between STOPS 4 and 5 an increase in pegmatitic material seemingly due to partial melting is seen. The metamorphic temperature estimated for the pelitic rocks at STOP 4 is close to that of the granite minimum, above the staurolite out reaction and below the second sillimanite isograd (fig. 3b; also Eusden and others, this volume, fig. 6). Presence of calc-silicate enclaves within the MGC at STOPS 5-7 with assemblages bearing clinopyroxene + calcic amphibole + plagioclase + quartz ± microcline incidate equilibrium at high temperatures. These data are consistent with the interpretation that some of the MGC rocks may have been formed from partial melting of the Berwick Formation and that some of the enclaves may be xenoliths within the Massabesic orthogneiss. Consequently, late Precambrian metamorphism appears to be recorded in the Berwick Formation as proposed by Olszewski and others (1984).

The petrologic data presented above indicate that there is a consistent increase in metamorphic grade from east to west toward the MGC. A similar increase in metamorphic grade occurs from STOP 13 northwest toward the MGC. At STOP 13 calcic amphibole + epidote + carbonate + plagioclase + quartz ± biotite represent the calc-silicate assemblage. Between STOPS 12 and 13 calc-silicate layers are composed of calcic amphibole + epidote + carbonate + plagioclase + quartz ± biotite and calcic amphibole + orthoclase + carbonate + plagioclase + quartz ± biotite. The following reactions bring epidote and K feldspar into the assemblage, respectively, and depend on fluid composition as well as temperature (see Ferry, 1976, and Hewitt, 1973, among others): calcite + anorthite + H_2O = epidote + CO_2 and biotite + calcite + quartz = tremolite + K feldspar + CO_2 + H_2O. Closer to the MGC, at STOP 12, clinopyroxene + calcic amphibole + plagioclase + garnet + quartz occurs in calcsilicate layers indicating that the diopside isograd occurs on the southeast side of the FHFZ southeast of the MGC.

The formation of clinopyroxene in the calc-silicate layers occurs before the formation of sillimanite + biotite + quartz + muscovite in the pelitic layers (seen at STOP 9) toward the MGC from the southeast, consistent with the correlation Thompson and Norton (1968, tbl. 24-1) have made for calc-silicate and pelitic rocks in general.

Evidence for more than one period of mineral growth is seen at several stops. At STOP 2 grains of staurolite and garnet have inclusion-choked cores and nearly inclusion free rims. Also biotite is being replaced by chlorite. Discontinuously-zoned amphibole with darker-colored cores than rims occurs at STOP 1. Anhedral, embayed clinopyroxene rimmed by calcic amphibole ± carbonate occur at STOPS 12 and 4. Further petrologic and related field and structural studies are underway to discern the relative effects of polymetamorphism versus changes in mineral assemblage/compositon due to progressive metamorphism.

Summary

Field criteria for a gradational contact between the proposed late Precambrian Merrimack Group and the bulk of the Massabesic gneissess are emphasized in the following outcrop descriptions in the road log. Considerable work remains to be done, however, before a full understanding of the rocks of the Massabesic Anticlinorium and Merrimack Trough and concensus on the sequence of events is reached. Isotopic results, however, imply the following geologic framework:

1200 to 650 Ma -	Deposition of the Merrimack Group on unknown basement
650 Ma	Metamorphism and deformation of the Berwick Formation and generation and intrusion by orthogneiss in the area of the Massabesic Anticlinorium

- Deformation, metamorphism, and intrusion

473 to 435 Ma - Intrusion of the Exeter Diorite (and similar diorites), Newburyport Quartz Diorite, and Ayer Granite

405 to 275 Ma - Intrusion of two-mica granites, probably accompanied by migmatization and partial melting of country rock in many areas (e.g., Milford granite and Massabesic paragneiss)

225 to 100 Ma - Intrusion of the White Mountain Series and Mesozoic diabase and lamprophyre dikes

The history outlined above is significantly different from other nearby areas in southeastern Massachusetts and central New Hampshire. We anticipate and welcome discussion at the outcrop!

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Road Log for Trip B5

Access to the new section of Rt. 101 from Raymond to Candia has been made possible by permission from the Midway and Palazzi contractors. We travel and examine road cuts AT OUR OWN RISK. For this trip we have tentative approval to stop along I-93 in the vicinity of Manchester, NH, and access roads to I-93. Do not stop at other times without contacting the NH Dept. of Highways, Concord, NH.

The assembly point is a large parking lot (Gulf Station) at the intersection of NH Rts. 125 and 47 (old 101 from Raymond) in Epping, NH, just east of the Lamprey River Bridge. Actual mileage may be different from Raymond to Candia than that reported here and will depend on work progress on the new highway. There will be several opportunities to reset odometers (*).

Mileage

<u>Epping 7-1/2-minute quadrangle</u>

0.0 Gulf Station parking lot; junction Rts. 125/47, Epping, NH; proceed S on 125

1.4 Rt. 101W turn right toward Raymond, low outcrop of Berwick on northside

3.9 <u>"STOP 0"</u>: S Side 101, long continuous crop of typical Berwick Formation (Stop 2 of Sundeen in the 1971 NEIGC Guidebook in which he considered this the Eliot Formation). The crop consists of brown weathering well bedded, purplish-gray medium-grained biotite + plagioclase + quartz granofels and thin, green calc-silicate layers composed of quartz + light green calcic amphibole + plagioclase + epidote + sphene + carbonate + pyrite. Rusty pelitic schist occurs as thin interbeds and layers locally, as do very minor pinky brown weathering carbonate-rich layers. All are tightly folded about shallow northeast plunging axes. Axial surfaces are filled with quartz and quartzo-feldspathic veins. A Mesozoic porphyritic diabase dike, presumably from the nearby Rattlesnake Hills Complex, cross-cuts the Berwick Fm. here. No stop this year, but an instructive crop for future wanderings.

4.1 If still here, with all the construction. Crops of Freedman's microcline granite, Berwick Formation, and dikes from the Rattlesnake Hills (Mz, White Mountain Series volcanic center, 111 Ma K/Ar, Foland and Faul, 1977)

Mt. Pawtuckaway 7-1/2-minute quadrangle

5.5 Turn right at lights off exit from 101 construction onto Rt. 107/156

6.3 Left at lights on old 101 (107, 156) toward Raymond and Manchester, low crops just before lights and a few hundred meters E are Berwick with calc-silicate layers carrying quartz + plagioclase + calcic amphibole + garnet + biotite + sphene + opaque.

6.6 Turn left at fork (Cumberland Farms Store) and into the town of Raymond

7.4 Raymond Common, turn left on Main Street

*8.2 Left between underpass and up onto new Rt. 101 and continue E about 0.1 mi to crops.

8.3 STOP 1: Berwick Formation is exposed on both E and W bound lanes and perhaps best in the median strip. Fresh purple brown biotite granofels (quartz + biotite + chlorite + plagioclase + sphene) composed of frequently laminated 10-20 cm beds intercalated with thin pelitic schist "partings," and 3-5 cm green calc-silicate (calcic amphibole + quartz + plagioclase + biotite + chlorite + sphene ± garnet) frequently as discontinuous stringers, pods, and occasional boudins. Tops have not been observed here; rocks dip steeply southeast and are cut by pegmatite carrying muscovite books + garnet + beryl + fluorite (S-side and median strip). We consider this typical Berwick. It is essentially the same as the crop at "STOP 0."

8.4 Turn vehicles around and head W, carefully recross Main Street to continue on the new highway.

8.9 STOP 2: Pelitic schist (Gove Member of the Berwick Formation$\underline{=}$Gonic in the Rochester, NH area) crops out on both sides and the median strip as shiny, silvery gray, medium to coarse-grained biotite + muscovite + quartz + staurolite + garnet + minor sillimanite schist. Biotite is aligned and plunges about down dip; larger (3-5 cm) staurolite grains, occasionally twinned, are cross-foliate. Perhaps "staurolumps" is an appropriate term (☺). Late crenulation cleavage plunges gently SW. Minor biotite granofels and calc-silicate is exposed at the east end of the crop.

9.5 STOP 3: West end of Batcheldor Road exchange, brief stop. Flint Hill FZ crosses the highway about here. Dark green sulfidic and brecciated biotite granofels of the Berwick. Small clasts in much silicified microbreccia may be seen locally in outcrop. Milky quartz typical of the Flint Hill FZ is exposed here. Undated, light gray, fine-grained, two-mica garnet-bearing granite is exposed just west of brecciated Berwick. On the exit ramp White Mountain series dikes cross-cut the granite and perhaps the FHFZ. Very sheared metasedimentary rocks characterize the fault zone on the ramp and may reflect later movement than that responsible for the silicified microbreccia.

*Depending on construction conditions we will either proceed about 1.8 miles directly on the new highway, or exit at Lane Road.

9.8 Turn right onto Lane Road

10.0 Left at Lane Road/Batcheldor Road

10.1+ Left at Lane Road/Old Manchester Road and again left onto Green Road

12.0 Right, back onto new 101

12.0 STOP 4: Crops just west of Green Road/New 101 and west of the FHFZ, about 0.5 km east of contact with Massabesic gneiss. Steep southeast dipping, medium grained, biotite granofels (biotite + plagioclase + quartz) in 10-20 cm layers with 1-2 cm, light green,quartz + plagioclase + clinopyroxene + calcic amphibole + sphene ± carbonate ± garnet ±calcite calc-silicate layers of the Berwick Formation, rusty and nonrusty weathering quartz + sillimanite + biotite + white mica ±garnet pelitic schist (a few very large sillimanite bundles near pegs, please don't hammer upon them) dominates the western part of the crop. Glacial pavement atop the median strip and along the S-side of the east bound lane reveal probable tops, cross-cutting biotite granodiorite, and thick diabase.

Candia 7-1/2-minute quadrangle

12.8 Possible stops between here and STOP 5 will depend on time and interest. The crops for the next two miles are dominantly gray foliated biotite + hornblende + quartz + plagioclase + microcline gneiss and is cross cut by pink coarse-grained quartzofeldspathic gneiss with large (4-5 cm) microcline grains, minor garnet, and some magnetite and biotite. Foliation in the gray gneiss is parallel to that of the Berwick at the last stops, contains abundant small tight to isoclinal folds at scales of 10's of cm's, and dismembered layers of biotite quartzite and granofels. Much of this rock is considered to be paragneiss. We recognize no major dislocation between STOPS 4 and 5.

14.6 STOP 5: "The Big Bend" low near continuous median crop dominated by gray biotite quartzofeldspathic gneiss with relict quartzite layers slightly transverse to foliation and clinopyroxene-bearing calc-silicate xenoliths to 25 cm. Interfoliated, undated pink granite (microcline granite of Freedman, 1950?) is slightly discordant. Both are cross-cut by coarse-grained pink pegmatite. Foliation is flat to moderately NW dipping providing some evidence for a structural arch for the Massabesic Anticlinorium.

Depending on road conditions we will exit the construction area and rejoin old 101 between Patten Hill Road and the end of construction or drive directly to the end of the new road and proceed W on 101 toward Manchester.

*0 Start new mileage at the Auburn/Candia town line

Manchester North 7-1/2-minute quadrangle

4.5 STOP 6: 0.2 mi east of Rt 101/I-93 exchange, park as far off the road as possible. We will examine only the crops along the north side of the westbound lane. This outcrop is dominated by medium- to coarse-grained, light gray quartz + plagioclase + perthite + biotite + magnetite ±white mica gneiss (orthogneiss) containing magnificent rotated xenoliths of light gray, medium-grained, biotite + quartz + plagioclase ±magnetite ±garnet ±white mica ±pyrrhotite gneiss with biotite rich borders. Flow foliation in the orthogneiss, particularly around the xenoliths, is very well devleoped. Enclaves of medium coarse-grained, green gray, banded calc- silicate gneiss here are interpreted to be xenoliths of Berwick Formation. Mineral assemblages within distinct layers include quartz + plagioclase + calcic amphibole + biotite + sphene + zircon, clinopyroxene + epidote + quartz + plagioclase + magnetite + sphene, plagioclase + quartz + epidote + amphibole + garnet + biotite + magnetite + zircon, and clinopyroxene + calcic amphibole. Amphibole and plagioclase contain vermicular inclusions of quartz.

4.7 Stay right onto I-93 N.

5.4 Take Exit 8 to Wellington Road. Parking will be difficult here and it is likely that we will have traffic safety escort.

STOP 7: Wellington Road. Much to see! Exposed here in near continuous three-dimensional cuts are paragneiss and migmatite cut by relatively large bodies of pink and gray two-mica granite and pink magnetite-bearing pegmatite. Light gray quartz + feldspar rock and dark gray amphibole + feldspar rock (resembling diorite) make up the bulk of the migmatite and intergrade with greenish calc-silicate rock. These are highly gradational, perhaps representing partial melt and restite. Two mica granites are mutually cross-cutting; no clear cut age relation has yet been established although the schlieren-bearing granite appears earliest. An anatectic origin from bimodal volcanic or calcareous pelitic-felsic volcanic protoliths will be discussed. A typical assemblage within calc-silicate enclaves is commonly: clinopyroxene + plagioclase + quartz + microcline + biotite + calcic amphibole + sphene.

Continue to Exit 9S

~6.2 Exit 9S, Follow the cloverleaf system. Probably parking on Rt. 28 and walking back along the S-bound exit ramp.

STOP 8: I-93, just north of the 9S Exit. An important stop to demonstrate the presence of banded calc-silicate rock northwest of the Flint Hill (Silver Lake) Fault Zone (and southeast of the Campbell Hill fault system) and to provide a basis for comparison with the Berwick seen and to be seen farther south. Polydeformed thin-bedded calcareous subgraywacke is intercalated with sulfidic carbonaceous metapelite and metagraywacke on the southeast and are probably downdropped by a fault (098,50N) against dark gray popcorn migmatite intruded by two-mica pegmatite dikes on the northwest. The calc-silicate rock is composed of quartz + carbonate + feldspar

+ biotite + diopside (as STOP 7). The rusty weathering rock is probably equivalent to the rusty pelitic schist seen at STOP 4. We suggest this sequence is correlative with the Berwick Formation.

~6.5 Rejoin I-93 and continue S - We will have a slow drive by the next set of crops between STOP 8 and the Wellington Road exit (STOP 7) near the VA Hospital.

8.2 Large two-mica mica granite "footballs" are well displayed in the median strip. These are thought to represent insitu melts "frozen in ascent" and may correlate with the 385 Ma (U/Pb zircon) dated footballs on I-93 near Windham (J.B. Lyons, oral communication, 1983).

<u>Manchester South 7-1/2-minute quadrangle</u>

11.4 Continue S on I-93 to I-293. Depending on time we will EITHER follow I-293 (1.6 mi) to Exit 1 (Rt. 28) S of Manchester. Follow Rt. 28 S about 0.8 mi, turn right onto Harvey Road for 0.5 mi. and right into the parking lot of True Value Hardware Distributors.

<u>OPTIONAL STOP 9</u>: If made we will follow Harvey Road S to Grenier Field Road to its intersection with Rt. 28; STOP 10 is 0.2 mi S of this intersection.

This outcrop shows many elements of the MGC migmatite; coarse-grained gneiss, restitic xenoliths, two-mica granite, and pegmatite. The compositional banding of the coarse-grained gneiss is defined by continuous layers of coarse-grained biotite, commonly with abundant sillimanite + white mica, alternating with light gray feldspathic layers. The banding strikes about N50E and generally dips steeply to the southeast. The orientation of the banding is more irregular near the granite and near the xenoliths. The xenoliths are concentrated in the west part of the outcrop. They are so similar in appearance to the surrounding gneiss that it can be difficult to distinguish them from the gneiss. The two-mica granite at the east part of the outcrop clearly truncates the banding of the coarse-grained gneiss, although the contact is somewhat obscured by aplite and pegmatite.

Proceeding south, Harvey Road turns into Grenier Field Road, which curves around a hill to the northeast. On the south end of the hill restitic xenoliths of biotite + plagioclase + quartz schist are intermingled with two-mica granite. The coarse-grained gneiss crops out on the north end of the hill.

<u>OR</u> Continue S (~4.1 mi) on I-93 to Exit 5.

15.5 Exit 5, turn west on Rt. 28 for 1.8 mi.

16.3 <u>STOP 10</u>: park just south of the intersection of Rt. 28 and Page Road. There are several outcrops along this stretch of road. The northernmost of these shows the coarse-grained gneiss of the Massabesic migmatite juxtaposed against weakly pinstriped biotite granofels and schist typical of the more pelitic Berwick Formation.

The foliation of the schist roughly parallels the compositional layering of the coarse-grained gneiss. Calcareous lenses in the schist are contorted about axes plunging moderately to the northeast. Boudin-like lenses of two-mica granite are oriented roughly parallel to the foliation of the schist. Biotite-rich layers from the schist protrude through the granite lenses. Gradational transitions from pegmatite to granite can be observed here.

Derry 7-1/2-minute quadrangle

*18.1 Return to I-93, Exit 5

18.3 STOP 11: South bound entrance ramp I-93/Rt. 28, Londonderry, NH. A very fine-grained, chloritized, broken, and somewhat brecciated outcrop marks the trace of the Flint Hill Fault Zone. Porphyroblasts of amphibole are present as well as some pegmatite.

Continue S

18.8 STOP 12: 0.5 mi S of Rt. 28 (south-bound lane).
The thick-bedded, fine- to medium-grained biotite granofels typical of the more pelitic Berwick Formation has been highly sheared and broken here. The shearing is especially apparent in the more pelitic layers. Fractures may be filled with chlorite. The calcareous lenses here are folded and distinctly zoned with rims of hornblende + amphibole with vermicular quartz + biotite + plagioclase + quartz + sphene ± garnet and cores of diopside + hornblende + amphibole with vermicular quartz + plagioclase + quartz + sphene ± garnet. The diopside and amphibole with vermicular quartz are generally embayed and rimmed by amphibole and may be relict phases. The garnet is isotopic and anhedral. Diopside has not been identified between here and Derry, NH.

Continue S

22.0 Exit 6, Derry, NH

22.2 STOP 13: Southbound entrance ramp to I-93 from Rt. 102, Derry, NH
The fine-to medium-grained, purple-gray granofels of the Berwick Formation is well-bedded here, although poorly graded. Calcareous lenses are poorly zoned and contain tremolitic amphibole, epidote, sphene, biotite, quartz, plagioclase, and calcite. The granofels is interbedded with finer grained, dark gray, moderately rusty schist composed of quartz, plagioclase, white mica, biotite, chlorite, sphene, pyrite, and tourmaline. All of the rocks are folded isoclinally with axes plunging shallowly to the southwest and axial planes dipping steeply to the northwest. Intrafolial fold hinges have the same orientation.

END OF TRIP - Head S to "Headquarters" and dinner (not too late we hope)!

THE GEOLOGY OF THE SADDLEBACK MOUNTAIN AREA, NORTHWOOD QUADRANGLE, SOUTHEASTERN NEW HAMPSHIRE

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Introduction

Saddleback Mountain is a four-peaked saddle-shaped mountain located 39 km west of Portsmouth, NH (Figure 1). The mountain lies along the boundary between the New England Uplands and the Seaboard Lowlands Sections of the New England Physiographic Province of Fenneman (1938). It also lies adjacent to the western boundary of the Massabesic Gneiss Complex as designated by Anderson (1978). Figure 2 of this text is a generalized Geologic Map of Southeastern New Hampshire compiled from Billings (1956) and later workers cited in the references from Figure 1. Portions of Figure 2 show that the Saddleback Mountain area is on strike with two convergent fault zones which are postulated to be of regional tectonic significance: the Nurambego of southern Maine (Hussey and Pankiwskyj, 1976; Hussey and Newberg, 1978), and the Pinnacle-Campbell-Hill-Hall Mountain Faults mapped by Carnein (1976). (See Lyons et.al., 1982). This area may be an example of one of the critical areas referred to by Hamilton and Meyers (1967).

Freedman (1950) mapped the Saddleback Mountain area as an isolated, faulted metamorphic roof pendant of supposedly Littleton Formation (Mid-to Late Devonian, ?) within granitic units of the Fitchburg Pluton of the New Hampshire Magmatic Series, as designated on the 1956 Geologic Map of New Hampshire (Billings, compiler). Recent isotopic studies of previously presumed Devonian granitic units have yielded Precambrain ages (Besancon et.al., 1977, Kelly, 1980): these units lie south and westwards of the Saddleback Mountain area, but belong to the units of the "Fitchburg Pluton", which has been redesignated the Massabesic Anticlinorium by Lyons et.al., (1982). Alienikoff, et. al., (1979) reported Precambrian age dates for units even farther south and west. Still, the structural complexities of the Massabesic Anticlinorium have yet to be deciphered.

The Saddleback Mountain area is an advantageous one to study for these reasons:

1. There are over 600 feet of vertical relief (250m) between the topographic base of the mountain and its top.

2. In comparison to many areas of southeastern New England, there is relatively good exposure.

3. The metamorphic units on Saddleback Mountain are isolated from other metamorphic units along strike, thus allowing treatment of the problem structurally, in a non-stratigraphic context without regional prejudice.

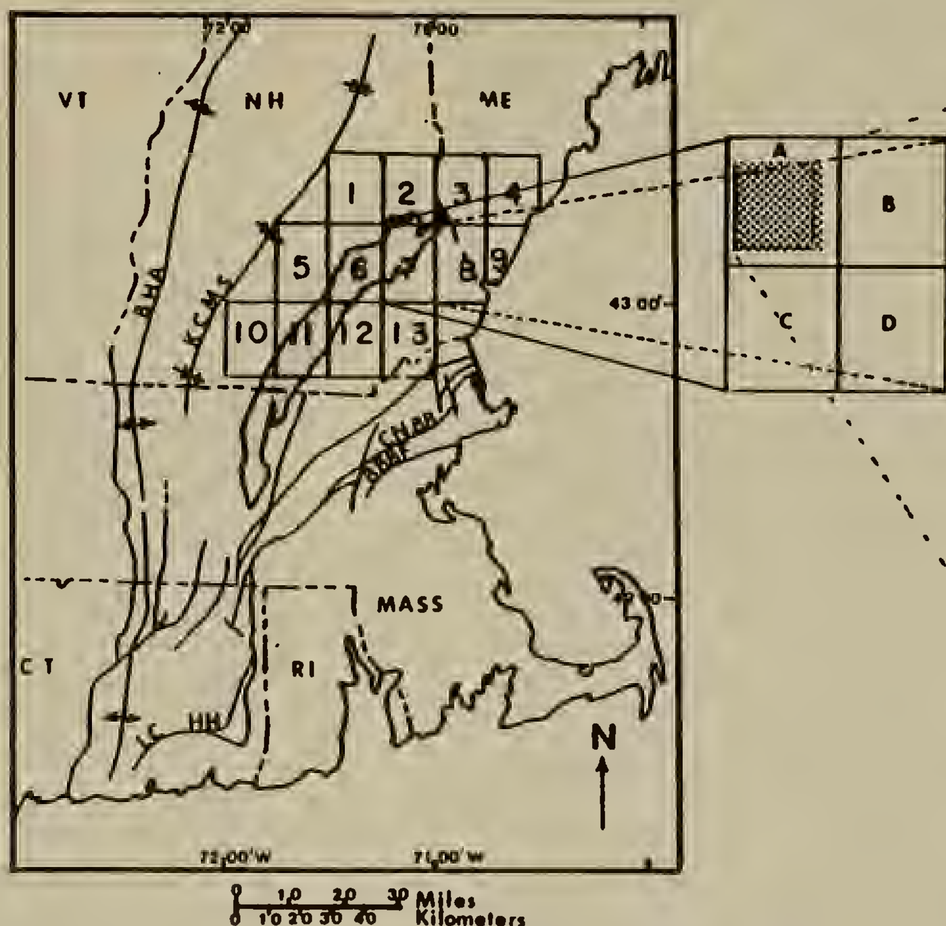


FIGURE I LOCATION MAP WITH REGIONAL STRUCTURAL FEATURES AND QUADRANGLE INDEX

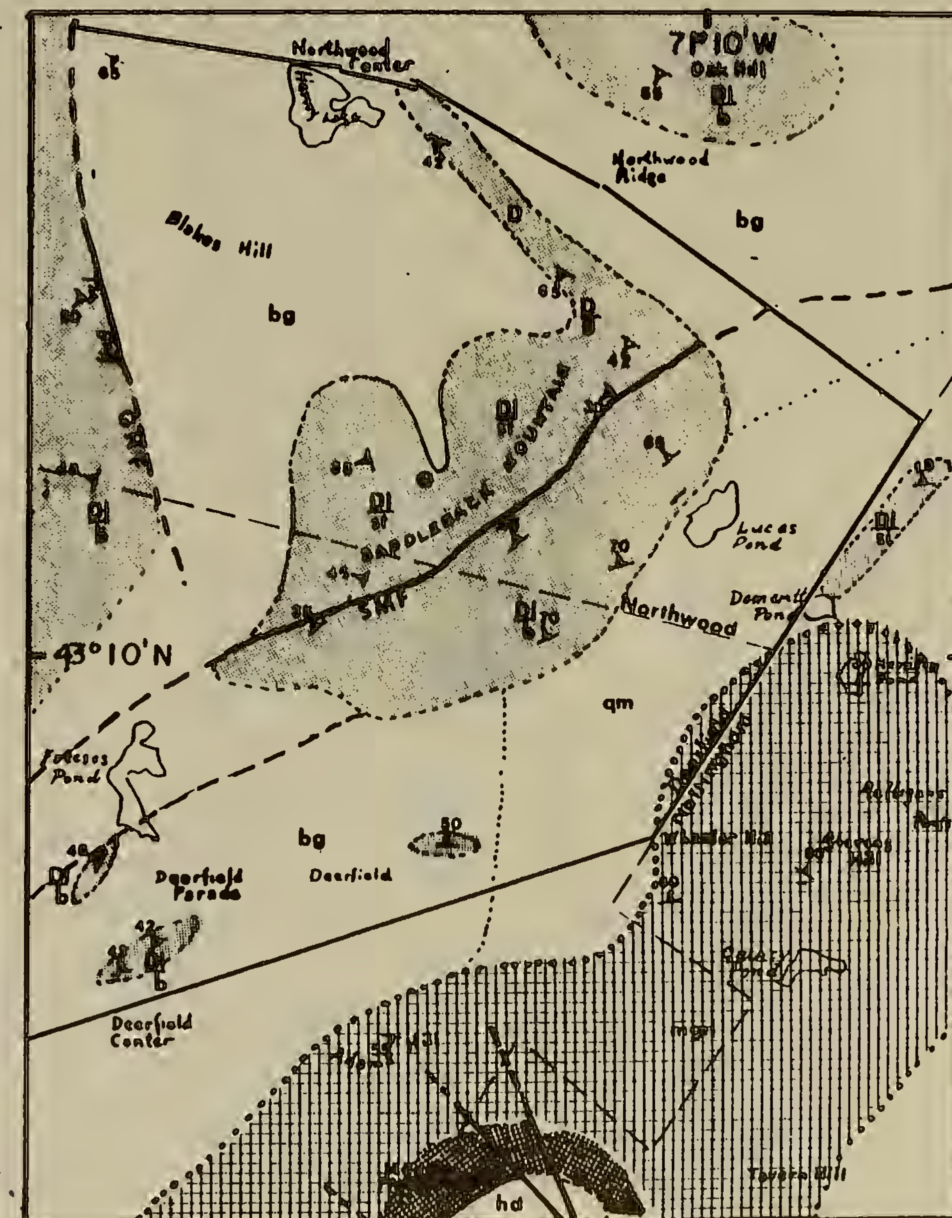


FIGURE IA LOCATION MAP

GUIDE TO FIGURE I

STUDY AREA

MASSABESIC ANTICLINORIUM (FITCHBURG PLUTON)

15 MINUTE QUADRANGLE INDEX

1. GILMANTON (HEALD, 1955)
2. ALTON (STEWART, 1961)
3. BERWICK (HUSSEY, 1962; HUSSEY & PANKIWSKYJ, 1976)
4. KENNEBUNK as above
5. CONCORD (VERNON, 1971)
6. SUNCOOK (CARNEIN, 1976)
7. MT. PAWTUCKAWAY (FREEDMAN, 1950)
   - A. NORTHWOOD 7 1/2
   - B. BARRINGTON 7 1/2
   - C. MT. PAWTUCKAWAY 7 1/2
   - D. EPPING 7 1/2
8. DOVER [NOVOTNY, 1963, 1969]
9. YORK: as 3 above
10. PETERBOROUGH [GREENE, 1970]
11. MILFORD [ALIENIKOFF, 1978]
12. MANCHESTER [SRIRAMIDAS, 1966]
13. HAVERHILL [SUNDEEN, 1971]
14. EXETER [NOVOTNY 1963, 1969]

ABBREVIATIONS

BBBF - BOSTON BASIN BOUNDARY FAULT
BHA - BRONSON HILL ANTICLINORIUM
CNBB - CLINTON-NEWBURY BLOODY-BLUFF FAULT ZONE

KCMS - KEARSARGE CENTRAL MAINE SYNCLINORIUM
LCHH - LAKE CHAR HONEY HILL FAULT ZONE

GUIDE TO FIGURE IA

---- GEOLOGIC CONTACT (INFERRED) BETWEEN 'LITTLETON' FORMATION AND IGNEOUS UNITS

.... GEOLOGIC CONTACT (INFERRED) BETWEEN (bg) BINARY GRANITE and 'qm) QUARTZ MONZONITE

•••••• GEOLOGIC CONTACT BETWEEN 'MASSABESIC GNEISS' AND QUARTZ MONZONITE according to ANDERSON (1978)

FAULT

FAULT INFERRED

HM-CHF = HALL MOUNTAIN-CAMPBELL HILL FAULT (as projected by ANDERSON, 1978 from CARNEIN, 1976)

GHF = GULF HILL FAULT

SMF = SADDLEBACK MOUNTAIN FAULT

ATTITUDE OF 'BEDDING' AND FOLIATION as mapped by FREEDMAN, 1950

ATTITUDE OF FOLIATION

LITTLETON FORMATION

$\frac{Dl}{b}$ = BIOTITE GRADE

$\frac{Dl}{st}$ = STAUROLITE GRADE

$\frac{Dl}{sl}$ = SILLIMANITE GRADE

mgn = MASSABESIC GNEISS

MOUNT PAWTUCKAWAY SUITE

COARSE GRAINED MONZONITE

hd HORNBLENDE DIORITE

BOUNDARY OF STUDY AREA

FIGURE 2: GEOLOGIC SKETCH MAP, S.E. NEW HAMPSHIRE

MODIFIED FROM BILLINGS (1956) AND LATER QUADRANGLE MAPS.

GUIDE TO MAP SYMBOLS

- **1–14** 15' Quadrangles as Indexed in Figure 1
- **15** ISLES OF SHOALS (Fowler-Billings, 1977; Bloomshield, 1975)
- Coastline
- ME/NH Stateline or territorial boundary
- Formation boundary after Billings (195)
- Formation boundary sketched following quadrangle maps
- Physiographic section boundary
- Silicified zone or fault
- Town or city
- State capital
- Anticlinal Axis
- Synclinal Axis
- Axis of Major structure

METAMORPHIC ISOGRADS

- —si— Sillimanite
- —st— Staurolite
- —g— Garnet
- —b— Biotite
- —c— Chlorite

SCALE

0, 5, 10, 15 mi

0, 8, 16, 24 km

GEOLOGIC FORMATIONS OR LITHOLOGIES

a. Igneous

- mgn Massabesic Gneiss
- g/big Granite/Two Mica Granite
- kqm Kinsman Quartz Monzonite
- nqd Newburyport Quartz Diorite
- agd Ayer Granodiorite
- qd Quartz Diorite

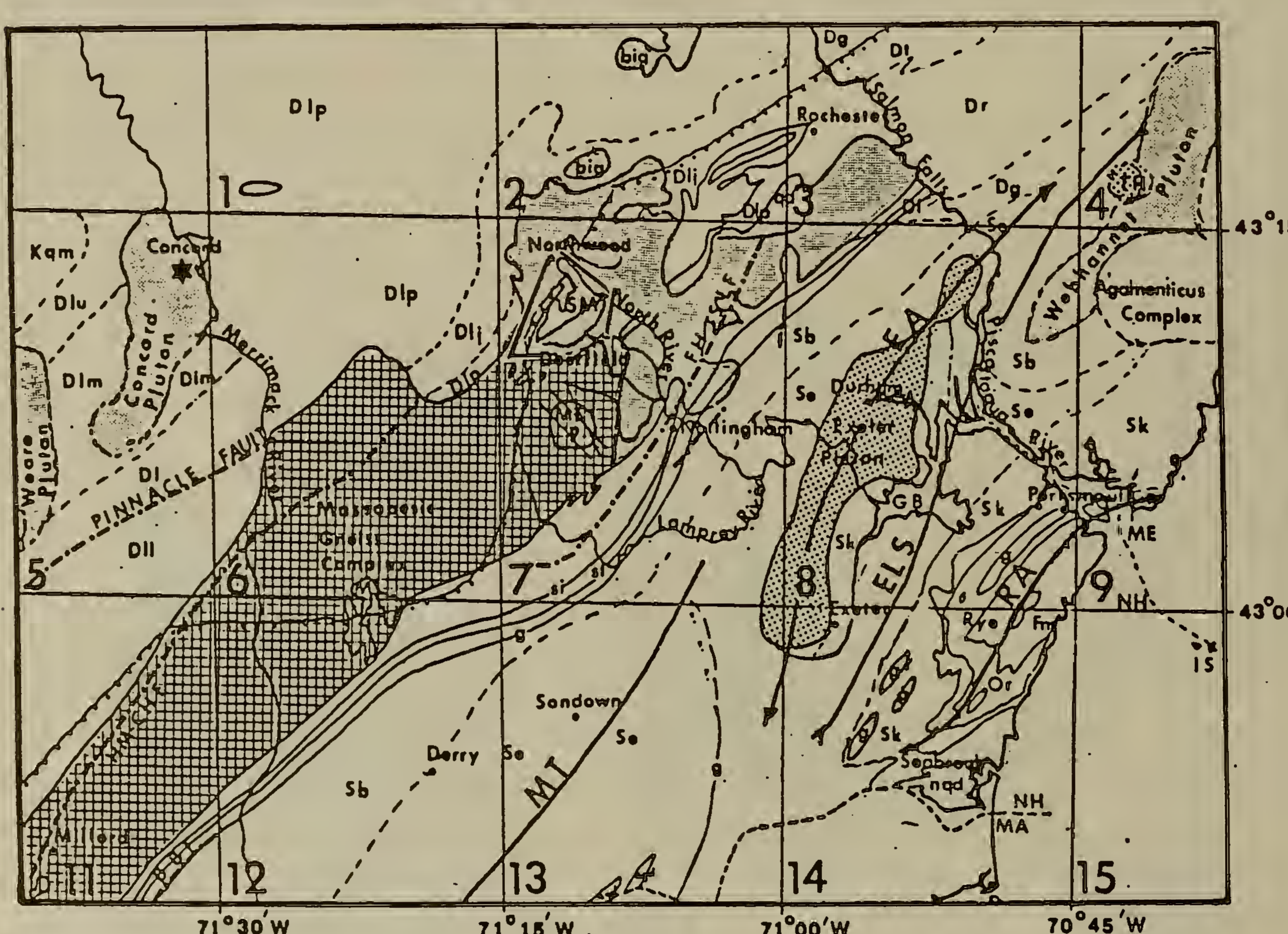
LITHOLOGIES (continued)

b. Metamorphic

- Dg,t,r Shapleigh Group } SW Maine
- Dl... Littleton Formation } SE, Central NH
- Sb Berwick Fm } Merrimack Group
- Se Eliot Fm } Merrimack Group
- Sk Kittery Fm } Merrimack Group
- Or Rye Formation

ABBREVIATIONS

- EA Exeter Anticline
- ELS Eliot Syncline
- MT Merrimack Trough
- RA Rye Anticline
- BBF Beaver Brook Fault
- FH-SLF Flint Hill - Silver Lake Fault
- HM-CHF Hall Mountain-Campbell Hill Fault
- MT P Mount Pawtuckaway
- SM Saddleback Mountain
- TH Tatnic Hills



4. The area serves as a test area for the possibility of through-going faults connecting with others mapped previously: i.e., Hussey and Pankewskyj, 1976; Hussey and Newberg, 1978; Carnein, 1976.

A study of Freedman's (1950) Geologic Map of the Mount Pawtuckaway Quadrangle indicated that alternative hypotheses to his interpretation of the structure of the area might be viable. The following hypotheses were maintained throughout the field mapping process:

1. The model provided by Freedman (1950) of a metamorphic roof pendant to the Devonian intrusives lying along a drag fold on a structural terrace on the eastern limb of the Merrimack Synclinorium.

2. In light of the isotopic studies to the south, the metasedimentary mass might rest nonconformably on older crystalline basement rocks, with the map pattern produced as a function of topographical surface in relation to the dip of the units.

3. The metasedimentary lithology might be a screen in a large granite-migmatite complex.

4. The metamorphic outcrop pattern is a reflection of a horst or graben structure.

5. The metamorphic mass might represent a klippe, or erosional remnant of a larger decollement surface, or nappe structure.

6. The pattern might represent a basin and dome interference pattern between major folds.

These hypotheses were carried through, or modified throughout the mapping process.

The geographic area of concentration for detailed mapping was the immediate vicinity of Saddleback Mountain which lies diagonally across the coordinate grid in a N55 to 60 degrees E trend, encompassing an area of approximately 37 km^2. The reconnaissance mapping in the rest of the Northwood Quadrangle, the Mt. Pawtuckaway, and Suncook Quadrangles to the south and west encompassed approximately 102 km^2.

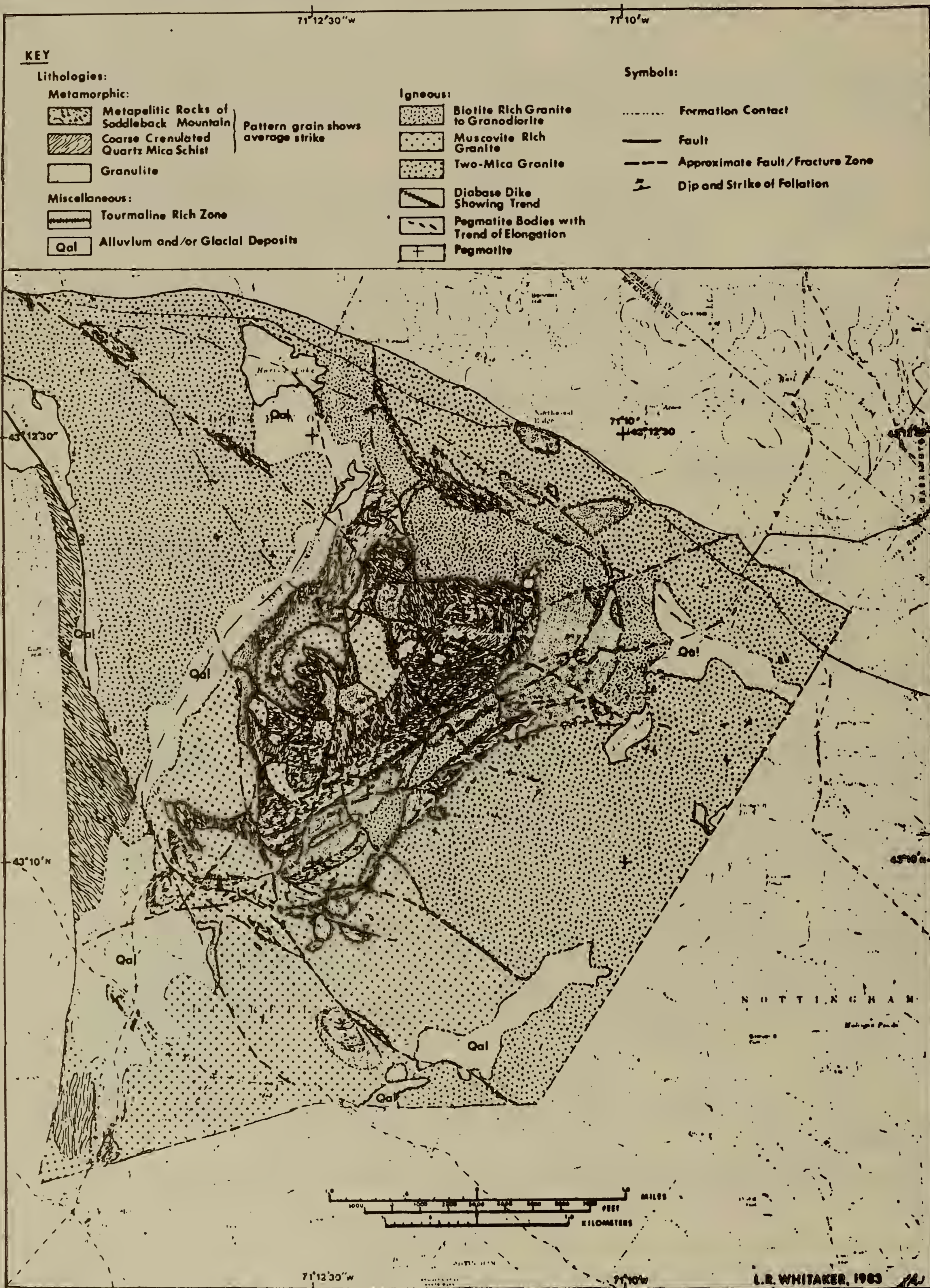
Transposition structures and tight minor folds were recognized early in the mapping process, indicating that the dominant foliation trend seen in outcrop was a second-order or higher feature. The mapped area was divided into five structural domains based upon the nature of the dominant foliation trend and other fabric elements for that geographic area. (See Plates 1-3).

Geology

Lithologies:

Igneous and metamorphic rocks are exposed in about equal portions (Plate

PLATE 1: GEOLOGIC MAP OF SADDLEBACK MOUNTAIN, NORTHWOOD QUADRANGLE, N.H.



1, Geologic Map). Igneous units tend to be more poorly exposed, and to create a more rounded topography than areas underlain by metasedimentary rocks. In exception to this, pegmatite bodies commonly crop out as knobs above the surrounding topography.

The igneous lithologies are subdivided into a felsic to intermediate category which is in turn, subdivided and, a mafic one which is represented by basalt dikes.

The metasedimentary lithologies are divided into two broad categories: granulites and metapelites. The granulitic units are interlayered with more schistose units, and sequences in which the granulitic layers predominate over metapelitic ones will be termed granulites. In this report the term granulite is used in its textural context and not as an indicator of metamorphic facies.

The granulitic metasedimentary rocks crop out on the east and south sides of the mountain along a northeast-southwest trending band which narrows in the central zone, then fans southwards. Exposure width of the band inceases also northwards with decreasing elevation. Drill chips from a well on the northeast flank of the mountain confirm the presence of granulites at depth in an area where they are not well exposed at the surface. The elevation of the exposures of granulites on the east side of the mountain ranges between 470 ft. (145m) to 700 ft. (215m). On the upper reaches of the mountain granulites are found within migmatites between the elevations of 800 ft. (246m) and 1000 ft. (308m). Exposures along the western flank of the mountain crop out between 700 ft. (215m) and 1000 ft. (308m). There layering is nearly horizontal, or gently dipping southwestwards, and occasionally eastwards. Units on the west flank of the mountain are commonly interlayered with and overlain by felsic igneous sills, or grade texturally into tough crystalline gneissic feldspathized schists.

The lithologies of the granulites are: calc-silicate granulite; quartz-biotite-plagioclase-microcline granulite; quartz-biotite schist with minor muscovite; impure quartzite; and tourmaline-bearing granulitic gneiss.

Outcrop exposure is rarely continuous for more than 65m along strike, and tends to be along joint surfaces sub-parallel to the strike of the foliation, but steeply dipping in the opposite direction to the direction of the dip of foliation. Outcrops tend to be low-lying and knoll-like; commonly there are relicts of pegmatite bodies, or felsic to intermediate sills overlying granulitic metasedimentary exposures.

Layering ranges from 1 cm to 10 cm, with occasional 10 cm layers. The average thickness is 5 cm for the calc-silicate-bearing units, and 2 cm for the flaggy and more schistose units. Grain sizes range from .3 mm to 1 mm, and microscopic textures range from protomylonitized (Higgins, 1972, p. 15) to annealed (Spry, 1969, p.222). The samples which show annealing tend to have a lower percentage of opaque minerals, while cataclastic specimens show alignment of opaques parallel or subparallel to the dominant fabric, or have a large number of bubbles, possible fluid inclusions and dustings of opaque minerals within grain boundaries.

Calc-silicate-bearing units may constitute as much as 30 percent of the granulites. The calc-silicate minerals are both disseminated in the matrix and contained in pods and bands. These pods and bands may appear as elongate lenses ranging from 5 cm to 1 cm or more in length. Pods also commonly range in size from 3 cm to 15 cm along the long ellipsoidal axis, and 3 to 7 cm along the short axis. They are commonly observed to occur at the juncture of two healed fracture systems or to be associated with fold patterns in the rocks. These will be seen at Stops 3 and 7. Thin sections of calc-silicate pods show high strain features such as undulatory extinction and ribbon texture of quartz, crushed grain boundaries of amphibole or pyroxenes and plagioclase as well as dislocations of plagioclase twin planes.

The metapelitic rocks crop out in a sliver-like band along the east and south sides of the study area between approximately 400-500 ft. (123-154 m) elevation in the northeast, and approximately 600 ft (185m) in the south. (See Geologic Map). The bulk of the metapelitic rocks are exposed above the 800 ft. (246 m) contour, intermingled with small patches of granulite contained within coarsely migmatitic micaceous quartzo-feldspathic matrix. Exceptions to this are small discontinuous exposures of granulities inter-layered with sills of two-mica granite. Because of the elevation discrepancies between the two areas where metapelites are exposed, those below 600 ft. (185m) will be termed the lower metapelitic sequence, and those above 600 ft. will be termed the upper metapelitic sequence.

In contrast to the granulites, the metapelites show positive and sharper topographic relief. Exposure is often continuous for long distances along strike along the scarp-like en-echelon surfaces of "master joints". Excellent exposure of the metapelites is found on the upper reaches of the south peak of Saddleback Mountain opposite the WENH microwave tower. Exposure of the lower metapelites was not readily available at the time of Freedman's field work. Exposure has since been made available by road cuts and excavations. Good exposure is available at the Camp Yavneh septic lagoon in Northwood, N.H., and at the junction of Coffeetown Road and N.H. Route 43 in Deerfield, N.H.

There is a high degree of internal deformation within the metapelites, with a wide variety of fold styles exhibited at different scales.

Layering within the metapelites ranges from 2 mm to 5 cm or more in thickness. The rock texture is generally medium to coarse grained, in contrast to the generally finer grained granulites. Porphyroblasts of staurolite, tourmaline, and/or garnet commonly lend the rocks a knotty texture following weathering of the outcrop. Layers composed of quartz and tourmaline only are common in distinct sections of the sequence. Besides these layers, compositional layering of the rocks is delineated by staurolite and/or garnet-rich zones, layers of intergrown micas and quartz, quartzo-feldspathic rich layers, or layers rich in sillimanite. Well-indurated quartzite layers up to 10 cm thick are traceable in some areas. Strong cataclastic rodding of quartz, crenulation of kink axes in micas, and sillimanite needle alignment are common lineations present in the metapelites.

The rock colors vary depending upon the relative amount of biotite present. Fresh surfaces tend to be shiny light silvery grey to mottled white to bluish black, depending upon the grain size and width of compositional banding. Sillimanite bearing lithologies have a characteristic satiny sheen. Staurolite and garnet may give the rocks a pinkish tinge. Weathering colors range from chocolate brown with yellow tinge to rusty-to-black. Micas cause the golden yellow weathering. Mineral assemblages are shown on Thompson AFM diagrams in Figure 3. Noteworthy textural characteristics are shown by several metamorphic minerals: garnet, sillimanite, staurolite, and opaques. Garnets found in the metapelites are commonly poikilitically embayed, fractured, and flattened in the plane of the dominant foliation. Sillimanite occurs in both fibrolitic and tabular forms, with some sillimanite rich rocks showing both forms together. The fibrolitic form of sillimanite tends to be concentrated in shear zones. Less sillimanite rich rocks tend only to contain fibrolite. The rocks exposed by the WENH Microwave tower contain muscovite whose texture is suggestive of replacement of sillimanite by muscovite, then regrowth of sillimanite. Rock staining indicates minor potassic feldspar in these rocks. Staurolite is almost without fail poikilitically intergrown with quartz, and is often contained preferentially in discrete, folded layers within the rocks. The crystals of staurolite commonly define lineation by preferred orientation, or define fold noses of minor structures. Opaque minerals are noteworthy because they may often comprise a major phase of the rock, or more than five modal percent. The opaques may outnumber other minor phases of the rock sample, even if they are not a major phase.

Two generations of biotite are documented from metapelitic samples, as well as two distinct populations of sillimanite.

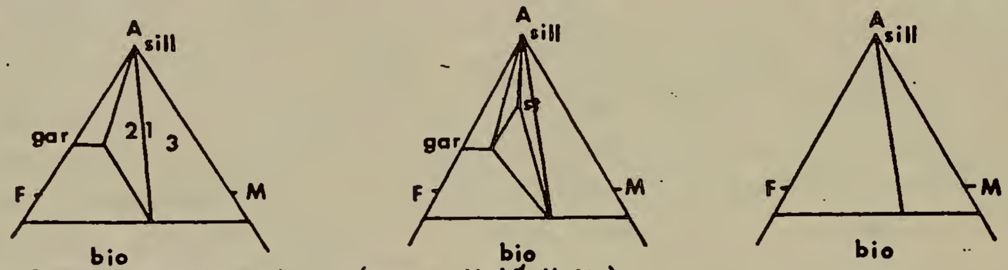
Significant portions of the upper metapelitic sequence are migmatitic, as are the rocks of the Camp Yavneh Lagoon. Portions of the quartzo-feldspathic material occur as veins and pods, and there is a gradual increase in quartzo-feldspathic content of the granulitic metasediments, along with a coarsening of fabric until the granulite is engulfed in a sea of more mobile matrix and left as a raft, or scholle to use the terminology of Mehnert (1968).

The mineral assemblages in the metapelitic rocks are compatible in metamorphic grade with the granulitic rocks. The stable mineral assemblages of both groups place them in the upper amphibolite facies of metamorphism (Eskola, 1920). Staurolite persists in the metapelites, albeit in segregated lenses throughout the sequence. This indicates that the P-T conditions of the rocks were sufficiently high enough to produce sillimanite in coexistence with potassic feldspar, but were below the staurolite breakdown temperature. These conditions would be fulfilled at about 4 Kb pressure, and between 550 and 600 degrees centigrade (Labotka, 1978).

There are relict minerals which indicate the rocks have undergone multiple periods of metamorphism. These are:

1. The overgrowth of diopside by amphiboles, and amphiboles on each other, to include core-rim reactions.

FIGURE 3: MODIFIED THOMPSON AFM DIAGRAMS FOR METAPELITIC MINERAL ASSEMBLAGES



Projections from Quartz, Muscovite (approx Mol% Units)

1. Quartz, aligoclase, albite, biotite, muscovite, sillimanite, opaques
2. Quartz, muscovite, biotite, garnet, sillimanite, opaques
3. Quartz, biotite, sillimanite, aligoclase, muscavite, chlarite (late), tourmaline, opaques
4. Quartz, biotite, muscovite, garnet, stauralite, chlarite (late), opaques,
5. Quartz, biotite, muscovite, sillimanite, garnet, staurolite, opaques, tourmaline, chlorite
6. Quartz, muscovite, biotite, opaques, garnet, staurolite
7. Quartz, aligoclase, muscovite, biatite, chlarite, apaques
8. Quartz, garnet, muscavite, biatite, sillimanite, opaques, microcline
9. Quartz, biotite, sillimanite, muscavite, oligaclase, garnet, opaques, microcline
10. Quartz, microcline, muscovite, biatite, stauralite, garnet, opaques

(8–10) Not projected

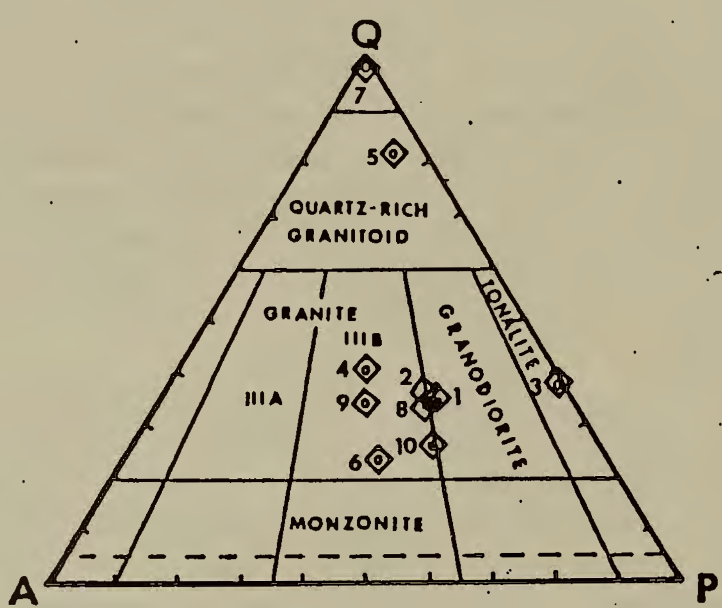


FIGURE 4
I.U.G.S. CLASSIFICATION OF FELSIC IGNEOUS ROCKS
(STRECKEISEN, 1979)

SAMPLES PLOTTED: SEE Whitaker, 1983

1. 666	5. 7189-1	9. 105G2
2. 62179-1	6. 7199-2	10. 105sso
3. 7679-2	7. 889-2	
4. 7179-1	8. 105G1	

2. The presence of two populations of sillimanite, and possibly two generations of garnet in the metapelites.

3. Two generations of biotite in both metapelites and granulites: in some granulite samples grains are crossed, and extinct at distinctly different stage orientations, while the different biotite generations in the metapelites show differences in crystal habit, inclusion pattern and indices of refraction.

4. The flattening and embayment of garnet, with alteration of the rims to biotite in the metapelites.

Items 3 and 4 above indicate that metamorphism was accompanied by deformation.

The igneous rocks are predominantly felsic to intermediate. Mafic rocks are represented by dikes, ten of which have been positively located. The divisions of the felsic to intermediate rocks are:

1. Two-mica granite
2. Biotite rich granite-to-granodiorite
3. Muscovite rich granite
4. Pegmatite
5. Tonalite.

These lithologies are gradational with each other, but one type will tend to predominate over the others in a given domain. Muscovite-rich granites are dominant in the Southern Domain, but they grade to tonalites and biotite-rich granite. The biotite-rich granite to granodiorite predominates in the Northern Domains and the Northeastern Domain, and exists in narrow discontinuous bands along the eastern side of the study area parallel to the Saddleback Mountain Fault Zone. An increase in the biotite content of the igneous rocks, particularly in the eastern margin of the area is a common precursor to the discovery of a small enclave of granulitic metasediment, or a scholle. The only cross-cutting relationships between rocks of the felsic to intermediate category are of pegmatite dikes crossing two-mica granite, or biotite rich granite-to-granodiorite.

Figure 4 is a plot of the pointcounted normalized samples on the I.U.G.S. Q-A-P Triangle (Streckeisen, 1979). The majority of these samples plot in the granite field where they cluster near the granite-granodiorite boundary. One sample plots as a tonalite: the remaining two samples which are muscovite rich granite and pegmatite plot in the silexite, or quartz-rich granitoid field.

The mineralogy of these rocks indicates that they are peraluminous, as biotite, muscovite and garnet with accessory tourmaline are the only femic constituents.

The essential minerals are: plagioclase (var. oligoclase, An 20-26), microcline or orthoclase, quartz and micas. The ratios of the micas one to another is variable, but muscovite and biotite are ubiquitously intergrown. Myrmeckite and perthite are common, with myrmeckite being particularly common in more highly deformed zones of the rock. Garnets are common accessories, often occurring in two populations: one pink fine-grained (1-2 mm) cinnamin colored; the other darker, and larger (2-5 mm). Characteristic accessories are: apatite, sphene, zircon, opaques, and an amorphous red adiamantine-reflecting fracture filling which may be responsible for lending the rocks a pink color in outcrop.

Foliation is evident in almost all outcrops. Cataclastic features such as flaser texture, or dents-du-cheval may also be present. Ghost structure, or relict foliation (Pitcher in Newall and Rast, 1969) is evident in many outcrops, especially along the northern boundary of the study area. Contacts between igneous rocks and the metasediments indicate that the metasediments behaved in a ductile manner during the emplacement of the igneous units: the igneous rocks themselves have features indicating post emplacement shearing. Contacts between granodiorite and pelitic schist are less commonly seen in the field than contacts between granulite and felsic-to-intermediate lithologies: i.e., the muscovite-rich granite is commonly associated with the metapelites, or pegmatites.

There are over 100 bodies of felsic pegmatite exposed throughout the area. These tend to be oblately elliptical lenses ranging from 1 m in length to over 30 m, and lying subparallel or parallel to the local strike of the host rocks. They are commonly responsible for the localized steepening of dip, and/or deflection of the foliation trend. They may also bottom out as detatchment zones for the metasediments. Pegmatites occur within the metasediments, commonly along the contact between the metapelites and granulites, and within the igneous bodies, commonly close to or along the contacts with the metasediments. The pegmatite-metasediment contact may be the only portion of the contact exposed. Larger pegmatites 30 m or more in length crop out as knobs 3-15 m high above the surrounding topography, particularily along the eastern boundary. Many pegmatites are simple, but complex mineralogies occur, notably in the contact zones, showing high tourmaline or apatite content, as well as beryl. Borders between the pegmatites and country rock may be gradational, with gradual coarsening of the fabric of the host rock, an increase of micas in the host, and the presence of tourmaline laths, or zones of intergrown micas, staurolite or garnet. Structures such as slickensides, or rodding and mullion structures are common along the contacts. These contacts tend to show a higher percentage of opaques than the metasediments farther from the contacts. Tourmaline-rich zones occur near the contacts between the metasediments and the pegmatites, and in definite zones in metapelitic units, and in granulites near the Saddleback Mountain Fault Zone.

Ten dikes of dark-gray, red-brown weathering, slightly porphyritic basalt have been located. They range in width from .3-.6 m to an estimated 22 m width. Basalt float on the northwest side of Saddleback Mountain indicates that more dikes are present at depth, but these erode away preferentially, and are not represented in outcrop in direct proportion to their presence.

Quartz veins and stringers are ubiquitous in all metasedimentary and felsic igneous rocks of the area. Quartz segregations in the granulites tend to be ptygmatic and rolled, dimensionally on the order of 2-3 cm thick to 8 cm. These are commonly aligned en-echelon parallel or subparallel to the strike of the dominant foliation. Large crosscutting veins of commonly diffusion banded quartz occur within the metapelites and granites, often near the contacts between the two. These veins are usually fractured and commonly off-set.

Structures:

The pattern of the Geologic Map (Plate 1) indicates that the area shows alternation of igneous and metamorphic units with respect to the topographic surface. Plates 2, 2A and 3 provide structural information broken down by domains.

There are many minor folds within the study area. The earliest folding is isoclinal with recognizable folds ranging from a few centimeters to three or more meters. These folds are often difficult to discern in outcrop because they have been transposed and/or, are contained in the plane of the dominant foliation. Such folding is responsible for conflicting readings in facing direction within the same outcrop where grading of units may be present. Intrafolial folds and rotated sheath folds (Cobbold and Quinquis, 1980) are also present. These are common in metapelitic units and in the granulites near contact zones (Figure 5). Later folding is on various scales ranging from 10 cm to 10m or more, and refolds early isoclines. Interlayered pelitic and granilitic units will demonstrate differing deformational responses: granulites will warp, while pelites may be kinked, or isoclinally folded intrafolially. Folds with monoclinic geometry and complex asymmetric folds also occur. These styles will be examined in the field. There are two zones where crenulation cleavage (Gray, 1970) occurs (Stop 1B) along the southernmost trace of the Gulf Hill Fault, and on the northeastern border of the study area.

Fold superposition is seen in a number of localities (Stops 1D, 5, 7, 9). Metamorphic mineral growth is also associated with directed fabric orientation; thus "primary" features are commonly obliterated or masked, and cataclastic/tectonically induced properties of the rocks mimic them. Both brittle and ductile features are present in folding, and "late" brittle fracturing may affect the foliation orientation locally, or show cataclastic (ribbon texture) in thin section.

Jointing is common in all rocks of the area, but may become intense locally in stream beds, or fracture/fault zones. A rhombohedral to tetrahedral fracture pattern is common in the rocks of this area. Joint orientations are displayed on Plate 3.

The location of faulting in the Saddleback Mountain area was determined from accumulated lines of evidence such as: degree of cataclasis, shearing of veins, or pegmatites; abrupt change of lithologies along strike, or on localized detatchment zones; topographic evidence, and map and aerial photo linears; ground topography such as swamp/stream alignment and silicified zones, (see Plates) and also, fold styles. There is also unreduced magnetic data which substantiates faulting in two areas, and a 2° back azimuth compass anomaly in one. Mafic dikes also are aligned spatially with fracture traces. The fracture/faults of the Saddleback Mountain area from a

PLATE 2A: LINEATIONS: RODDING, ROSE DIAGRAMS AND EQUAL AREA FOLD AXIS PROJECTIONS BY DOMAIN; PI OF FOLIATION SMALLER CIRCLES

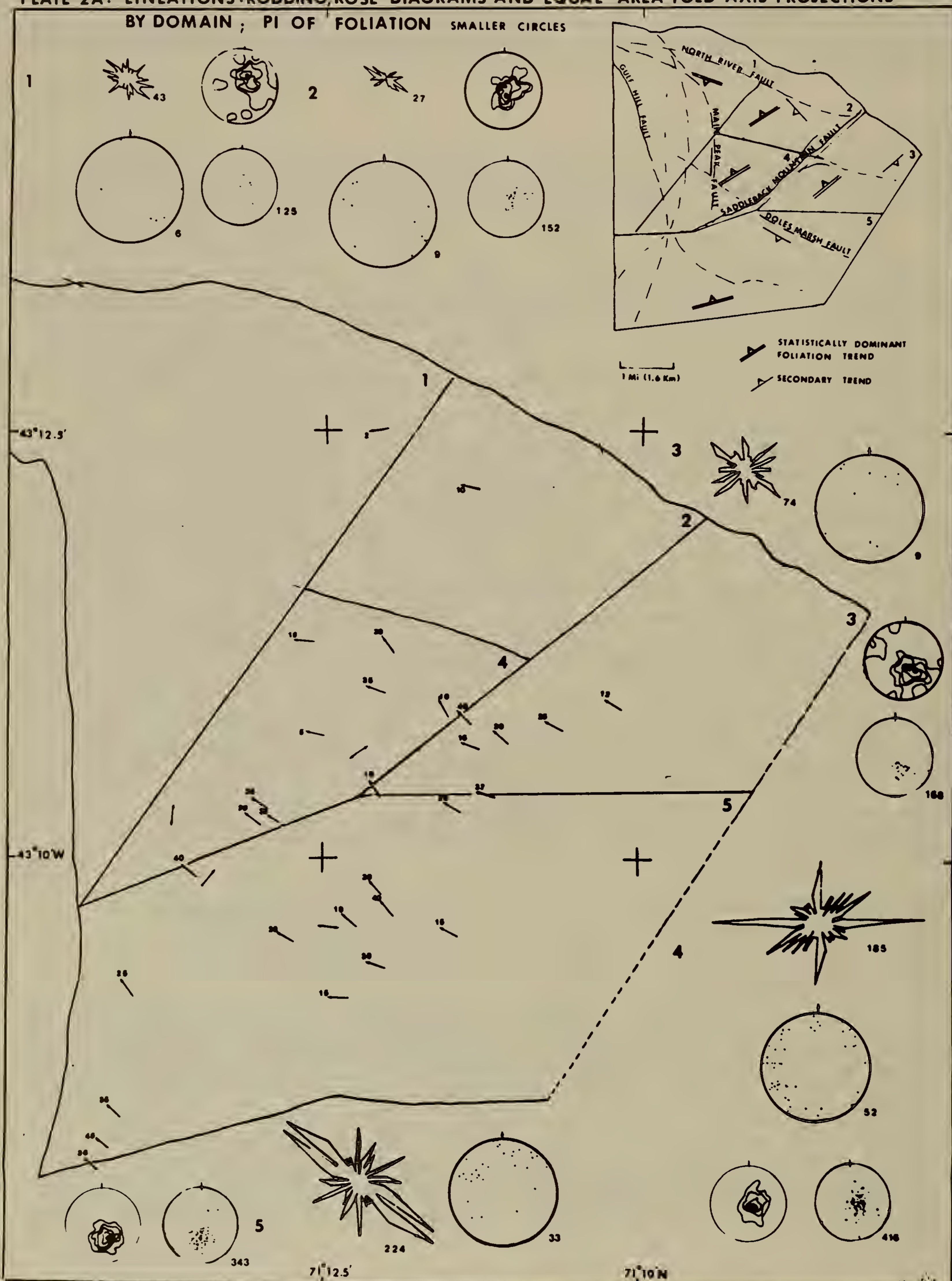


PLATE 2 : FOLD AXES AND LINEATIONS

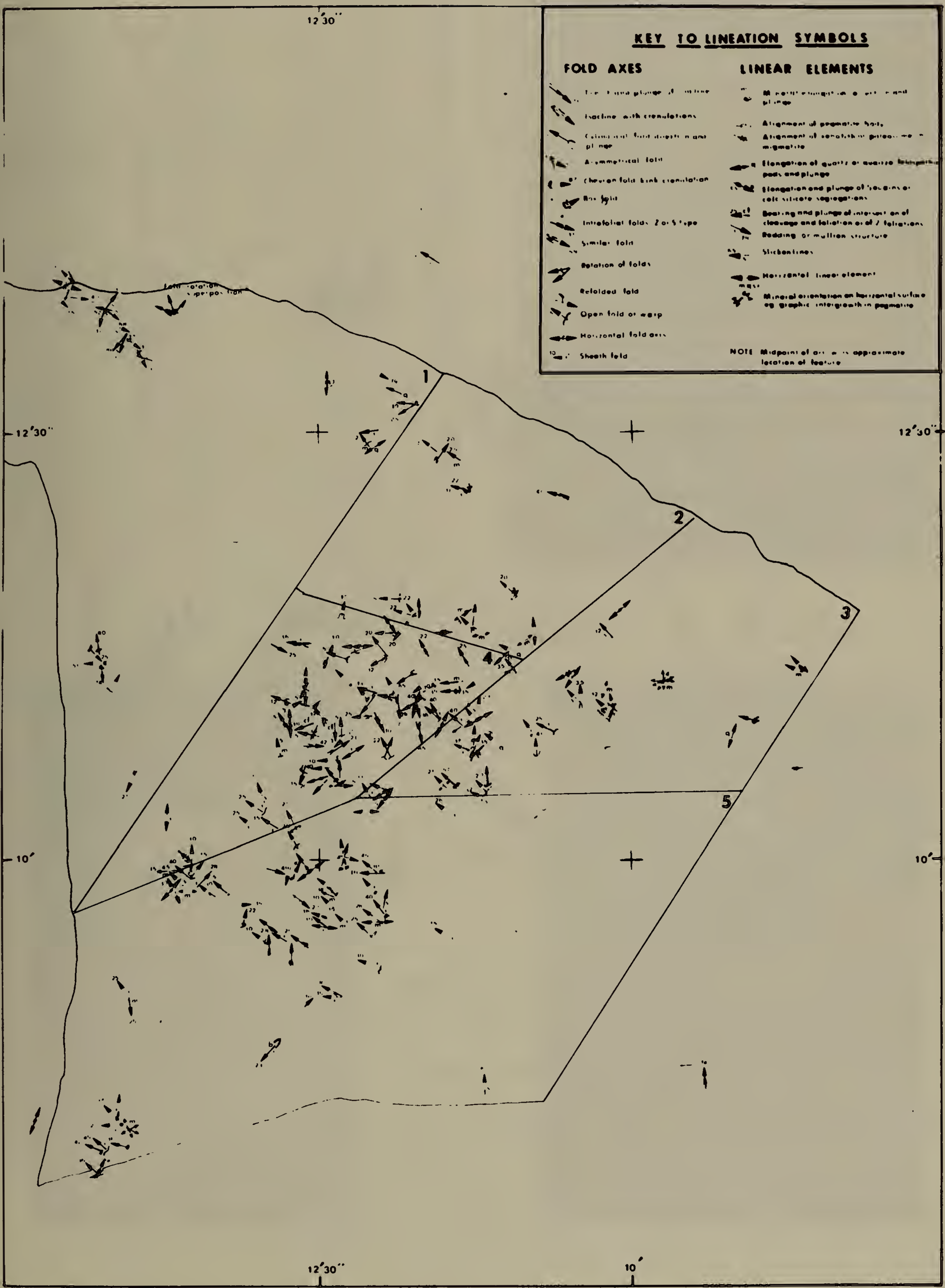


PLATE 3: JOINT PATTERNS, VEINS AND DIKES

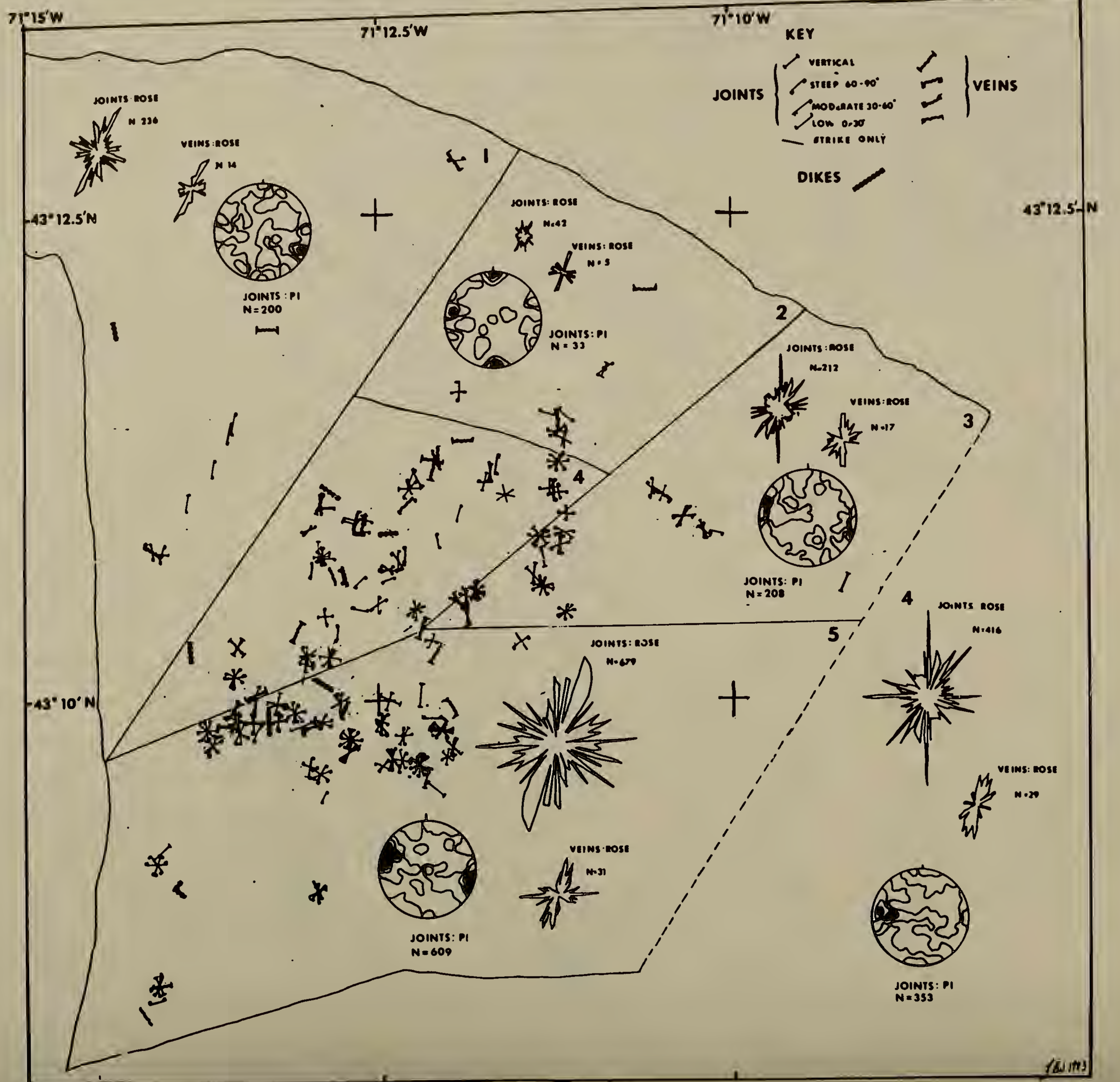


FIGURE 5A

FIGURE 5B

FIGURE 5: FOLD STYLES

A Outcrop of granulite showing tight folding in the third dimension, but when viewed along strike it appears eavenly bedded. Stop 9 NW Peak, 980' above Northwood town trail.

B Tightly folded granulite exposed along joint. Hammer shows contact with granodiorite. Stop 9, headwaters of Bear Brook.

C Small intrafolial sheath-like folds exposed near the contact of granulite and pelite. Stop 12A.

FIGURE 5C

pattern similar to perianticlinal faults (Sitter, 1956) or fractures over diapirs.

Any interpretation of the structure of the area must explain the criteria above.

The following structural models were considered:

1. Structural terrace model of Freedman (1950) i.e., Saddleback lies close to the nose of a plunging fold on the northeast limb of a major synclinorium.

2. Migmatite complex.

3. Nappe above a master decollement.

4. Basin and Dome pattern produced by fold superposition.

5. Graben or Horst.

6. Diapir i.e., the metasedimentary mass might have been domed by the injection of magma rising from below/by a rising basement block.

7. Braided fault zone as described by Kingma (1958).

The favored model is that of the braided fault zone, (Model 7) because it can incorporate the greatest number of features observed in the field: i.e., the alternation of igneous and metamorphic units with depth; differing degrees of cataclasis; the presence of localized detatchment zones; minor folding; fracture orientation such as is caused by a piercement body; the potential presence of large and small scale horst and graben blocks. This model also explains the presence of migmatites. The braided fault would serve as a tectonic pumping mechanism, or migmatite mixer. Differences in slope produced by vertical movement along a fault plane could also potentially generate the gravity mechanism needed to produce a major decollement. Transfer of motion, or torque between two splays of the major fault could explain fold rotation and superposition and the change in the dominant foliation patterns. It could also explain the presence of multiple metamorphic and deformational events by differential heat transfer and motion. One last far-out thought: heat transfer might also be effected by basalt intrusions (Pajari, et. al, 1981). Note the presence of basalts near or in granites - this type of relationship was also reported by Trygstad (1981).

The basalt dike at Stop 1 indicates tectonic mobility upon emplacement. A radiometric age date (i.e., 40-39) for this dike would be highly desirable, because it would allow bracketing of the lastest episode of mobility for the area. Were this date to be anomalously early, or late, then the sequencing of events for this portion of southeastern New Hampshire will need to be reconsidered.

The geophysical modeling of the northern termination of the Massabesic gneiss, which is mapped directly east of the area (Anderson, 1978) implies that there could be interleaving of igneous and metasedimentary units at depth. Such interleaving is seen in the Saddleback Mountain area. Radiometric dating of the Saddleback Mountain units and the units directly east might resolve some of the questions concerning the structure and age of the Massabesic. An open question remains concerning the stratigraphic correlation of this area: what if the composition of the lithologies is more a function of tectonism and partial melting than original lithologic composition? If this might be true, then the rock units would be a function of the induced P-T-X conditions of tectonism, then metamorphic "formations" should not be assigned stratigraphies. This problem needs to be resolved by detailed geochemistry.

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Road Log

Mileage	
0	Starting point is Johnson's Dairy Bar, North side of N.H. Route 4 in Northwood, N.H. To get there from Danvers, Mass, take I-95 North to the Portsmouth, N.H. Interchange. Exit at the intersection with N.H. Route 4 and travel westwards through Durham, N.H. Cross the Lee, N.H. traffic circle where Routes 4 and 125 intersect, and continue westwards on Route 4. Johnson's Dairy Bar is approximately 14 miles west of Lee traffic circle on N.H. Route 4. Starting time is 8:30 A.M. Please be sure you have your lunch and are prepared to bushwhack. Turn eastwards onto N.H. Route 4 from the parking lot of Johnson's Dairy Bar.
0.1	(Stop 1 will be interspersed between several localities: It encompasses the NW domain of the map of the area.)
(1 hour)	Stop 1. Intersection of N.H. Routes 4, 107 North and Blakes Hill Road. Cataclastic two-mica granite is exposed in the cut on the Blakes Hill Road side of the intersection (south side of Route 4). The cataclastic foliation in the granite follows domains which are related to anastomosing fracture surfaces. Foliation directions are affected by proximity to movement surfaces (Figure 1RL-A); joint and foliation surfaces are commonly listric. There are varying degrees of deformation within the outcrop. A highly fractured boudinaged slightly porphyritic basalt dike intrudes the granite (Figure 1RL-B).
0.1	Turn southwards onto Blakes Hill Road.
0.8	Fork in road: left fork is Kelsey Mill Road, right fork Blakes Hill Road. Stay to the right.
2.0	View of Saddleback Mountain looking eastwards across field.
2.1	Intersection of Mountain Road and Blakes Hill Road.
2.7	Granites exposed under power lines.
2.8	Turn around and park at the top of hill: walk down the hill to the gravel pit for:
	Stop 1A. This gravel pit lies at the south end of the Gulf Hill Fault as it was mapped by Freedman. Glacial gravels are being worked for road metal but the bedrock is also being sold as gravel. The bed rock is a highly friable crenulation-cleaved and kinked quartz-mica

FIGURE 2RL Stop 1D

FIGURE 4RL Stop 7

FIGURE 5RL Stop 10

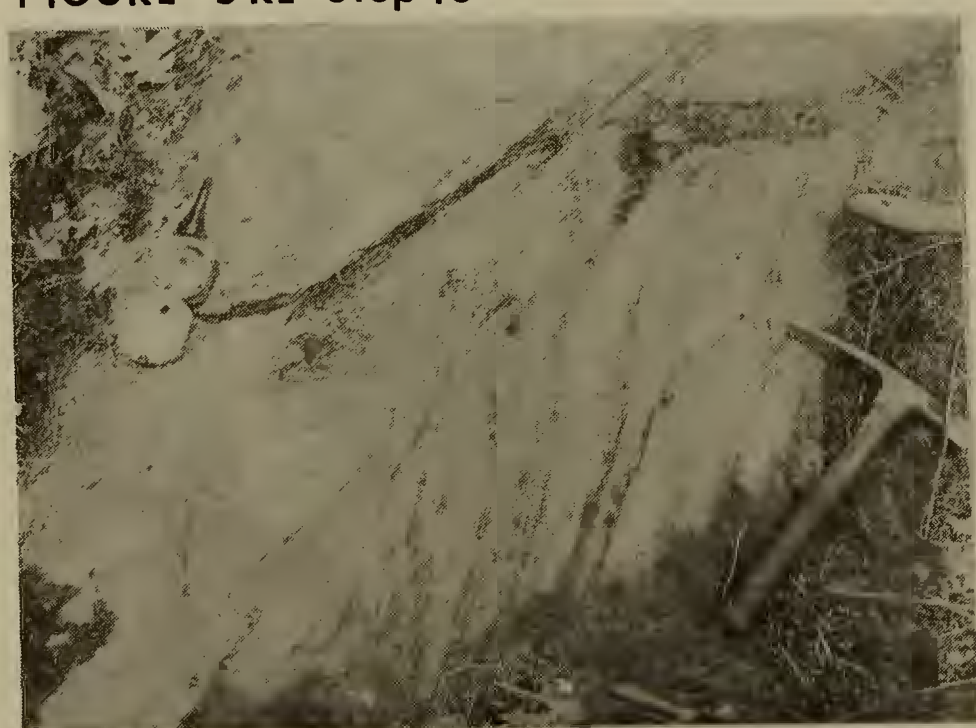

FIGURE 6RL Stop 11

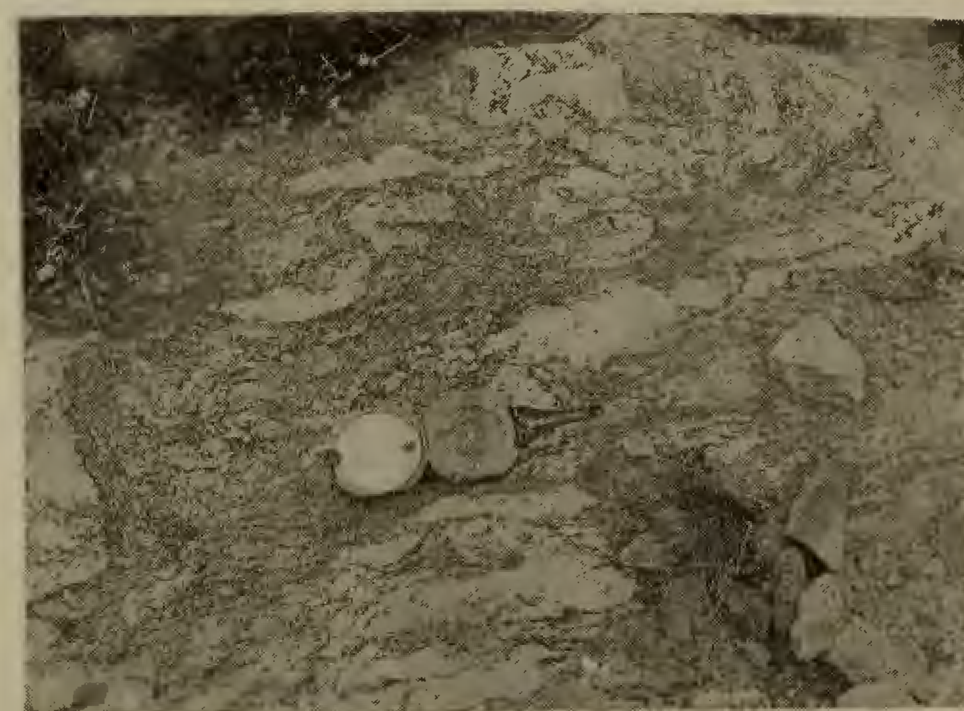

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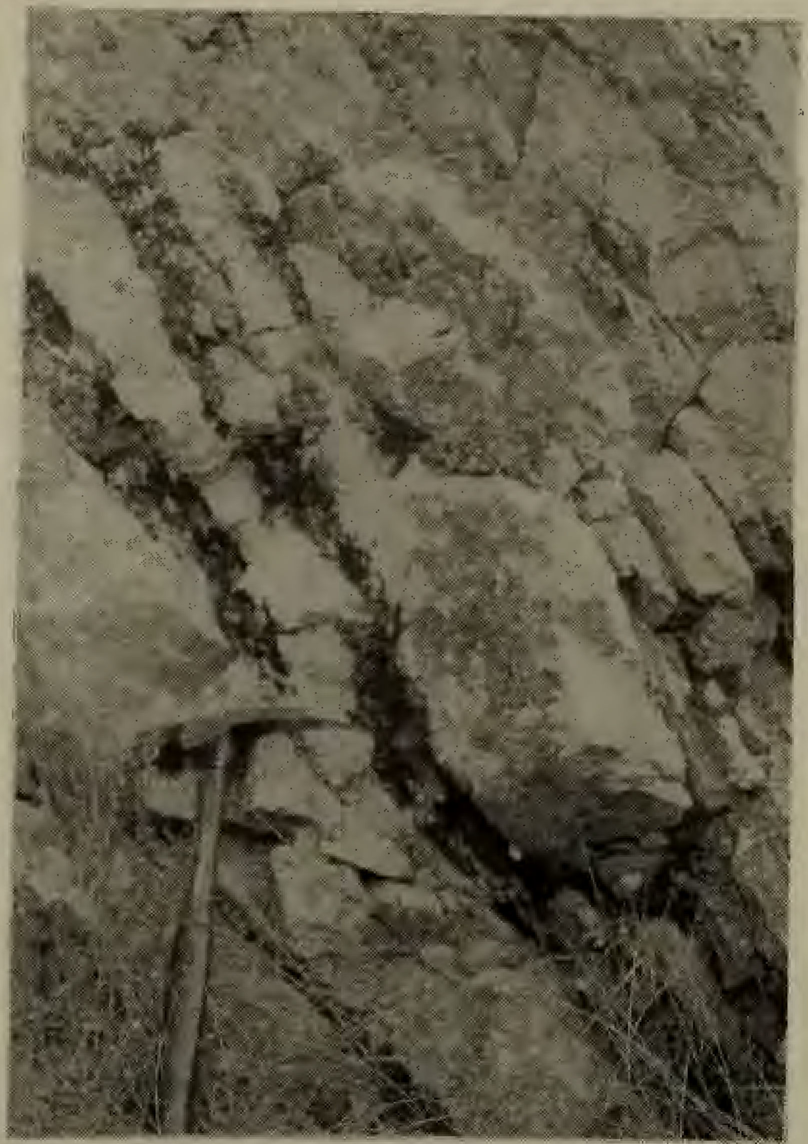
FIGURE 1 RL

A

B



A. LOCALIZED REORIENTATION OF CATACLASTIC FABRIC OF GRANITE NEXT TO "BRITTLE" FRACTURE

B. FRACTURING AND BOUDINAGE OF BASALT AND GRANITE

Mileage

phyllonite. The average foliation of the rocks is variable depending upon the depth of exposure, because the minor structures are being exposed.

Proceed up the hill, and travel northwards on Blakes Hill Road.

3.7 Pass junction of Mountain Road and Blakes Hill Road, Northwood.

3.9 Pass Junction of Harmony Road and Blakes Hill Road.

4.4 Cross area where power line traverse leads to the Gulf Hill Fault to the west.

5.0 Fork between Kelsey Mill and Blakes Hill Roads - take right fork onto Kelsey Mill Road heading eastwards.

5.2 Stop 1B. (Optional). Depending on conditions. North end of swamp where Kelsey Brook flows under Kelsey Mill Road. Note exposures of granite, gneissic schist (recrystallized mylonite?) and granulite.

Proceed northeast along Kelsey Mill Road.

5.7 (5 min.) Stop 1C. (Optional). Folds in schist in assocation with the contact with granite, exposures in the yard of the Northwood Oil Company, junction of Kelsey Mill Road and N.H. Route 4.

Turn eastwards onto N.H. Route 4.

6.6 View of west face of Saddleback Mountain looking southeastward from Harvey Lake on south side of the road.

7.5 Junction of Harmony Road and N.H. Route 4. Turn southwards onto Harmony road on south side of the road.

8.15 Turn around in turn around area of M.E. Johnson residence, head northwards on Harmony Road.

8.5 (30 min.) Stop 1D. (Optional). Dependent upon conditions. Exposure is along the shore line of Harvey Lake. If the water level of Harvey Lake is low enough, the contact zone between schist and garnetiferous granulite will be exposed, along with a folded outcrop of granulite showing development of micas on slip surfaces, fold superposition, healed cross fractures and minor thrust surfaces within the fold outcrop. If conditions do not permit viewing in, i.e., the outcrop is submerged, Figure 2RL will have to suffice.

8.8 (20 min.) Junction of Harmony Road and N.H. Route 4.

Mileage

Stop 2. Park cars on Harmony Road, go around the corner to outcrop on south side of NH Route 4.

This outcrop shows the contact between schist, granulite and granite. An example of a "juiced zone of movement?" There are tightly folded minor folds with NW-SE axial trend.

Turn right onto N.H. Route 4 and head eastwards.

12.7 View of Northeast side of Saddleback Mountain looking southwards from Northwood Ridge which we have been traveling along.

12.75 View of Mt. Pawtuckaway to the south.

12.9 Junction of NH Routes 4 and 43. Turn right onto Route 43 and go southwards.

14.0 Junction of N.H. Route 43, Lucas Pond Road on east and Mountain Road on west along east flank of Saddleback Mountain.

14.4 Junction of Old Deerfield Road and Mountain Road, Northwood, N.H. Turn left onto Old Deerfield Road and head south .15 miles. Elevation 574'.

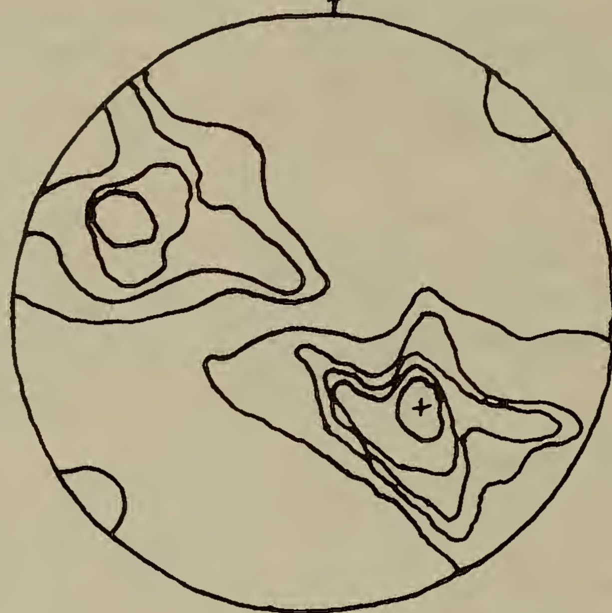
14.55 Stop 3. Outcrop of folded calc-silicate granulite exposed at elevation 580' approximately 60 yards west of Old Deerfield Road.

Thin sections of samples from this exposure reveal cataclastic properties (fracturing and millings of feldspars, pyroxene and amphibole phenocrysts). Quartz is sutured, also showing ribbon texture. Diopside is overgrown by amphiboles, and amphiboles are of two generations, and/or overgrown by biotite, indicating a complex structural and thermal history for these rocks. The calc-silicate pods may represent original compositional layering, but these are now transposed into rolled segregations. Folding is complex; the entire outcrop plots in a monoclinic, or kink mode. (Figure 3RL).

14.7 Exit Old Deerfield Road to NH Route 43, turn left, head north on Route 43.

14.95 Exposure of granulites in road cuts.

15.1 Junction of Woodman Road and Route 43. Turn right onto Woodman Road head southwest.



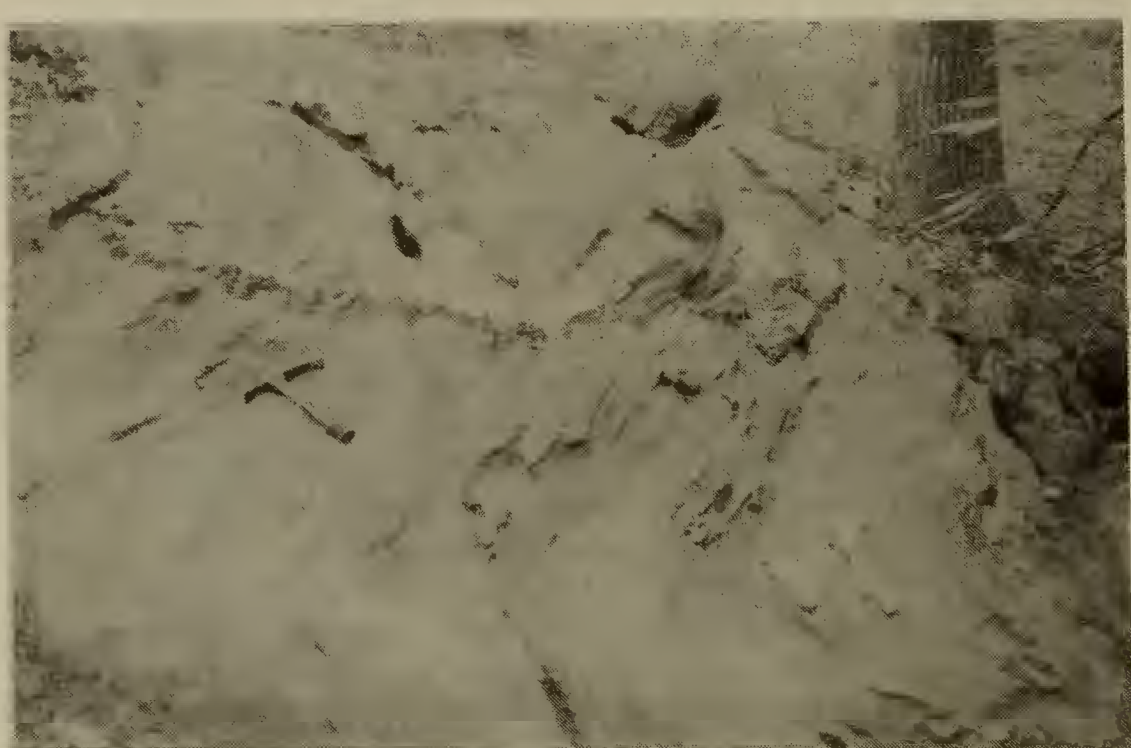


FIGURE 3RL OUTCROP OF FOLD .1 MI SOUTH OF THE INTERSECTION OF MOUNTAIN ROAD AND OLD DEERFIELD ROAD, NORTHWOOD, N.H., WITH PI DIAGRAM OF FOLIATION MEASUREMENTS (STOP 3)

Mileage	
15.2	Turn left onto service road opposite first trailer on right on Woodman Road. Park cars.
(30 min.)	Stop 4. Walk eastwards down trail to Camp Yavneh Lagoon.
	Exposures here are of silicified pegmatites and granular sillimanite-rich gneissic schist. There are two generations of sillimanite in these rocks: tabular and fibrolitic as well as two generations of biotite. This is an area of structural rotation of foliation. There is also a NW oriented rodding in this exposure.
15.3	Turn around, head back to the intersection of N.H. Route 43 and Woodman Road. Turn left onto Route 43, head southwards.
15.55	Stop 5. Road cut of granulites: note, directly north and west are exposures of pegmatites and biotite-rich granodiorite. Isoclines are present in these rocks even though not apparent at first. Rock staining enhances minor structures, which are delineated by the presence of potassic feldspar.
	Continue southwards on NH Route 43.
16.1	Outcrop of gneissic schist exposed on west side of Route 43.
16.2 (30 min.)	Stop 6. Masonry supply yard - exquisite exposure of folds in association with fractures and contact with biotite rich granite - granodiorite. Excavations for the foundation of the masonry supply warehouse revealed the changing structural orientation. This exposure in the yard of the masonry supply warehouse and the outcrop at NH route 43, as well as some in the Waterville Camp Ground indicate that there is a splay of the Saddleback Mountain Fault running directly through this area. Brittle and ductile features, as well as features indicative of subsolidus melting, and, or chemical mobility (lit-par - lit injection).
16.6	Proceed southwards on NH Route 43.
	Exposure of sillimanite-rich schist in outcrop on NH Route 43 near the entrance to Doles Marsh Wildlife refuge.
17.5	Deerfield, N.H. Junction of Coffeetown Road and NH Route 43.
19.0	Junction of Parade Road and N.H. Route 43, Deerfield, NH. (This is the turn off for Stop 10 later in the day).

Mileage	
	Route 43 changes direction to westerly. Continue on 43.
19.2	Exposure of granite pegmatite.
19.7 (30 min.)	<u>Stop 7A</u>. Exposure of granulite in road cut on north side of road. Contact with speckled biotite-rich granitic(?) rock similar to that seen at Stop 6. Move cars .2 miles down N.H. Route 43 to Stop 7B.
19.9	<u>Stop 7B</u>. Exposure of granulite, granite and basalt dike in road cut opposite the intersection of Mountain View Road, and N.H. intersection of Mountain View Road and N.H. 43, Deerfield, N.H., elevation 600' (Figure 4RL).
19.95	Junction of N.H. Route 43 and Mountain View Road.
20.1	Note silicification of pegmatites. Beryl-bearing pegmatites are exposed in the yard of the cape at the turn of Route 43. Note the fracture patterns associated with folding of the foliation planes of the granulites metasediments, and the development of coarse micas along shear surfaces.
	Continue southwards on N.H. Route 43.
21.2	Junction of N.H. Route 43, and N.H. Route 107 N-S and south Parade Road. Turn southwards onto 43-107 south. (There is a mafic dike buried on the north end of the lot of the Quonsett Hut at the junction of Parade Road and N.H. Route 43-107. This dike is highly fractured and is deflected indicating that it was moved upon emplacement. There is a high degree of sausiterization in the thin section of this dike and development of secondary calcite pyrrhotite. (Pyrrhotite forms fracture fillings.)
21.4	Cross Lamprey River on Route 43.
21.9	Junction of James City Road and Route 43 (N.H. Historical marker).
22.3 (30 min.)	<u>Stop 8</u>. (Also rest stop. Facilities graciously provided by the American Legion.) Deerfield American Legion Hall - Contact between gneissic schists and pegmatites. Pegmatites are zoned, there are rotating directions of feldspar aligment, and shears in the large feldspar phenocrysts-graphic granite was exposed in the excavation alignment of graphic intergrowth has spatial distribution akin to fracture orientation. Rimmed garnets in pegmatites. Walk .1 mile S on 43 exposure of schists which are folded. Pseudo-cross bedding is produced by

Mileage	
	fracturing in the rocks. Thin sections show evidence of retrogression in these rocks. This outcrop is in alignment with the trace of the Gulf Hill Fault.
	Compare these rocks to the outcrop of the Camp Yavneh Lagoon at Stop 4.
	Turn around, (note - last chance to buy lunch at one of the two country stores in the vicinity of the American Legion Hall.)
	Proceed northwards on N.H. Route 43-107.
23.2	Recross Lamprey River.
23.4	Junction of N.H. Route 43-107 (DEFINITELY LAST CHANCE TO STOCK UP AT STORE).
	Take N.H. Route 43 northwards.
24.4	Pass outcrop of Stop 7 on right.
24.5	Junction of N.H. Route 43 and Mountain View Road, Deerfield, N.H.
	Turn left onto Mountain View Road. Follow Mountain View Road up Saddleback Mountain.
25.8 (3 hours)	Elevation 900'. Turn around at road to tower. Park cars.
	Stop 9. (Lunch) - take lunches and field packs, walk up to WENH Microwave tower (Elevation 1100'). (Note, the elevation of the top of Saddleback Mountain reported on the 1981 7 1/2 minute Northwood Quadrangle is incorrect, total elevation is 1175'). Eat lunch and begin treck across the top of Saddleback Mountain.
	Views of Gulf Hill, Fort and Nottingham Mountains to the west, Blue Hills to northwest (on a clear day Mt. Monadnock can be seen to the southwest).
	The rocks exposed at the microwave tower are gneissic schists of variable composition often with abundant tourmaline. (See description in text). We will examine these rocks, then view the contacts between these rocks and igneous units, examine structures within biotite-rich granite to granodiorite, and structures within granulitic rocks on the northern side of the uppermost peak of the mountain and in the saddle of the mountain. The tranverse begins at approximately 030 azimuth from the south peak of

Mileage

the mountain. The contact between granite and gneiss occurs at the north end of the south peak in association with small enclaves of granulitic metasediments. Isoclinal folding/fracturing mimicking it is seen in the granites here. The traverse changes then to almost due east to the top of Saddleback Mountain's highest peak where tourmaline rich gneissic schists are exposed. A northwesterly traverse is then taken across the west flank of the highest peak. The contact between schist and granulite occurs in an approximately E-W trending direction. Granulites showing complex folding and fracture patterns are exposed on the west flank of the highest peak, (interlayered with granites on the north end), and on the saddle. The northwesterly traverse will take us through granulites, then granite, and back down through granulites, gneissic schist. At the top of the north west peak, granite pegmatite with quartz veining is exposed. The Town of Northwood Nature trail will be followed in a westerly direction, descending to the valley of the NW trending Lamprey River tributary known as Bear Brook and crossing it at approximately the 700 foot contour.

From there, a logging trail which trends approximately S25°E (155° Azimuth) will be followed. Along this traverse are migmatites with rotated granulite enclaves, and restites showing reaction rims. Exit the trail onto Mountain View Road at approximately 880-900' elevation by point where cars were parked.

Proceed southwards along Mountain View Road.

27.1 Junction of Mountain View Road, and NH Route 43, Deerfield, NH.

Turn right onto NH Route 43.

27.7 Junction of N.H. Route 43 and north end of Parade Road, Deerfield. Turn right and head southwards on Parade Road for .2 miles.

27.9 Stop 10. Junction of logging trail and Parade Road, Deerfield, NH.

(15 min.) Be careful parking cars; this is by a blind corner in the road.

This outcrop shows transposition structures characteristic of ductile shear at the borders between porphyritic granite and gneissic schist (Figure 5RL).

Turn around, head northwards toward NH Route 43.

Mileage

28.1 Junction of Parade Road and N.H. Route 43. Turn right onto N.H. Route 43, head northwards.

29.6 Junction of Coffeetown Road and N.H. Route 43.

(15 min.) <u>Stop 11</u>. Elevation 600'. Exposure of crenulated and rodded tourmaline-rich gneissic schist with granulite layers exposed in road cut at the corner of Coffeetown Road and Route 43. Also, refolding of ptygmatic folds in the yard of trailer .1 mi. north of road cut. (Figure 6RL).

Proceed northwards on N.H. Route 43.

The remaining stops are optional-depending upon time. If the trip ends here, return to Danvers. may be by heading southwards on NH 43-107S until it intersects NH Route 101 East. Follow 101E until it intersect I-95. Take I-95 south to Danvers.

30.5 Outcropping of sillimanite-rich schist at entrance to Doles Marsh Wildlife Refuge.

Turn around, head northwards on N.H. Route 43.

31.5 Turn left off of Route 43 onto spur of Old Deerfield Road .1 mile north of turn off to Doles Marsh Conservation area.

<u>Stop 12</u>. Folded schist in association with speckled biotite-rich granodiorite.

Proceed northwards on Old Deerfield Road.

31.8 Turn left, park and proceed up old trail.

<u>Stop 12A</u>. Follow trail westwards - trail will be following part of the trace of the Saddleback Mountain fault. Note contacts between "schists" and granulites, and intense fracturing in stream beds near traces of E-W fault. Folds resembling sheath folds described by Cobbold and Quinquis (1980). (See Figure 5 of text.)

Return to N.H. Route 43 north, follow it to N.H. Route 4, turn right, head eastwards towards Lee, Durham and Portsmouth. Take I-95 south to Danvers, Mass. from Portsmouth, N.H. interchange.

MOUNT PAWTUCKAWAY RING-DIKE COMPLEX

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University of Lowell, Lowell

Introduction

The Mount Pawtuckaway ring-dike complex, a member of the White Mountain magma series, is located in Rockingham County, New Hampshire. The north central portion of the recently published Mt. Pawtuckaway Quadrangle provides essentially complete coverage of the Mount Pawtuckaway complex. The complex occupies an area of approximately 3 square miles, is roughly circular in outline, and the maximum relief is on the order of 500 feet. The mafic rocks, which are easily eroded, underlie the lowlands while the more resistant monzonites form ridges. The only notable exception is a gabbro body which underlies Meloon Hill along the southern edge of the complex. The rocks of the Mount Pawtuckaway complex are intruded into foliated quartz monzonites of the Fitchburg Pluton.

The earliest studies on the Mount Pawtuckaway complex (Jackson, 1844; Hitchcock, 1878; and Smith, 1922) identified the rocks as being largely syenites and camptonites. Roy and Freedman (1944) were the first to completely map the complex and a further modification of the geology is found in Freedman (1950). Shearer (1976) carried out a largely geochemical study of the major units, but he does report modal data for some of the lithologies. Between 1981 and 1984 Eby and several University of Lowell students (J. Dadoly, M. Lambert, and J. Plunkett) partially remapped the complex and conducted a relatively extensive petrographic study. This has led to a preliminary revision (Fig. 1) of the geologic map found in Roy and Freedman (1944).

General Geology of the Mount Pawtuckaway Complex

The Mount Pawtuckaway complex is a plug-like structure which apparently extends to depth (Joyner, 1963; Bothner, 1976). Field relations indicate that the mafic rocks preceded the felsic rocks. The earliest mafic rocks are apparently pyroxenites which are now preserved as isolated blocks in the foliated diorite (see Fig. 1 for the locations of the various units). The foliated diorite and hornblende diorite are mineralogically similar, the difference being the presence or absence of foliation. Not all samples of the foliated diorite show foliation and, conversely, some hornblende diorite samples are foliated. These rocks do not show cross-cutting or intrusive relationships. A finer-grained phase of the diorite does seem to intrude some of the coarser-grained rocks. The porphyritic diorite and hornblende-biotite diorite are mineralogically similar, the difference being grain size, and are most likely variants of the same lithology. Mineralogically and chemically they can be viewed as late stage differentiates of the dioritic magma. The arcuate gabbro body found along the southern margin of the complex is distinguished by the anorthite content of its plagioclases (An_{60} to An_{46}), the essentially complete absence of apatite (which is a common accessory in the

Fig. 1. Generalized geologic map of the Mt. Pawtuckaway complex showing the major lithologic units. Dashed lines indicate possible outline of a second ring dike. Numbers indicate field trip stops. Geology modified from Roy and Freedman (1944).

diorites), and its distinctive trace element geochemistry. The position of the gabbro in the sequence of mafic rock emplacement is ambiguous since it is not intruded by any of the other units. The mineral and rock chemistry of the gabbro indicate that the magma from which it crystallized did not serve as a precursor to the dioritic magma.

The fine- and coarse-grained monzonites intrude the mafic units and locally carry mafic inclusions. The only exception to this generalization is the gabbro body in which no monzonite dikes have been found. The central coarse-grained monzonite body locally contains fine-grained monzonite inclusions indicating that it post dates the fine-grained monzonite. In places the central monzonite grades to a syenite. Scattered, isolated outcrops of both fine- and coarse-grained monzonite (and syenite) have been found both northwestward and southwestward of the central body. The northwestward extension is represented by a ridge. This extension of the central body is shown by the dashed lines on the geologic map, and apparently outlines a second ring-dike structure. Scattered outcrops of felsite on a small knoll along the western edge of the hornblende diorite unit indicate the presence of a small felsite intrusive. Similar rocks are found in a quarry in the porphyritic diorite. Lamprophyre and aplite dikes apparently represent the last phase of igneous activity in the Mount Pawtuckaway complex.

Ages determined by a number of methods are currently available for several of the units. The coarse-grained monzonite has been dated as 124 Ma (K-Ar biotite, Foland et al., 1971), 126 Ma (Rb-Sr whole-rock isochron, Eby, unpublished), and 134 Ma (K-Ar biotite, Krueger, personal communication). A Rb-Sr whole-rock isochron gives an age of 129 Ma for the central fine-grained monzonite plug (Eby, unpublished). An apatite fission-track date gives an age of 130 Ma for the foliated diorite (Eby, unpublished), and ages of 129 Ma (K-Ar amphibole) and 123 Ma (K-Ar biotite) have been determined for a sample from the hornblende diorite (Krueger, personal communication). The age data indicate that the igneous activity was confined to a period of 10 m.y., and most likely an even shorter time interval. The one exception is an apatite fission-track age of 107 Ma obtained for a lamprophyre dike located at the top of South Mountain (Eby, unpublished). Similar ages to that of the lamprophyre dike have been found elsewhere in the White Mountain igneous province, including the geographically near Little Rattlesnake Hill (Foland and Faul, 1977), so a minor period of igneous activity of this age in not unlikely for the Mount Pawtuckaway complex.

The available data suggest that the Mount Pawtuckaway complex represents a cross-section through a vertical conduit up which moved a series of magmas closely spaced in time. At least three separate mafic magmas are required to account for the pyroxenites, the diorites, and the gabbros. On the basis of preservation it might be suggested that the magma forming the diorites represents the last of these pulses. In each of the pulses crystallization apparently proceeded from the walls inward, and a number of the mafic rocks are largely composed of cumulus minerals. The conduit eventually became plugged and differentiation of a mafic magma in a chamber at depth led to a monzonitic liquid. The partial evacuation of this chamber and upward passage of the liquid, most likely to the surface, is represented by the fine-grained monzonites. The overlying column of mafic rock then collapsed into the chamber and the extrusion of additional liquid led to the coarse-grained monzonite units. Chemically the fine-grained and coarse-grained monzonites are quite similar suggesting that they were derived from the same magma. A later minor period of activity is represented by the felsites, lamprophyres, and aplites.

Petrography of the Various Lithologies

Pyroxenite

The pyroxenites are coarse-grained and largely composed of cumulus olivine and augite with interstitial labradorite and opaque minerals. The augites show a pink tint and are spotted and rimmed by red-brown amphibole. The augites contain minute opaque inclusions which are oriented parallel to crystallographic directions. A green mineral (hercynite?) is associated with the opaque minerals. Apatite occurs in trace amounts.

Gabbro

The gabbros are medium- to coarse-grained and locally show a well-developed foliation due to the alignment of plagioclase laths and seggregation of mafic from felsic minerals. Plagioclase (An_{60} to An_{46}) and a light pink augite are the major minerals, but olivine is locally abundant. The augites contain oriented minute opaque inclusions. The augites are rimmed and spotted by red-brown amphibole. Apatite occurs in trace amounts.

Hornblende diorite

The grain size is variable from medium-fine-grained to coarse-grained and locally foliation can be found. The plagioclases are generally andesine, but can be zoned to oligoclase. The pyroxenes are generally light green but pink cores are not uncommon. The pyroxenes are often extensively replaced by reddish-brown (hornblende) and green (hastingsite?) amphibole. Red-brown biotite occurs both as separate grains and replacing pyroxene and amphibole. Apatite is a common accessory ranging in modal abundance from 1.6 to 3.5%. Olivine, extensively altered, is occasionally found.

Foliated diorite

The grain size is variable from fine- to coarse-grained and most specimens show a foliation due to the alignment of plagioclase grains and the separation of mafic from felsic minerals. The plagioclase is generally andesine, but it may be zoned to oligoclase and occasionally labradorite cores are found. The pyroxenes are light pink and light green, and where the two varieties occur together the light pink pyroxene constitutes the core. The pyroxenes are spotted and replaced by red-brown and green amphibole, and locally this replacement is almost 100%. The pyroxenes contain oriented minute opaque inclusions, and the preservation of these inclusions in the amphiboles indicates the prior existence of pyroxene. Olivine occurs in most specimens and locally is an important accessory. The biotites are straw brown to red brown and generally occur as large flakes. Apatite is an important accessory ranging in modal abundance from 1.0 to 2.6%

Porphyritic diorite and hornblende-biotite diorite

These two rocks are mineralogically similar, the difference being the much coarser grain size of the porphyritic diorite. The plagioclase is generally oligoclase and K-feldspar occurs in minor amounts. Particularly in the porphyritic diorite, the oligoclase is often rimmed by K-feldspar. A green pyroxene is occasionally found, usually largely replaced by red-brown amphibole. Large flakes of reddish-brown to straw-brown biotite usually exceed amphibole in modal abundance. Apatite and opaque minerals are the accessories.

Table 1
Representative Modes for the Mafic Rocks

	Pyroxenite	Gabbro		Hornblende diorite		Foliated diorite		Porph. diorite	Hb-Bio diorite
	MP73	MP81	MP83	MP1	MP50	MP9	MP75	*	*
Plagioclase	8.3	61.0	66.7	45.6	63.5	45.3	62.9	47	55
K-feldspar	—	—	—	—	—	—	—	10	5
Olivine	18.2	4.8	20.5	—	—	7.0	0.9	—	—
Pyroxene	50.6	10.0	0.4	16.4	1.0	11.0	6.1	—	5
Amphibole	18.2	19.1	5.9	20.4	25.9	23.6	20.0	18	12
Biotite	—	—	—	8.8	1.4	2.4	3.9	17	18
Opaque	4.5	5.1	6.3	6.1	4.7	8.1	4.4	3	5
Apatite	0.2	—	0.2	2.7	3.5	2.6	1.8		

*Modes from Roy and Freedman (1944)

Coarse-grained monzonites and syenites

The grain size varies from medium- to coarse-grained and the monzonites and syenites are gradational into each other with changes in the K-feldspar/ plagioclase ratio. The syenites tend to occur towards the center of the large monzonite bodies suggesting that differentiation proceeded from monzonites to syenites. The plagioclase is generally oligoclase and the alkali feldspars are microperthitic. The plagioclases are often rimmed by perthite. The pyroxenes are colorless to light green and are partly replaced by red-brown and dark green amphiboles. The biotites are reddish brown to straw brown and replace both pyroxene and amphibole. Quartz is interstitial, and some sections contain fayalitic olivine.

Fine-grained monzonites

The grain size varies from very-fine-grained to fine-grained and some sections have phenocrysts of biotite and hornblende. The major minerals are oligoclase and microperthite. Quartz is interstitial. The amphiboles are green to dark green and the biotites are reddish brown to straw brown. Opaque minerals and apatite occur as accessories.

Felsite

Small phenocrysts of alkali feldspar, plagioclase, pyroxene, and amphibole are found in a fine-grained groundmass of alkali feldspar, plagioclase, pyroxene, amphibole, and opaque minerals. The phenocrystic plagioclase is high oligoclase while the groundmass plagioclase is low oligoclase. The pyroxenes are green and slightly pleochroic. The phenocrystic amphibole is red-brown while the groundmass amphibole is green to dark green. The plagioclase phenocrysts are surrounded by reaction rims and the pyroxene phenocrysts are rimmed by dark green amphibole.

Table 2
Representative Modes for the Felsic Rocks

	Coarse-grained monzonites & syenites				Fine-grained monzonite
	MP8	MP12	MP15	MP49	*
Plagioclase	19.4	31.3	23.8	8.4	40
K-feldspar	46.2	33.6	53.6	83.9	35
Quartz	—	0.5	2.2	2.5	2
Olivine	0.7	0.9	—	—	—
Pyroxene	12.0	13.9	4.0	0.2	4
Amphibole	0.4	6.2	11.3	3.2	7
Biotite	14.6	7.5	3.9	—	9
Opaque	5.4	5.3	1.2	1.8	3
Apatite	1.3	0.8	—	—	

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Road Log

Mileage

0 Entrance road to Pawtuckaway Mountains and fire tower. The entrance is marked by a small brown sign on the east side of NH Route 107, 3.2 miles north of the juncture of Routes 107 and 101.

1.5 Stop 1. Small pull-off on north side of road. Walk 200 feet north into small clearing for scenic vista of the Pawtuckaway Mountains. Return to the road and walk west past the park service sheds. Foliated outcrops of quartz monzonite of the Fitchburg Pluton are found along the road. Proceeding back toward pull-off, outcrops of fine-grained gabbro on hill just east of park service sheds, which apparently represent a chill zone. The rock coarsens rapidly over a distance of 100 feet. Scattered outcrops in woods on south side of road occasionally show foliation and several lamprophyre dikes are exposed.

2.0 Stop 2. Park by small farm cemetary on south side of road. Follow logging road departing from west side of cemetary 300 feet to large flat outcrops of gabbro. Take left hand fork and continue another 1000 feet and then proceed southwesterly to SW side of Meloon Hill. Outcrop is essentially continuous along this side of the hill. The gabbro is medium-grained and locally shows a well-developed foliation. The foliation strikes parallel to the contact and dips steeply towards the center of the intrusion. This unit is cut by several fine-grained mafic dikes. The gabbro is differentiated from the other mafic units on the basis of the anorthite content of the plagioclases (An_{60} to An_{46}), the essentially complete absence of apatite, and its distinctive trace element geochemistry.

2.5 Stop 3. Outcrops of medium- to coarse-grained hornblende diorite to the west of the road. Some of the outcrops show a poorly-developed foliation. Also found in the immediate area are outcrops of fine-grained monzonite and diorite. One of the hornblende diorite outcrops is cut by an intermediate dike showing a trachytic texture. Syenite and monzonite dikes are also found cutting the hornblende diorite.

2.8 Stop 4. Walk east 200 feet along logging road. Small quarry located just to the north of the road. This is the best exposure of the porphyritic diorite. At the western end of the quarry coarse-grained monzonite has been engulfed by fine-grained bluish-gray material which may be classified as monzodiorite. Isolated outcrops of this monzodiorite are found in the immediate area. This rock apparently represents one of the last igneous events in the Mount Pawtuckaway complex. Proceeding eastward in the quarry, outcrops of porphyritic diorite are observed. These outcrops are cut by both felsite and fine-grained monzonite dikes. The felsite appears to be identical to the rock which underlies a small knoll along the western edge of the hornblende diorite body (shown as a small felsite plug on the geologic map).

Stop 5. Proceed southeastward from the quarry up Middle Mountain. A series of outcrops provide almost 100% exposure of the fine-grained monzonite. CAUTION: This rock is very brittle and fragments come off the outcrop like shrapnel. Do not wound yourself or a fellow geologist. There are slight variations in grain size throughout this unit, but they do not appear to be correlated with distance from the contact. On fresh surfaces the rock is greenish-black.

Stop 6. At the top of Middle Mountain outcrops of coarse-grained monzonite are found. Proceed a short distance eastward through this unit. In this area the outcrops are deeply weathered and fresh pieces are difficult to obtain. Inclusions of fine-grained monzonite are found in some outcrops of coarse-grained monzonite.

3.4 Stop 7. Park at the intersection of the loop road and Round Pond road. During times of heavy rainfall, the road may not be passable between Stop 4 and Stop 7. Walk back (west) along the road several 100 feet to a road leading north into a primitive picnic area. Outcrops of coarse-grained monzonite are found on either side of the road. Diorite and fine-grained monzonite inclusions are found in the coarse-grained monzonite. On the east side of the road are found several outcrops of fine-grained monzonite. These outcrops are texturally interesting because blebs of coarse-grained monzonite are found in the fine-grained monzonite. In thin section no sharp boundaries are observed, but simply a distinct change in grain size. If time permits, continue westerly along the ridge. Scattered outcrops of coarse- and fine-grained monzonite are found along with outcrops of foliated diorite. This ridge may represent a continuation of the monzonite bodies which would suggest the presence of a second ring dike.

Stop 8. Return to intersection and continue on the Round Pond road in an easterly direction. Outcrops of pyroxenite are found along the road approximately 400 feet from the intersection. Outcrops of pyroxenite are found throughout this area and in places they have been intruded and engulfed by fine- to medium-grained diorite. The pyroxenites, therefore, must represent an early stage of the magmatic history.

Stop 9. Continue eastward onto an abandoned road. Outcrops of foliated diorite are found in and on both sides of the road. Both fine- and medium-grained varieties of the foliated diorite are observed. Where the two varieties are found in contact, the fine-grained diorite appears to intrude the medium-grained diorite. The foliation parallels the contact with the coarse-grained monzonite and dips steeply inward.

Stop 10. Continue eastward to Round Pond. Outcrops of coarse-grained monzonite north of road and just west of brook carry inclusions of fine-grained diorite. A mafic dike cutting the monzonite is exposed in the stream bed.

4.2 Stop 11. Continue southward on loop road to parking area for fire tower trail. Proceed up trail to top of South Mountain. Excellent exposures of the coarse-grained monzonite are found along the upper portion of the trail. Fresh samples have a definite greenish cast. A number of lamprophyre dikes are exposed in the immediate area of the fire tower. An apatite fission-track date for one of these dikes gives an age of 107 Ma. At the southern end of the fire tower outcrop, a lamprophyre dike has been disrupted and engulfed by aplite.

Stop 12. Optional. Climb Middle Mountain directly across from the fire tower. The fine-grained monzonite tends to be slightly coarser on this side of Middle Mountain and mafic inclusions are relatively common. Continue on into the coarse-grained monzonite unit which locally grades into syenite.

Continue around the loop road to the starting point. At the time this field guide was written, the telephone company was putting in a line and blasting for the poles provided fresh samples of hornblende diorite and foliated diorite. These samples can be collected along the east-west segment of the loop road.

CAMBRIAN ROCKS OF EAST POINT, NAHANT, MASSACHUSETTS

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Introduction

The quintessential stratigraphic component of Avalonian terranes of eastern North America is a Cambrian succession bearing the so-called Acado-Baltic trilobite assemblage. Spectacular sea cliffs at East Point, the easternmost extremity of Nahant, Massachusetts, afford an opportunity to examine a continuous and well exposed Lower Cambrian section on the Boston Platform. The Nahant Gabbro, sills, abundant dikes, and faults cut the Cambrian strata and add to the geological excitement. Indeed, it is difficult to move more than a few meters along the cliffs without discovering a feature that will arouse your curiosity. This is also a wonderful place to watch waves crash against cliffs and to stare across the Atlantic in the direction of Africa.

Regional Lower Cambrian Stratigraphy

Cambrian rocks of Nahant are correlated with distinctive green and red nodular slates of the Weymouth Formation (LaForge, 1932) exposed in the Mill Cove area of Weymouth (Figure 1). Trilobites have never been found in Nahant Strata (a slate pebble with *Strenuella* was found; Grabau, 1900), but a thin zone in the Weymouth Formation at the type locality (Burr, 1900; Grabau, 1900) yielded a diagnostic trilobite assemblage (*Strenuella*, *Callavia*, and *Weymouthia*) indicative of the late Early Cambrian of the Acado-Baltic faunal province and similar horizons in England and Morocco (Shaw, 1950; Theokritoff, 1968; Anstey, 1979). The Hoppin Formation at Hoppin Hill near North Attleboro, (Figure 1) also contains *Strenuella strenua* (Billings). A number of workers have noted (Foerste, *in* Shaler and others, 1899; Grabau 1900; Theokritoff, 1968; Landing and Brett, 1982) that in both the Hoppin and Weymouth Formations the trilobite bearing assemblages overlie or are separated from a fauna dominated by small conoidal fossils. Landing and Brett (1982) assigned the lower part of the Hoppin Formation, and the limestones of the Weymouth Formation at Nahant which lack trilobites, to the earliest Cambrian Tommotian Stage (Figure 1) and the microfauna associated with the *Strenuella* (trilobite) bearing beds at Hoppin Hill to the overlying Atdabanian Stage. The restriction of faunas to particular lithofacies, the very limited exposure of good stratigraphic sections, the intimate association of the trilobite and non-trilobite assemblages, and the occurrence of a trilobite(?) in limestone at Nahant suggest to me that the

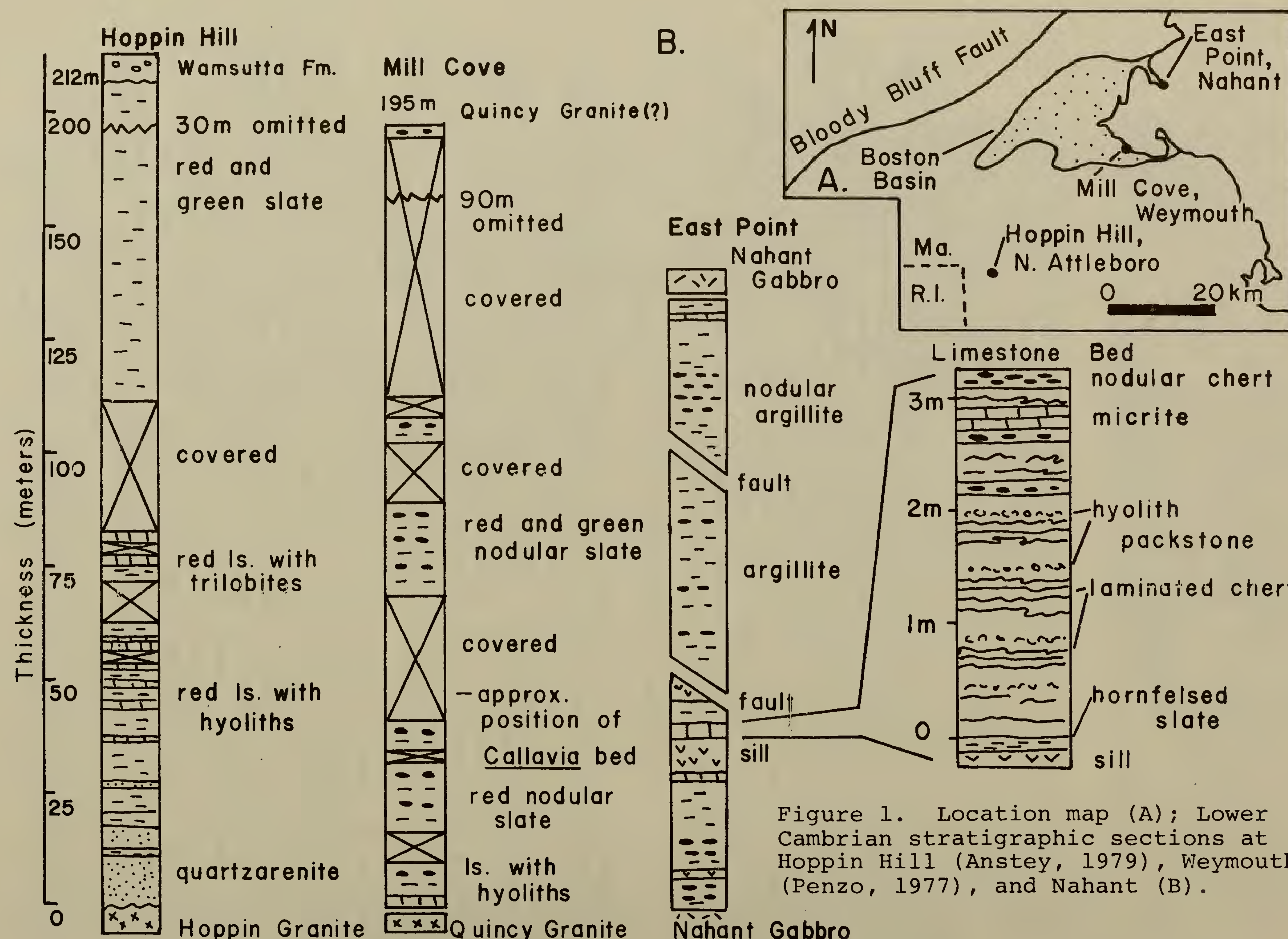

Figure 1. Location map (A); Lower Cambrian stratigraphic sections at Hoppin Hill (Anstey, 1979), Weymouth (Penzo, 1977), and Nahant (B).

Lower Cambrian rocks of eastern Massachusetts are probably not Tommotian but lower Atdabanian or slightly younger. Theokritoff (1968) explained the lack of co-occurrence of the trilobite and non-trilobite faunas as a probable result of facies control of assemblages and he therefore suggested that a non-trilobite Coleoloides fauna need not necessarily be earliest Cambrian. Landing, Nowland, and Fletcher (1980) also report the extension of several typical Tommotian phosphatic microfossil ranges well into the Callavia zone in Nova Scotia. We clearly need a better understanding of the taxonomy, stratigraphic ranges, and paleocecology of many of the non-trilobite index fossils of the earliest Cambrian. Until such knowledge is gained I suggest that application of the stage names of the Siberian Platform (Matthews and Missarzhevsky, 1975; Raaben, 1981) to the Cambrian successions of isolated areas such as eastern Massachusetts may be somewhat premature.

Strata equivalent in tectonic setting, depositional environment, and fossil content to the Hoppin and Weymouth Formations are excellently exposed in eastern Newfoundland and in more limited exposures in Nova Scotia and New Brunswick (McCartney, 1969). In Newfoundland and New Brunswick there is a profound, often angular, unconformity, with an overlying basal quartzarenite, as at Hoppin Hill, followed by the Lower Cambrian succession (Skehan, 1969). Kaye and Zartman (1980) have recently proposed that the Cambridge Formation of the late Precambrian Boston Basin grades upward into the Weymouth Formation to form a continuous and conformable succession. The lithologies of the Weymouth and the Cambridge are quite distinct and different, especially when viewed on the outcrop scale. The well documented nonconformity at the base of the Cambrian at Hoppin Hill (Billings, 1929; Dowse, 1950), less than 40 km to the southwest of the Boston Basin, is the result of transgression of shallow water facies onto a stable continental(?) block. The shallow water facies of the Weymouth and Hoppin Formations, the uncomformable association with stable basement, and the regional correlations do not support a basinal setting for the Weymouth Formation. Billings (1982) argues, on structural grounds, that an uncomplicated transition between the Boston Bay Group and the Cambrian strata in Weymouth and Quincy is unlikely.

Nahant Stratigraphy

About 130 m of strata are exposed in northerly dipping beds of East Point. It is difficult to present a simple stratigraphic section as there are numerous fault offsets and 2 thick sills that interrupt the sequence. The section consists of a dark silicified mudstone or argillite with interbedded nodular horizons and limestone beds. A 3m thick limestone bed is found on the southeast side of the point along the top of the cliff. The three basic Cambrian lithologies are discussed below.

Argillite

Most of the section is a brittle, dark gray to black, thinly laminated argillite. Bedding is faintly visible in hand specimen. Quartz silt and very fine quartz grains are present as thin, parallel laminae, and lenses (0.1-0.5 mm thick) or as scattered isolated grains. These laminae are separated by a dark structureless mudstone. Some undulating contacts resemble scour surfaces. Very rare burrow-like structures are present.

Nodular Argillite

The most characteristic lithology of the Weymouth is a red, green, or black slate or argillite containing elongate carbonate nodules. The nodules are from 0.5 to 3 cm in thickness and most are about 3 to 15 cm in length. They occur with varying frequency along bedding planes and occasionally are so abundant as to comprise a thin "limestone" bed. The nodules have been altered primarily by silicification, and by replacement with Ca-garnet, tremolite, wollastonite, and epidote. The highly altered nodules are often zoned or banded with chert and carbonate interiors and garnet and Ca-silicate rims (Bingham, 1977). Chertification initiated in the nodule often extended into the surrounding mudstone. I have seen no fossils in either the replaced or the relatively unaltered nodules. Underlying and overlying laminations in the mudstone are often deflected and appear to be displaced by nodules suggesting that the mudstone was unconsolidated when the nodules formed. The shapes of nodules are reminiscent of algal structures, but no definitive internal structure is present.

Limestone

White to light gray limestone beds range in thickness from 3 cm to 3 m. The very thin limestones usually occur in groups with interbedded, abundantly nodular, slate or argillite. Thicker limestones contain 0.5-3 cm thick, very thinly laminated, highly irregular, brownish or greenish chert beds. The cherts themselves do not contain fossils; however, patches of fossiliferous chertified limestones are associated with the greenish layers. Fossils occur in clearly defined thin zones within the limestones. These fossiliferous layers may be continous and traceable for several meters along the surface of the outcrops or they may be isolated as irregular masses.

Petrography of Limestones

The primary textures of limestones at East Point are remarkably well preserved given the proximity of igneous rocks. The rocks were originally biomicrites or fossil wackestones. The micrite has generally recrystallized to very finely crystalline sparry calcite although some patches of microspar are present (Figure 2). Abundant bioclast, primarily hyolithids, are composed

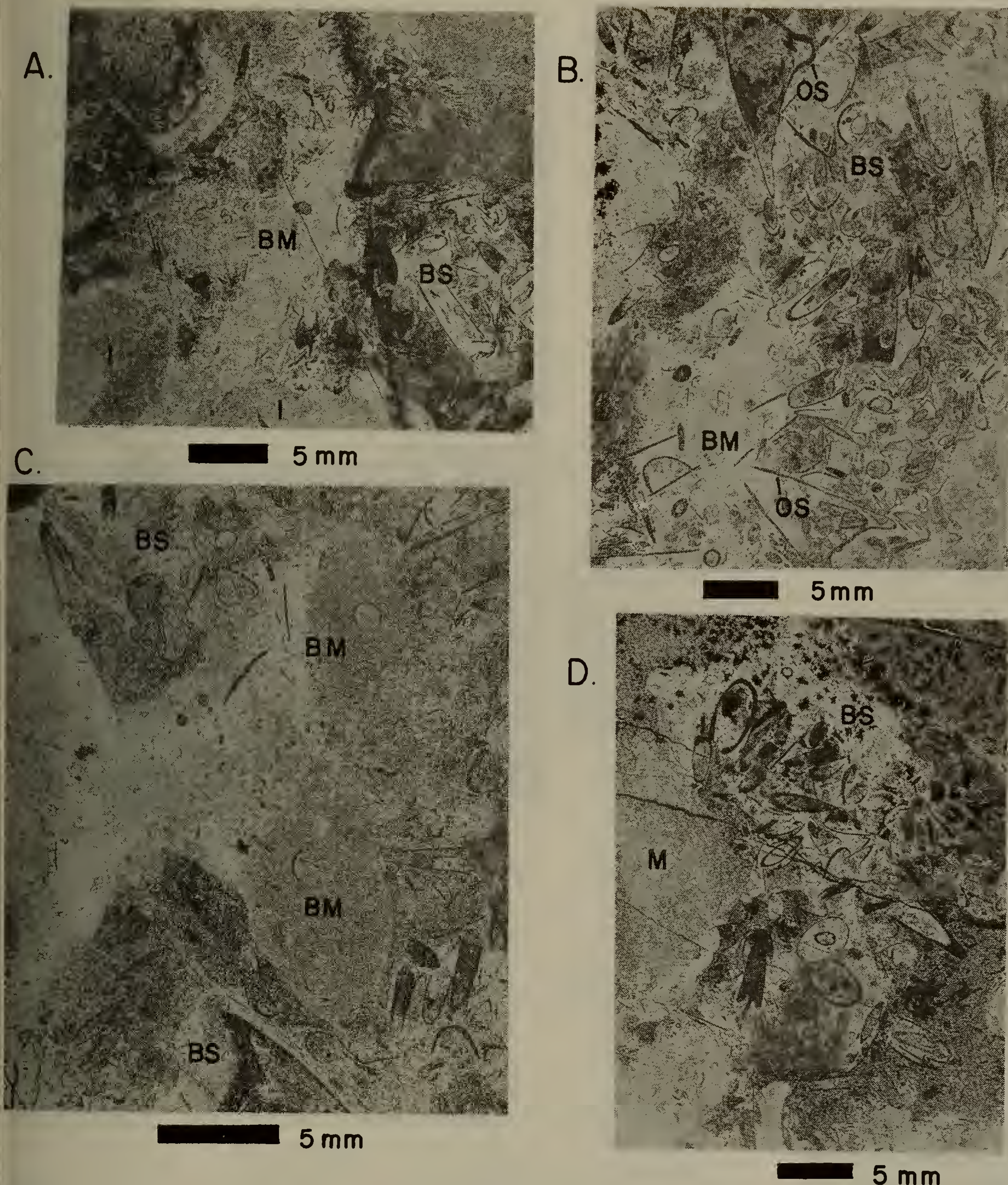

Figure 2
Negative prints of acetate peels from slabs cut parallel to bedding. Microfacies labeled micrite (M), biomicrite (BM), biosparite or packstone (BS), scale bar is 2 cm. A. intraclasts (I) surrounded by biomicrite and biosparite.; B. hyolith packstone and biosparite, note oblique sections of Orthotheca searsi (OS); C. irregular biomicrite mass surrounded by hyolith packstone; D. pocket of hyolith packstone, surrounded by micrite.

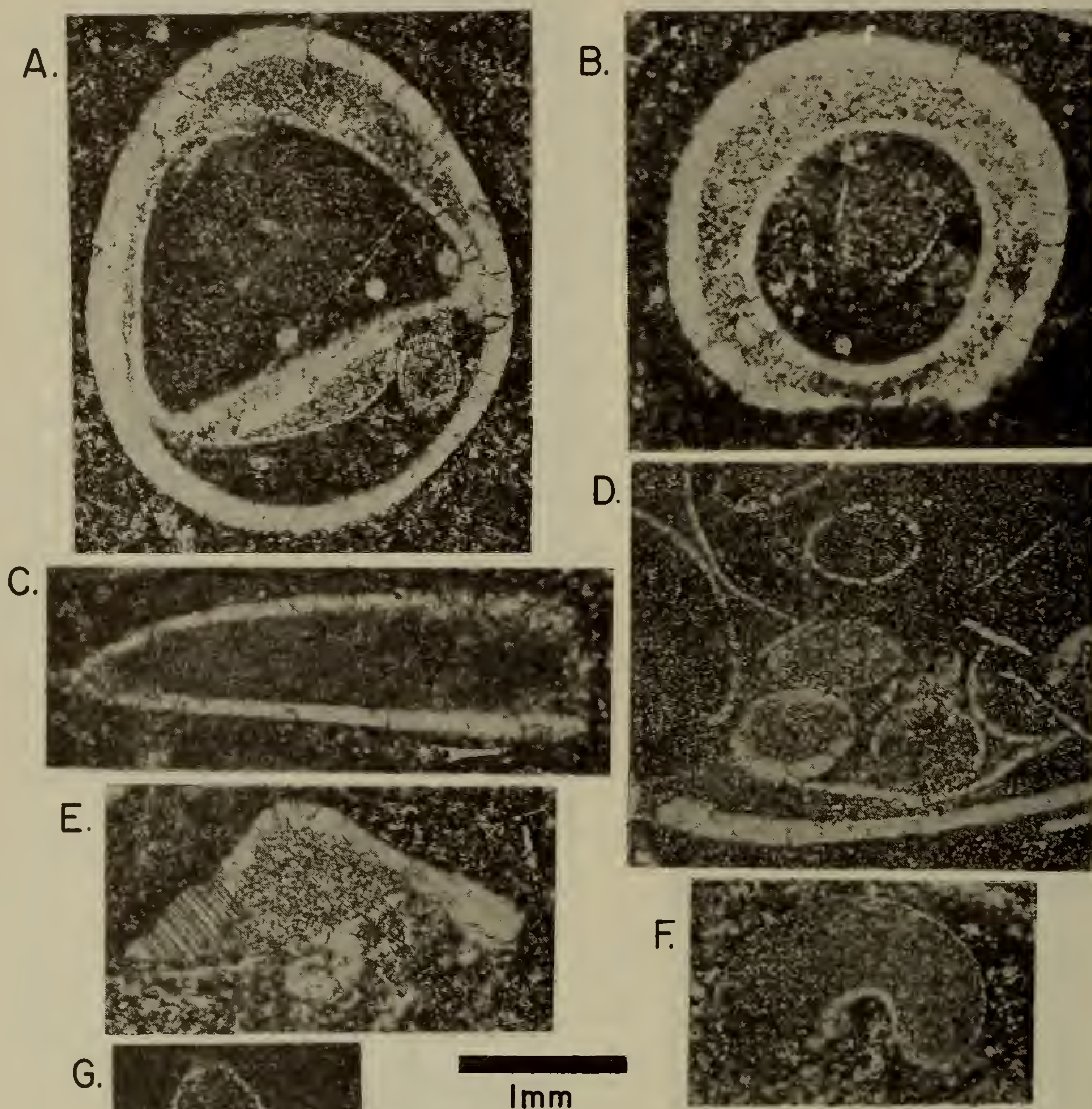

Figure 3.
Photomicrographs of thin sections taken with plane light. Scale bar is 1mm. A. Orthotheca princeps nested in Allatheca communis; B. nested individuals Circotheca cylindrica; C. longitudinal section of indeterminate hyolith; D. recystallized hyolith cross-sections and fragments; E. cross-section of hyolith operculum(?) F. section through about ½ whorl of Aldanella cf. A. attleborensis; G. longitudinal section through Stenotheca(?)

of blocky sparry calcite (Figure 3). This blocky spar may have filled voids created by the dissolution of original aragonitic shell material (James and Klappa, 1983). Although there are some remaining void-lining, finely crystalline, prismatic crystals most have been recrystallized or obliterated by recrystallization of micrite matrix. Common stylolites can be identified in thin section by truncation of fossils.

Three microfacies are present: 1) very sparsely fossiliferous to unfossiliferous micrite occurring in thin beds, irregular patches, or as intraclasts (Figure 2); 2) sparsely to moderately fossiliferous biomicrite or wackestone as thin beds or burrow (?) fillings that grade into unfossiliferous micrite or; 3) hyolith-biosparite, grainstone, or packstone. The latter microfacies is composed of abundant hyoliths and fragments. Some of these thin biosparite laminae overlie irregular scour(?) surfaces but more commonly they are found as isolated pockets or nests of fossils surrounded by micrite intraclasts or biomicrite. The inter-relationships of microfacies are illustrated in acetate peels of polished slabs cut parallel to bedding (Figure 2). Irregular masses of chert and epidote have secondarily altered areas of biomicrite and micrite.

Tubular and conoidal fossils do not show a strong preferred orientation. Long axis orientations (Figure 5) suggest only weak current sorting. Nested, or cone in cone hyolith specimens are fairly abundant and, along with the considerable abundance of fossil fragments indicate moderate post-mortem movement by currents. Long axes of many specimens of hyoliths are steeply inclined to bedding suggesting that the bottom may have been somewhat hummocky or irregular.

Environment of Deposition

A shallow subtidal environment is inferred for limestone strata at Nahant. Evidence supporting this conclusion is:

1) irregular laminae of biosparite and biomicrite with fragmental disarticulated shelly fossils
2) possible intraclasts around and between which accumulated biosparite and biomicrite
3) possible laminate and small domal or digitate stromatolites (now chertified)
4) a relatively diverse (for Early Cambrian) calcareous macrofauna.
5) a general association of Lower Cambrian hyoliths with shallow water facies in areas where more complete stratigraphy allows a confident asessment of paleo-environment. (especially Hoppin Hill and equivalent strata in Newfoundland).

This environment was interrupted by short term fluctuations in energy level, possibly storms, to produce the bioclastic rich layers mixed with intraclasts. Micrite accumulated during quiescent periods and during these periods stromatolites probably covered portions of the bottom. Limestone deposition ceased when the influx of fine clastics, possibly associated with concomitant deepening of the platform halted biogenic carbonate production. The thin nodular horizons result from a near balance between carbonate and clastic deposition. At no time during the deposition of the Nahant strata did a substantially uplifted extrabasinal source exist. Quartz silt is the coarsest extrabasinal material present. A stable platformal setting is further indicated by basal quartzarenite over granitic basement at Hoppin Hill and by quartzarenites in proximity to, but not in actual depositional contact with, the Weymouth Formation east of Mill Cove (Billings, 1982).

Paleontology

Foerste (1889) was the first to describe hyoliths from the Cambrian strata of Nahant. Louis Agassiz in 1850 and Sears in 1887 had also noted the presence of fossils in Nahant limestones (Grabau, 1900). By 1900 a well documented Lower Cambrian fauna was known from Hoppin Hill, Weymouth, Nahant, and from glacial cobbles and boulders at several localities. These Early Cambrian fossils are of great interest because they include some of the first shelled organisms to appear on earth. My paleontological studies, and those of my students, utilize thin sections, acetate peels, acid etched blocks, and specimens obtained by dissolving blocks. The discussion below lists and briefly describes taxa that have been found or reported from the limestone beds at Nahant.

Brachiopods

Grabau (1900) illustrated two species of inarticulate brachiopods, *Obolella* cf. *O. atlantica* and *Paterina bella*. I have found several internal molds and cross-sections indicative of the former species, but the material is of poor quality. *Obolella atlantica* occurs with the trilobite *Strenuella* at both Hoppin Hill and Weymouth.

Hyoliths

The Nahant fauna is dominated by bilaterally symmetrical, conical fossils known as hyoliths. The taxonomic status of hyoliths is uncertain, with some workers (Runnegar and Pojeta, 1974) regarding them as an extinct phylum, and some (Marek and Yochelson, 1976; Yochelson, 1978) placing them as an extinct class in the mollusca. A typical hyolith skeleton is a straight or slightly curved, rounded, flattened, or triangular cone closed by an operculum. Two curved calcareous "whiskers"or appendages extending from the aperture (Yochelson, 1974) may have assisted in locomotion and or feeding (Figure 4). These creatures were probably benthic detritus feeders (Marek and Yochelson, 1976). Much of the detailed systematic work on hyoliths, particularly those from the Lower Cambrian, is based on specimens from the well exposed strata of the Siberian Platform (Raaben, 1981; Matthews and Missarzhevsky, 1975). The taxonomic assignment of sectioned hyoliths to genera

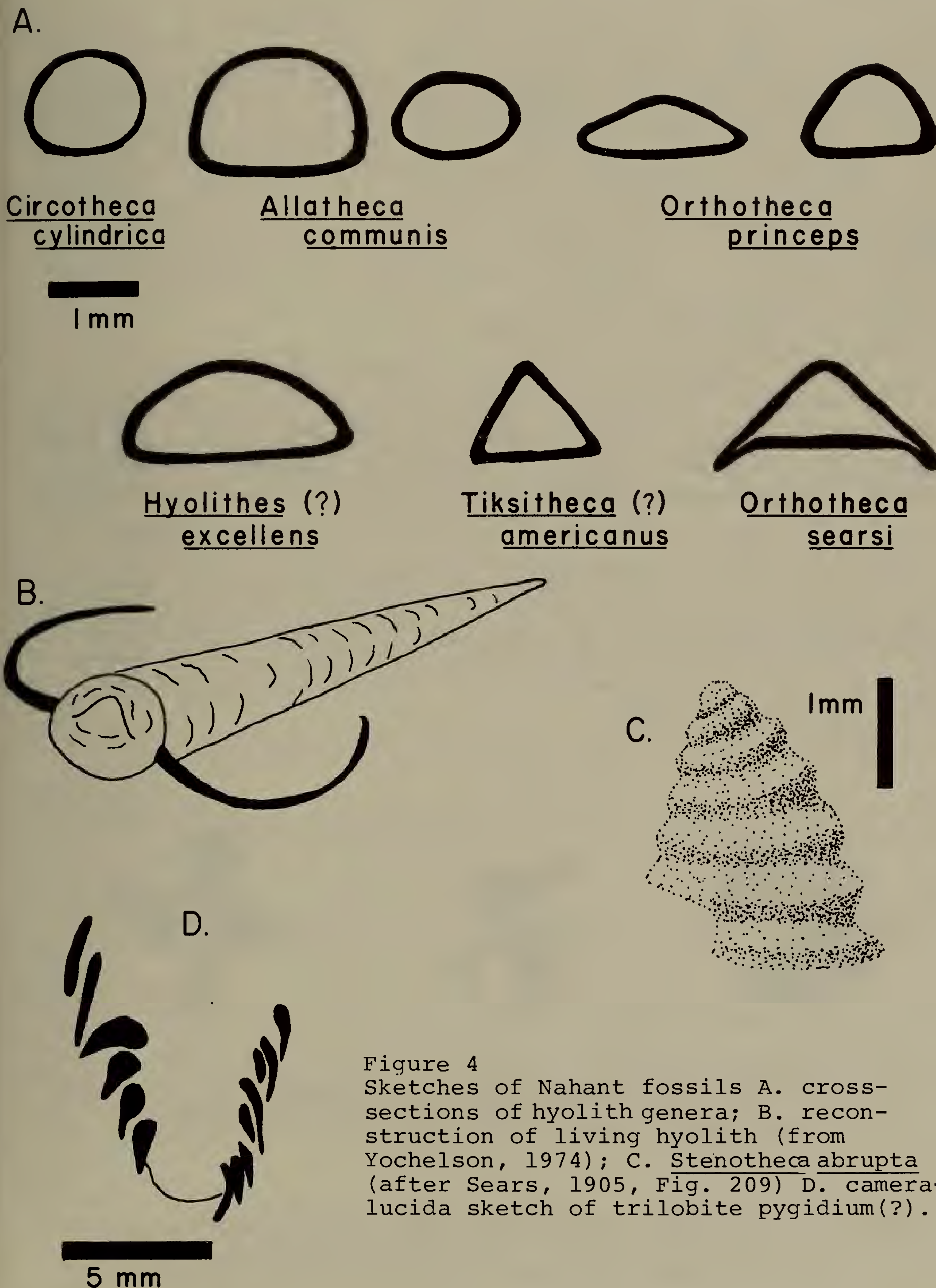

Figure 4
Sketches of Nahant fossils A. cross-sections of hyolith genera; B. reconstruction of living hyolith (from Yochelson, 1974); C. Stenotheca abrupta (after Sears, 1905, Fig. 209) D. camera-lucida sketch of trilobite pygidium(?).

and especially to species can be very difficult as one must constantly consider the geometric effect of obliquity of the section to the long axis of the specimen. Sections perpendicular to the long axis are most useful. The sections illustrated in Figure 4 are diagnostic for Nahant species. The apical angle is also a variable but taxonomically useful character. Grabau (1900) listed 8 species of hyoliths from Nahant of which 6 are probably valid. Some species are illustrated in Figures 2 and 3.
Identified species of hyoliths include:
Circotheca cylindrica (Grabau)
Allatheca communis (Billings)
Orthotheca searsi (Grabau)
Orthotheca princeps (Billings)
Tiksitheca(?) americanus (Billings)
Hyolithes(?) excellens (Billings)

Stenothecoids and Aldanellids

Stenothecoids are small, asymmetrical, inequivalved, bivalved organisms that possibly represent an extinct molluscan class. Specimens of Stenotheca abrupta (Shaler and Foerste) are present on acid-etched blocks and in thin section (Figures 3 and 4). Specimens have also been found at Hoppin Hill (Grabau, 1900).

Aldanella cf. A. attleborensis (Shaler and Foerste) was found in Nahant strata by Bingham (1977) and Landing and Brett (1982). This tiny, loosely coiled trochisprial shell (illustrated in Figure 3) is considered by Runnegar and Pojeta (1974) to be the oldest known gastropod. Other workers (Yochelson, 1978; Landing and Brett, 1982) do not consider them to be gastropods, but possibly coiled worm tubes.

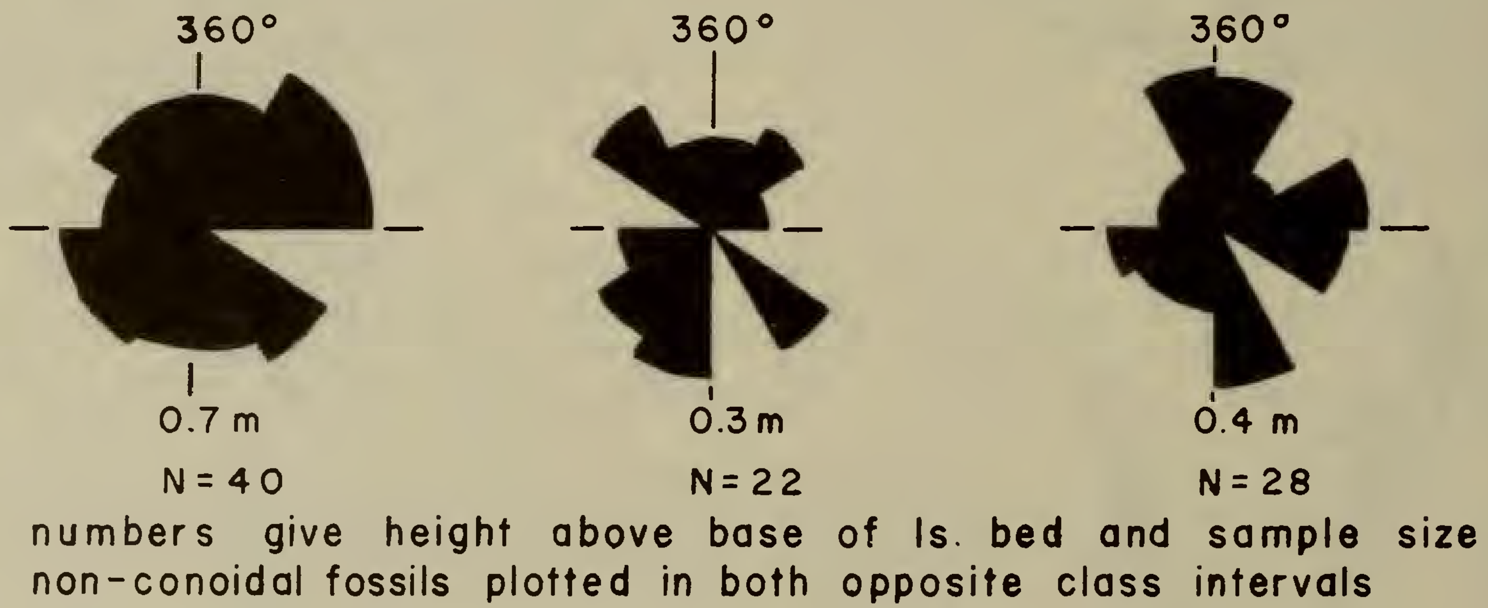

Figure 5. Circular histograms (equal area) showing orientations of long axes of hyoliths measured in slabs and peels parallel to bedding.

Miscellaneous Specimens

Small straight or nearly straight tubes probably represent *Coleoloides* sp., a possible annelid worm tube. Other enigmatic fragments may represent pieces of previously mentioned taxa or with imagination they may also be referred to other known species. One such specimen worth noting is illustrated in Figure 4C. This segmented organism was found *in situ* on a bedding surface in the upper limestone. It consists of 6 or 7 segments with spine-like extensions forming the lateral margins. The specimen is the proper size and shape to be the pygidium of *Callavia*; however it is to poorly preserved to be identified with certainty.

Acknowledgments

I would like to thank my secretary Mary Meehan, for typing two camera-ready manuscripts for me under more than a little pressure. Two of my students, Michael Bingham and Michael Penzo, completed undergraduate research projects on the Lower Cambrian of Nahant and Weymouth respectively. Their work, which brought several new ideas to light, will be covered more fully in a paper now in preparation.

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Road Log

Mileage	
0.0	Inn at Danvers Parking lot.
0.05	Exit onto Dayton St. on north side of Inn; 180° right on Dayton onto Armory Rd.
0.1	Left at sign for "Boston-Newburyport Rt. 95"
0.2	Right at intersection; follow sign for Rts. 95 and 1 Boston and Topsfield"
0.3	Left at intersection; follow sign for "95 south and Rt. 128 Burlington and Gloucester"; enter onto Rt. 95; continue south on 95/Rt. 1.
4.0	Cross over Rt. 128; continue south.
4.6	Exit on Rt. 129 east at sign "129 Lynnfield/Lynn."
4.9	Left turn at stoplight onto Rt. 129 east.
5.5	Enter rotary and bear right; follow Rt. 129.

8.8	Bear left on Rt. 129 at sign for Marblehead/Swampscott
9.2	Left (east) on Rt. 129.
9.8	Right on Rt. 129
9.9	Bear right on 129
11.3	Cross Rt. 1A
11.5	Right (south) on Lynn Shore Drive
12.5	Enter rotary; go about 180° around rotary and bear right onto Nahant Rd. (causeway); continue on Nahant Rd. past Little Nahant through town of Nahant.
16.1	Bear left on Nahant Rd.
16.3	Bear left to gate at entrance to Northeastern University's Marine Science and Maritime Studies Center and Edwards Laboratory.

Figure 6
Geologic map of East Point, Nahant (on following page). Thin dikes not shown at true thickness. Rectangular or circular areas represent gun emplacements or pill boxes.


N
35°
34°
Paved
Area
42°
38°
0
100m

16.35 Go through gate and park in paved area to right in front of lab. We will assemble in front of lab. Please do not wander off until group is assembled. There is no public parking anywhere near East Point. For future visits you must obtain permission before your visit, to enter the gate and to park, from the director of the Marine Science and Maritime Studies Center at Nahant. The Town of Nahant controls access to some areas of the cliffs and you should also obtain permission from the town well in advance of a visit with a larger group. Figure 6 presents a geologic map of East point and Figure 7, 8, and 9 are enlarged portions on the same map that show points of interest along the cliff. Items of geologic interest are discussed in the following section.

Cliff Log

Station

1 Looking south from this vantage point across the Boston Basin you can see the Boston skyline (250^{o}), the Blue Hills that form the southern margin of the Boston Basin, (220^{o}), and the drumlins in Boston Harbor (190^{o}). To the northeast you can see Cambrian strata dipping to the northwest.

2 Well exposed 1m thick basalt dike with distinctive chilled margins and gabbro xenoliths.

3 Nahant Gabbro with xenoliths(?).

4 Dikes galore! Only thicker dikes are shown in Figure 7 and thicknesses are not shown to scale. Many dikes have phenocryst-rich interiors and well defined chilled margins. Differential weathering and plucking leaves a trench or chasm where basalt is removed.

5 Dolerite sill cut by dikes.

6 Irregular ultramafic dikes with abundant xenoliths.

7 Argillite of Weymouth Formation with Ca-silicate nodules.

8 There are 3 prominent joint sets 1) strike 130^{o} -140^{o}, dip 90^{o}; 2) strike 235^{o},dip 40^{o} s; 3) strike 100^{o}, dip 55^{o} S. Many dikes and faults have approximately the same orientation as joint sets 1 and 3.

9 Highly silicified and metamorphosed limestone. Fossils in poor condition.

10 Nahant Gabbro. Kaye (1965) discusses several unusual features of the gabbro.

Figure 7. Geologic map showing stations 1-18.

11 Shear zone with breccia and some mineralization.

12 Fault with silicified breccia weathered in relief.

13 Argillite with nodules.

14 Thick dike with interesting assortment of xenoliths.

15 Good view of Blue Hills, Boston Harbor, and Atlantic Ocean. Please don't fall off cliff!

16 Good place to study weathered nodular argillite.

17 Very thick dike with abundant small xenoliths.

18 This fault must have significant displacement as the limestone bed on the left block is not present on the right block.

19 Deep chasm and small unsafe natural bridge produced by removal of dike. Bridge is formed by small wedged part of dike.

20 Argillite with very abundant nodules defining bedding planes. Some layers of nodules coalesce to form thin limestone beds.

21 Highly silicified nodular argillite and thin limestone. Note the delicate laminations in cherts and mudstones.

22 Dolerite sill.

23 Excellent fossils in limestones with thin, irregular, greenish chert layers. Please refrain from indiscriminate bashing of weathered bedding surfaces. See detailed cross-section of this bed given in Figure 1.

24 Heavily sheared limestone.

25 Low angle fault. Follow the breccia zone toward the water and along the underside of cliff to see breccia and slickensides. You can crawl into cliff along fault plane.

26 Yet another thick dike.

27 Stand here and search for limestone bed in wall across chasm. Obviously another fault of significant displacement.

28 Fault breccia. Note sill across chasm faulted into contact with breccia.

29 Dolerite dike with distinctive dark reddish-brown weathering surface.

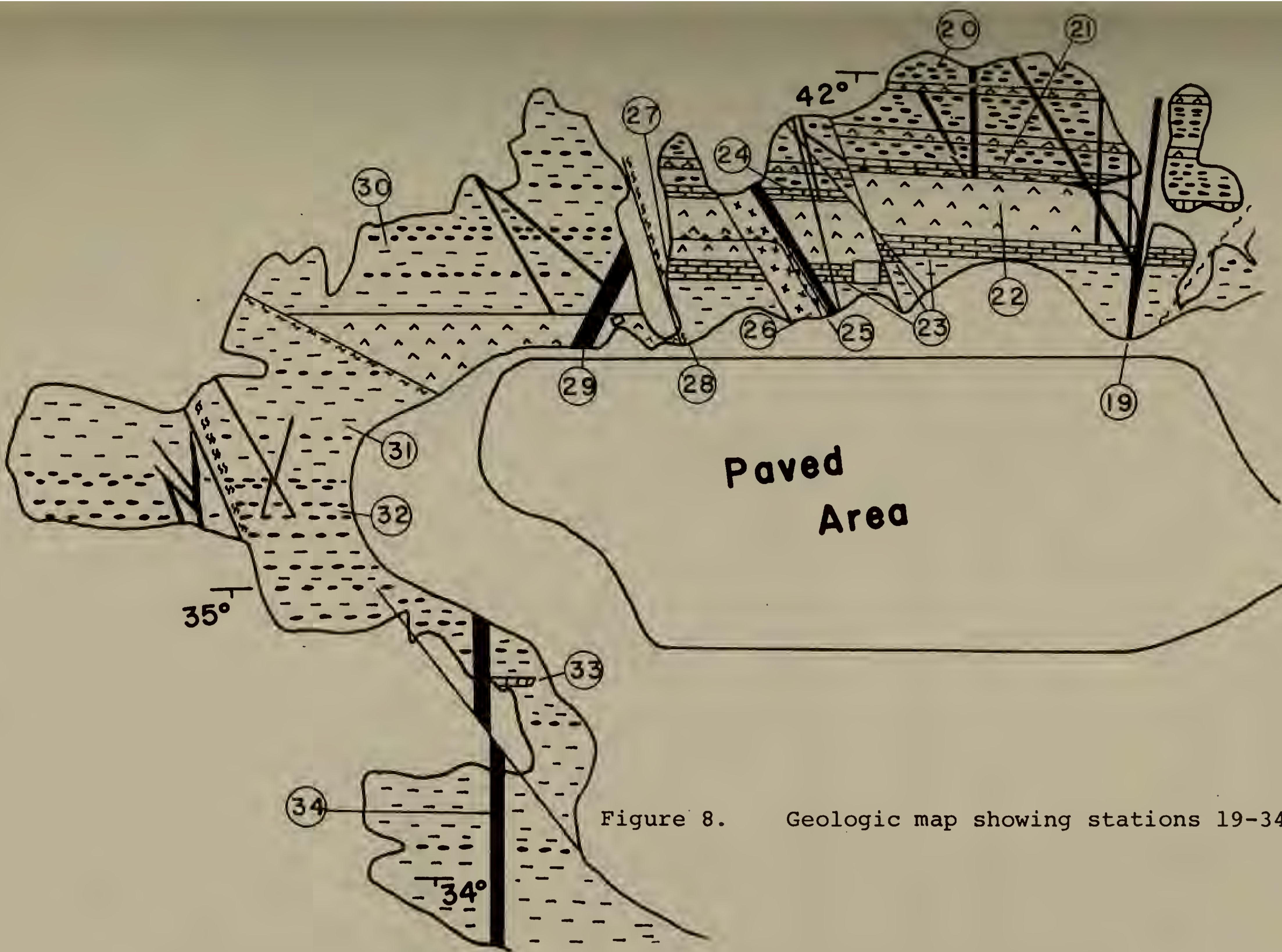

Figure 8. Geologic map showing stations 19-34.

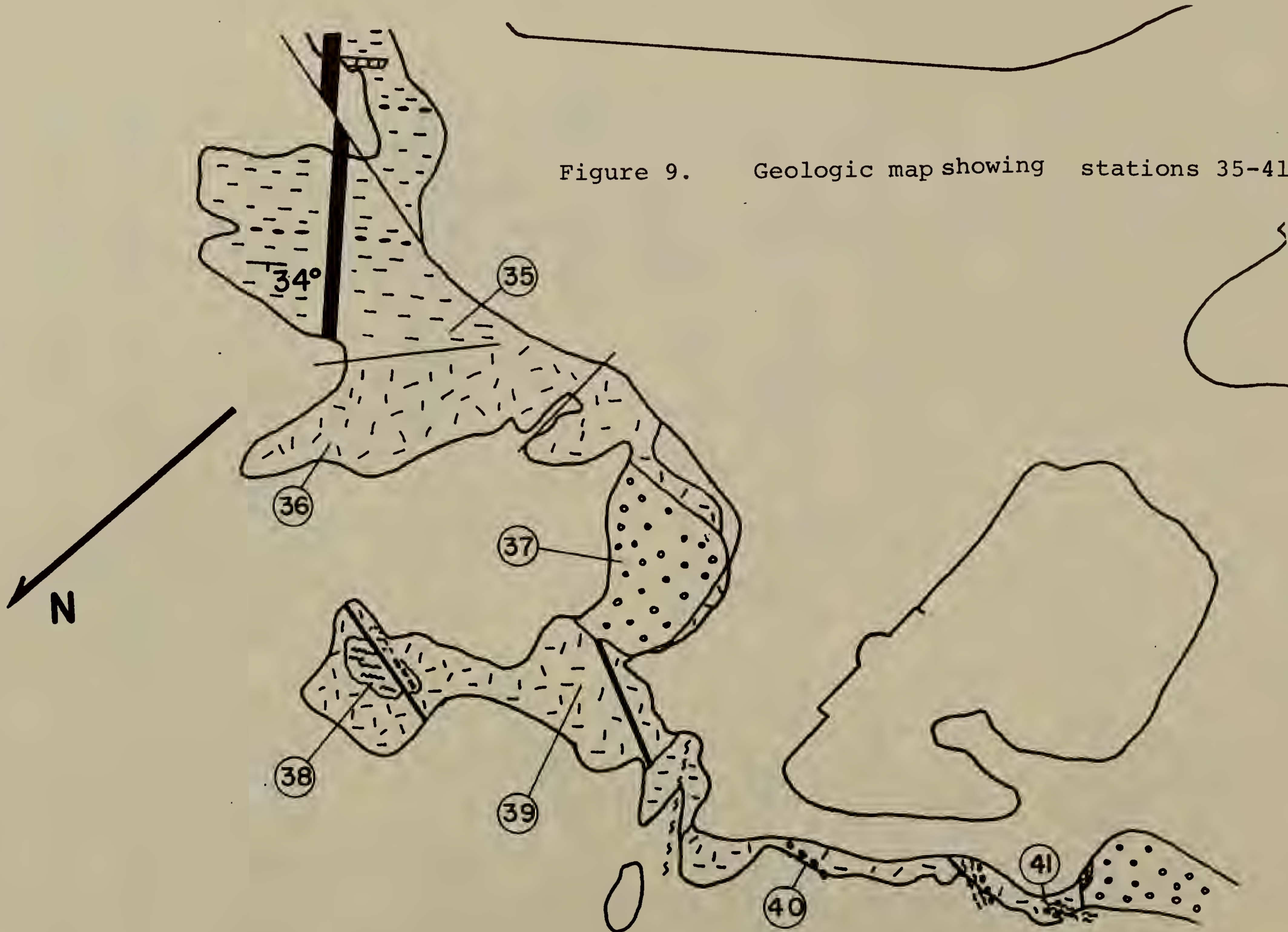

Figure 9. Geologic map showing stations 35-41.

30 Argillite with sparse nodules. Small dikes and shear zones not mapped.

31 Good exposure of dip surface of nodular argillite. Search for trilobites! Well exposed small dikes on this surface.

32 Argillite with abundant nodules.

33 Thin brecciated nodular limestone bed.

34 Stand on outer edge of dike at cliff and look east along strike. Note that piece of dike across gorge is offset by visible fault.

35 Nahant Gabbro in fault(?) contact with Weymouth Formation.

36 Nahant Gabbro.

37 High energy pocket "beach". P.S. Rosen of Northeastern University is currently studying cobble shape development and transport mechanisms on this beach.

38 Strongly foliated Weymouth xenolith in gabbro. Not accessible at high tide.

39 Many small dikes cutting gabbro, most not mapped.

40 Very unusual rounded and embayed quartzite (?) xenoliths in closely spaced dikes. Are these samples of older meta-sedimentary basement or Cambrian quartzarenites?

41 Carbonate and quartzite xenoliths associated with shear zone. Continue to left across beach to return to parking lot. End of trip.

THE MARLBORO FORMATION IN ITS TYPE AREA AND ASSOCIATED ROCKS JUST WEST OF THE BLOODY BLUFF FAULT ZONE, MARLBOROUGH AREA, MASSACHUSETTS

by Richard G. DiNitto[1], J. Christopher Hepburn[2], Kelly Durfee Cardoza[2], and Malcolm Hill[3]

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Introduction

The Marlboro Formation underlies the eastern portion of the Nashoba Block, an exotic terrane (Zen, 1983a) in eastern Massachusetts bounded by the Clinton-Newbury and Bloody Bluff Fault Zones (Figure 1). Movement on these faults was likely large, although their displacement cannot yet be established since no definite correlation of Nashoba Block rocks can be made with rocks elsewhere. The Nashoba Block has been metamorphosed to the upper amphibolite facies in contrast to the Boston Platform and the eastern Merrimack Trough (Lyons, et al., 1982) which have generally undergone only low grades of metamorphism. The Nashoba Block is underlain by a thick series of metabasaltic rocks on the east (Marlboro Formation) and metamorphosed sediments and volcanogenic sedments on the west (Nashoba Formation). These stratified rocks have been intruded by a distinctive sequence of Ordovician to Silurian calc-alkaline intermediate plutons (Sharpners Pond, Assabet and Straw Hollow Diorites) and peraluminous granites (Andover) (Zen, 1983b; Zartman and Naylor, 1984). The structure in the Nashoba Block is complex. The terrane is cut by numerous faults and ductile shear zones (Barosh, et al., 1977; Bell and Alvord, 1976) and many exposures exhibit multiple fold generations.

Emerson and Perry (1907) first used the name Marlboro Formation for mafic rocks in the upper part of the Blackstone Series of Rhode Island. Later, Emerson (1917) redefined the Marlboro to include black biotite schists and hornblendic schists of eastern Massachusetts, and described a type locality along the north side of Main Street in the City of Marlborough, Massachusetts. However, Emerson described a Brimfield-type Schist just north of the Marlboro Formation and included the type Marlboro section in the Brimfield-type Schist on the 1917 state geologic map. Emerson (1917) included, in the Marlboro, rocks from the Boston Platform in both eastern Massachusetts and Rhode Island, but noted that these rocks might be more appropriately divided into distinct units. The Marlboro rocks in Rhode Island were eventually separated as the Huntinghill Greenstone of the Blackstone Series by Quinn, et al. (1949). Bell and Alvord (1976) redefined the Marlboro in eastern Massachusetts to include all the mafic rocks in the stratigraphic interval between the Bloody Bluff Fault and the base of the Shawsheen Gneiss to the northwest. Marlboro-type rocks east of the Bloody Bluff Fault were included into the Middlesex Fells Volcanic Complex and the Greenleaf Formations (Bell and Alvord, 1976). Bell and Alvord further subdivided the Marlboro into a lower unnamed member at the base, overlain by the Sandy Pond Amphibolite Member. Three informal members of the Marlboro Formation above the Sandy Pond Amphibolite Member have been mapped in the Marlborough quadrangle (DiNitto, 1983). These three members appear

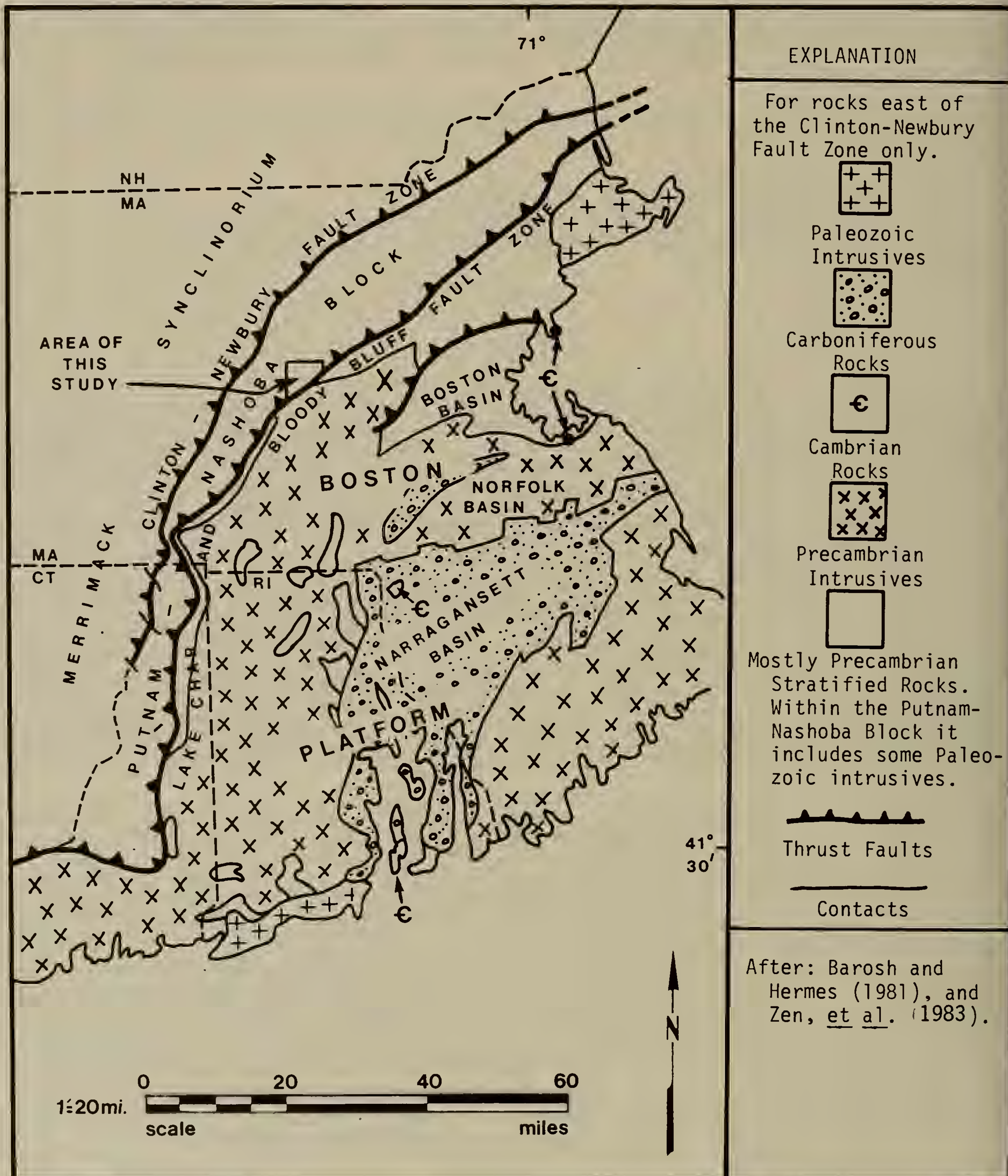


FIGURE 1 : Simplified Geologic Map of Southeastern New England.

to be at the same stratigraphic level as the Shawsheen Gneiss mapped by Bell and Alvord (1976) to the northeast (Figure 2). Lithologically, however, the three members in Marlborough do not appear to be similar to the Shawsheen Gneiss, making Bell and Alvord's (1976) definition difficult to apply in the Marlborough area.

Other work on the Marlboro in this region included mapping by Hansen (1956) in the Maynard and Hudson quadrangles, Nelson (1975a) in the Framingham quadrangle, Hepburn and DiNitto (1978) and Barosh (1978) in the Marlborough quadrangle. Skehan (1968), and Skehan and Abu-moustafa (1976) studied parts of the Marlboro in the Marlborough and Shrewsbury quadrangles during their mapping of the Wachusett-Marlborough Tunnel. Zen (1983b) on the recent state map of Massachusetts has summarized the previous mapping to date.

Our studies were taken to re-map and reevaluate Emerson's (1917) type locality for the Marlboro Formation and to determine whether any of the previously suggested correlations of the Marlboro rocks throughout eastern Massachusetts might still hold. As these studies progressed, other studies of the geochemistry of the Marlboro amphibolites were initiated to investigate the possible environment of emplacement and age for the Marlboro Formation.

Stratigraphy

Two formations have been mapped within the Marlborough area: the Marlboro Formation to the east and the Nashoba Formation to the west (Figures 2, 3). The Marlboro Formation is fault bounded near its presumed base by the Bloody Bluff Fault and is separated from the Nashoba by the Assabet River Fault. Both formations strike northeast and have steep to moderate northwest dips. Repetition of strata is not observed, but some units or portions thereof are cut out by faulting. Bell and Alvord (1976) indicate that the rocks in the Nashoba Block likely form a homoclinal sequence topping toward the northwest. While the structure of the Nashoba Block is complex and topping directions within this sequence are as yet unknown, we will describe the units below in a general southeast to northwest direction away from the Bloody Bluff Fault Zone, consistant with Bell and Alvord's interpretation.

Marlboro Formation

The Marlboro Formation is subdivided into five members (informal) that are composed of interstratified amphibolite, amphibolite gneiss, quartzofeldspathic gneiss, rusty sillimanitic schist, quartzofeldspathic granulite and minor amounts of quartzite, coticule and marble (Figure 4). Approximate thicknesses of each member are given in parenthesis after each members name.

Gneiss Member (600 m) - The gneiss member is comprised predominantly of a medium- to coarse-grained plagioclase-hornblende-biotite schist. These rocks are typically dark colored, black to gray, but weather lighter. Compositional layering is common, resulting from alternating mafic and felsic layers, 1 to 70 cm thick. The layering produces a rock that appears to a banded amphibolite. The schist is dark gray to silvery-gray, fine- to medium- grained and occurs typically as layers (10 to 25 cm) within the gneiss. The gneiss member is located at the eastern margin of the Marlboro Formation and lies directly northwest of the cataclastic rocks of the Bloody Bluff Fault Zone. The member is poorly exposed within the Marlborough area.

FIGURE 2:

REGIONAL GEOLOGIC MAP OF THE NASHOBA BLOCK, MA.

KEY FOR FIGURE 2

INTUSIVE ROCKS

LATE ORDOVICIAN TO EARLY DEVONIAN

- fg? — FITCHBURG GRANITE OF HEPBURN (1978), MUSCOVITE GRANITE OF SKEHAN & ABU-MOUSTAFA (1976).
- ihhg — INDIAN HEAD HILL GRANITE
- shd, aqd, sqd — STRAW HOLLOW DIORITE, shd; ASSABET QUARTZ DIORITE, aqd; SHARPNERS POND DIORITE, sqd
- sgd — SALEM GABBRO-DIORITE
- ag — ANDOVER GRANITE

AGE ?

- gg? — GRAFTON GRANITE (?)

PRECAMBRIAN

- dg — DEDHAM GRANODIORITE

AGE ?

- iu — INTRUSIVES, UNDIVIDED

STRATIFIED ROCKS

PRECAMBRIAN TO ORDOVICIAN

- as — ALTERED AND SHEARED ROCKS
- ch — CLAYPIT HILL FORMATION
- kg — KENDALL GREEN FORMATION
- tb — TADMUCK BROOK SCHIST
- n — NASHOBA FORMATION
- fbg — FISH BROOK GNEISS
- shg — SHAWSHEEN GNEISS
- m — MARLBORO FORMATION
- w — WESTBORO FORMATION

CONTACTS, dashed where inferred

FAULTS, dashed where inferred

THRUST FAULTS, dashed where inferred

QUADRANGLES

1. HUDSON, MA
2. MAYNARD, MA
3. CONCORD, MA
4. SHREWSBURY, MA
5. MARLBOROUGH, MA
6. FRAMINGHAM, MA
7. NATICK, MA
8. GRAFTON, MA

SOURCES OF DATA

CONCORD QUADRANGLE :	BELL & ALVORD (1976), BATTIN (1977).
FRAMINGHAM QUADRANGLE:	NELSON (1975a), BELL & ALVORD (1976). DINITTO (RECONNAISANCE, 1977 & 1978).
HUDSON QUADRANGLE :	HANSEN (1956), BELL & ALVORD (1976). DINITTO (RECONNAISANCE, 1977 & 1978).
MARLBOROUGH QUAD. :	BELL & AVORD (1976), BAROSH (1978), HEPBURN & DINITTO (1978), DINITTO (1983).
MAYNARD QUADRANGLE :	HANSEN (1956), BELL & ALVORD (1976), DINITTO (RECONNAISANCE, 1977 & 1978).
NATICK QUADRANGLE :	NELSON (1975b).
SHREWSBURY QUAD. :	BELL & ALVORD (1976), HEPBURN (1978), DINITTO (RECONNAISANCE, 1977).
ALL QUADRANGLES :	BAROSH, et al. (1977), Zen (1983a).

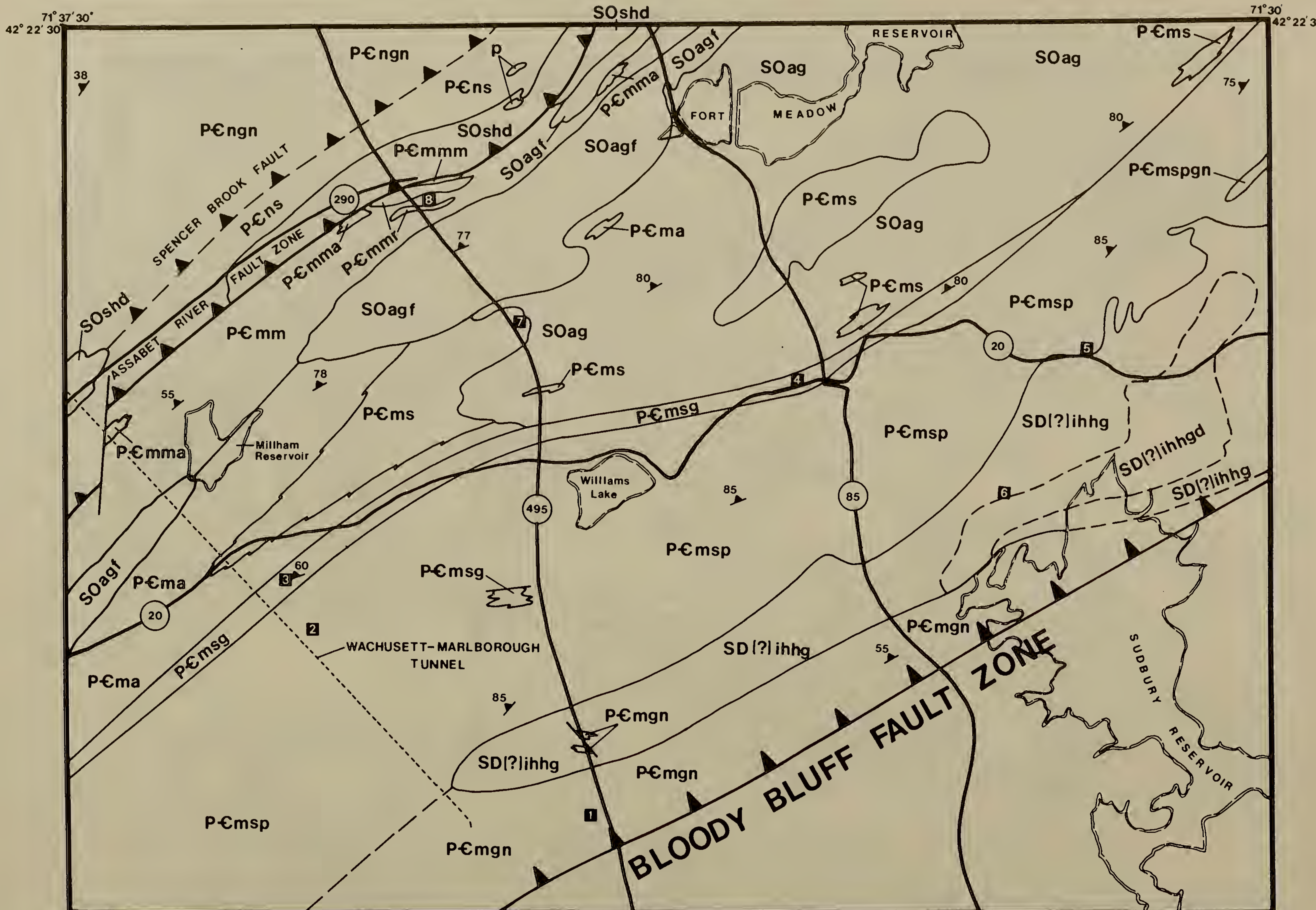


FIGURE 3: GEOLOGIC MAP OF THE NORTHERN HALF OF THE MARLBOROUGH QUADRANGLE, MA

KEY FOR FIGURE 3

INTRUSIVE ROCKS

SILURIAN TO DEVONIAN ?

P	PEGMATITES
SD(?)ihhg / SD(?)ihhgd	INDIAN HEAD HILL GRANITE, with dark phase, SD(?)ihhgd

LATEST ORDOVICIAN TO SILURIAN

SOshd	STRAW HOLLOW DIORITE
SOagf / SOag	ANDOVER GRANITE, with foliated phase, SOagf

STRATIFIED ROCKS

PRECAMBRIAN ? — NASHOBA FORMATION

PCngn	GNEISS MEMBER
PCns	SCHIST MEMBER

PRECAMBRIAN ? — MARLBORO FORMATION

PCmmm / PCmma / PCmmr / PCmm	MILLHAM RESERVOIR MEMBER, PCmm; with mylonite, PCmmm; amphibolite, PCmma; and rusty schist, PCmmr
PCma / PCms	AMPHIBOLITE AND SCHIST MEMBER, amphibolite, PCma; rusty schist, PCms
PCmsg	RUSTY SCHIST AND GRANULITE MEMBER
PCmsp / PCmspgn	SANDY POND AMPHIBOLITE MEMBER, with gneiss, PCmspgn
PCmgn	GNEISS MEMBER

■1 ROAD LOG STOPS

CONTACT

FAULT

THRUST FAULT

Contacts and faults drawn where approximately located, dashed where inferred and dotted where probable. Direction of movement along the faults was not be interpretated from field data.

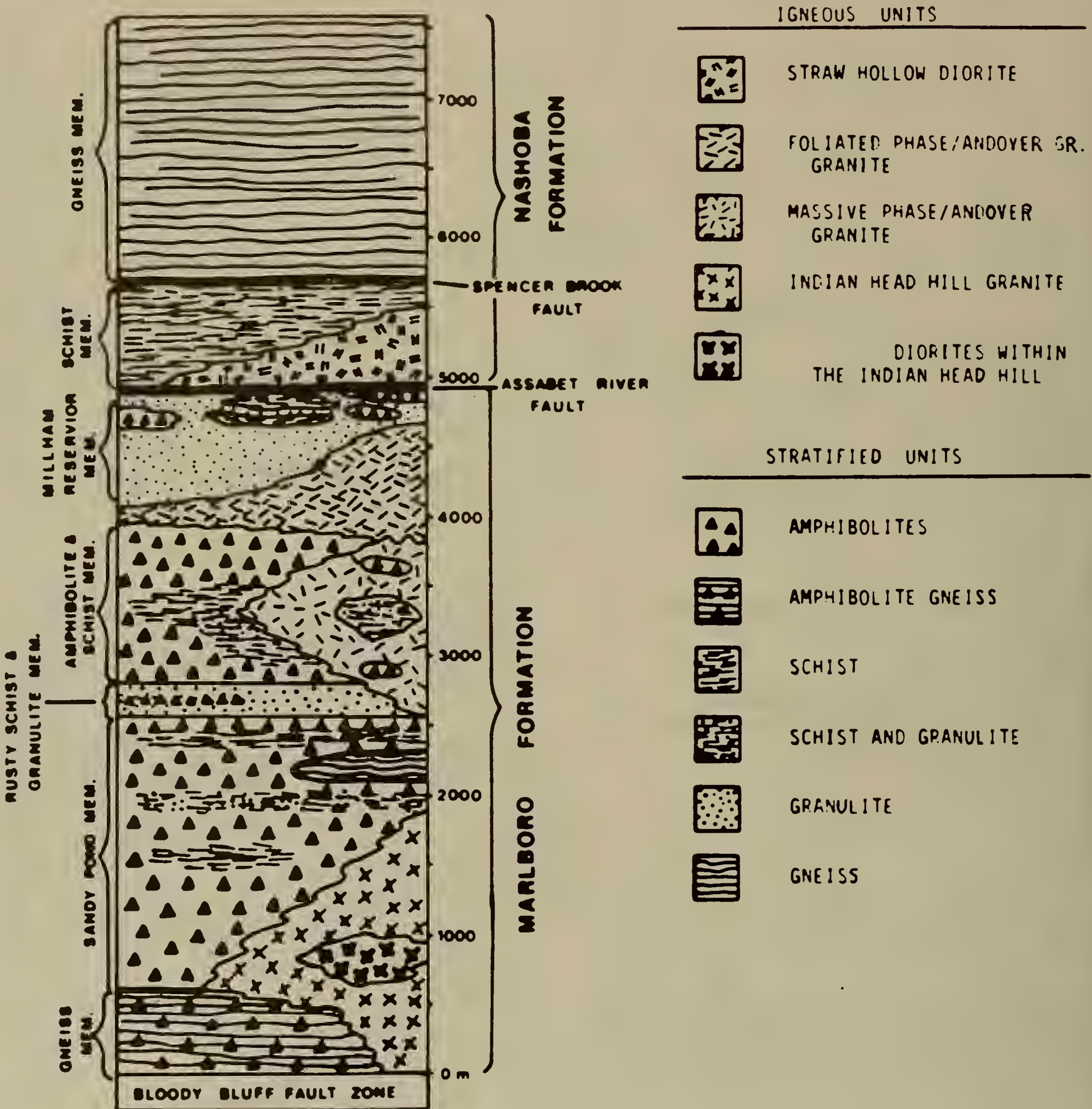


FIGURE 4 : Generalized stratigraphic column for the northwestern portion of the Marlborough quadrangle.

Sandy Pond Amphibolite Member (2000 m) - The Sandy Pond Amphibolite Member (Bell and Alvord 1976) consists predominantly of black, silvery-black, to dark gray, well-foliated, medium- to coarse-grained, massive to thinly layered amphibolites. Locally the amphibolites are interstratified with light gray to dark gray, garnet-sillimanite bearing biotite-quartz-plagioclase granulite and beige to light gray, moderately foliated, medium-grained biotite-sillimanite-plagioclase-quartz gneiss. Some of the foliation observed, results from the interlayering of thin, light gray granulite layers with the darker colored amphibolites and schists.

Amphibolite predominates in the western two-thirds of the member and often contains large lenses or pods of epidote that may reach 60 cm in legth. Schist and granulite layers (1 cm to 1 meter) are exposed mostly in the lower half of the member.

Rusty Schist and Granulite Member (225 m) - The rusty schist and granulite member consists of interlayered garnetiferrous-mica schists and biotite-quartzofeldspathic granulites with minor amounts of amphibolite, schist and quartzite. These rocks are typically medium gray, with a distinct foliation produced by fine laminae. Garnet occurs within the schists as crystals up to 5 cm in diameter or as coticule. Hornblende occurs locally as slender coarse prismatic crystals (up to 1 cm long) within the fine-grained medium gray granulites.

Amphibolite and Schist Member (1200 m) - This member is composed of two mappable submembers: the amphibolite submember and the schist submember. The amphibolite member consists of black to dark gray, fine- to coarse-grained, thick bedded amphibolite. The schist submember consists chiefly of a medium- to fine-grained, silvery-gray, rusty weathering biotite-muscovite-sillimanite-garnet schist. The amphibolite is exposed primarily to the southwest of the Marlborough area, being replaced along strike to the northeast by the schist submember. Much of the schist submember has been intruded by the Andover Granite and is now exposed predominantly as roof pendants.

Millham Reservoir Member (915 m) - The Millham Reservoir member is composed chiefly of fine- to medium-grained poorly- to moderately-foliated, white to light bluish-gray biotite-plagioclase-quartz granulite with minor amounts of rusty-weathering medium-grained muscovite-sillimanite-biotite-schist and coarse-grained, amphibolite. The granulites contain minor amounts of potassium feldspar, hornblende, garnet and sillimanite. The Millham Reservoir member is exposed to the south of and parallel to Interstate 290 and is west of the amphibolite and schist member. Chemical analysis of the Millham Reservoir indicate that SiO_2 varies between 66% and 69%. While it is tempting to assign the granulites of the Millham Reservoir member a volcanic origin (e.g. dacite?), high aluminum, high calcium and low iron values indicate the possibility of a sedimentary origin.

Nashoba Formation

The Nashoba Formation, in the Marlborough quadrangle is separated from the Marlboro Formation by the Assabet River Fault. The Nashoba Formation can be divided into two informal units that consist of rusty-weathering sillimanite-garnet schist and quartzo-feldspathic gneiss. Recent mapping by Bell and Alvord (1976) has distinguished 10 members of the Nashoba Formation primarily in areas to the northeast of Marlborough. The extent of these members in the Marlborough area and to the southwest is not well documented.

Schist Member (760 m) - The schist member consists of a silvery-medium to dark gray, rusty-weathering, medium-grained, garnet-sillimanite schist. Many garnets are megacrystic. Rocks of this member are poorly exposed and highly weathered. The eastern contact of this unit is considered to be the Assabet River Fault, but is not exposed in the field.

Gneiss Member - The gneiss member consists of silvery-dark gray, medium- to coarse-grained, strongly foliated, muscovite-biotite-sillimanite-plagioclase-quartz gneiss. Interlayered, locally are thin (5 to 30 cm) beds of sillimanite-muscovite-biotite schists. The strong foliation within the gneiss is developed by segregational layering that is common to the Nashoba gneisses. The western contact of this member is not located within this study area. Previous workers (Bell and Alvord, 1976; Barosh, 1978) have mapped a fault, the Spencer Brook Fault, at a postition that coincides with the contact between the schist and gneiss members. While field evidence of this fault is lacking in the Marlborough area, aeromagnetic data indicates a possible lineament near the contact that may be the result of a fault (Castle, et al., 1976).

Plutonic Rocks

Three distinct igneous units are differentiated within the Marlborough area. These are the Indian Head Hill Granite, the Straw Hollow Diorite, and the Andover Granite.

Indian Head Hill Granite

The Indian Head Hill Granite includes both a granitic phase and an older dioritic phase (see Hill, et al., this volume). The rocks that form the main pluton are medium gray, fine- to medium-grained, non-foliated, equigranular, biotite granites. The dioritic phase includes a biotite-potassium feldspar-hornblende-plagioclase diorite and a quartz-diorite. Both diorites are medium gray to black, locally foliated, and medium- to coarse-grained. Locally, the diorite is seen as xenoliths (up to 0.5 m in length) within the granite.

The Indian Head Hill Granite intrudes the eastern two members of the Marlboro Formation and is named after exposures at Indian Head Hill in Marlborough. Emerson first included these rocks within the Dedham Granodiorite, but had noted the differences of the exposures at Indian Head Hill.

Andover granite

The Andover Granite, originally described by Clapp (1910) for exposures in Essex County, Massachusetts is the largest pluton within the Nashoba Block (Zen, 1983b). Within the Marlborough area, the Andover can be divided into two phases: a massive and a foliated phase. The massive phase consists of white to beige, coarse-grained, garnet-muscovite granite to granodiorite. The foliated phase consists of light to medium gray, medium- to coarse-grained, strongly foliated, garnet-muscovite-biotite granite that is mildly to strongly cataclastic in texture. The amphibolite and schist member and the Millham Reservoir member are both intruded by the Andover Granite.

Straw Hollow Diorite

The Straw Hollow Diorite, also referred to as the Assabet Quartz Diorite by Barosh (1978) and Bell and Alvord (1976), was first identified by Emerson (1917) and later mapped by Hansen (1956) within the town of Hudson. The Straw Hollow Diorite consists of a fairly uniform, equigranular, medium-grained, dark bluish-gray, weakly-foliated, biotite-quartz-plagioclase-hornblende diorite. The Straw Hollow Diorite intrudes both the Nashoba and Marlboro Formations along the Assabet River Fault.

Geochemistry of the Marlboro Formation Amphibolites

Despite the fact that the Marlboro amphibolites are metamorphic rocks,the geochemical data as a whole are quite coherent and retain igneous information. The alkali and alkaline-earth elements (K, Rb, Cs, Sr, Ba) have to some extent mobilized during metamorphism, causing the scatter seen on Figure 6, for these elements. The major element compositions (see representative data in Table 1) indicate the basalts were mildly alkaline to high alumina basalts (TiO_2 = 0.8-2.3%, Al_2O_3 = 15.5-18.5%, Mg numbers = 0.57-0.66). They are slightly Light Rare Earth Element (LREE) enriched (Figure 5); that is consistent with the higher concentrations of the incompatible elements compared to the more to compatible (left to right) elements on the "spidergram" (Figure 6). Also note on Figure 6 the distinctly lower Ta and Nb contents, relative to Th and Ce, for most samples for which data is available. This pattern is characteristic of basalts erupted at convergent plate boundaries or in marginal basins (Baker, 1984; Hole, et al., 1984; Marriner and Millward, 1984).

TABLE 1

WT. %	K1	K2	K3	K7	K9
SiO_2	47.57	45.99	48.43	47.06	49.18
TiO_2	2.22	1.64	2.04	1.33	1.75
Al_2O_3	15.30	16.72	16.15	18.37	13.30
*Fe_2O_3	12.65	11.29	11.30	9.09	11.17
MnO	0.40	0.34	0.35	0.20	0.25
MgO	8.12	10.21	6.66	7.71	10.12
CaO	9.97	9.80	11.02	12.28	11.27
Na_2O	2.95	2.74	3.79	3.34	2.35
K_2O	0.24	0.60	0.18	0.28	0.19
P_2O_5	0.24	0.18	0.18	0.28	0.14
Total	99.66	99.51	100.11	99.95	99.71

* Total Iron as Fe_2O_3

Neodymium isotope studies of 10 whole rocks (Hill, et al., 1984) did not yield an isochron, indicating either heterogeneity of the mantle source or, more likely, variable contamination with LREE-enriched crustal rocks. The Nd evolution diagram (Figure 7) shows clearly that the basaltic protoliths for the amphibolites formed by melting of a long-term LREE-depleted mantle source, quite similar to the source of most Mid-Ocean Ridge Basalts (MORB) and arc basalts today. Also shown on Figure 7 is the documented range in 143/144Nd with time for a MORB-source mantle, taken from Hart and Brooks

FIGURE 5

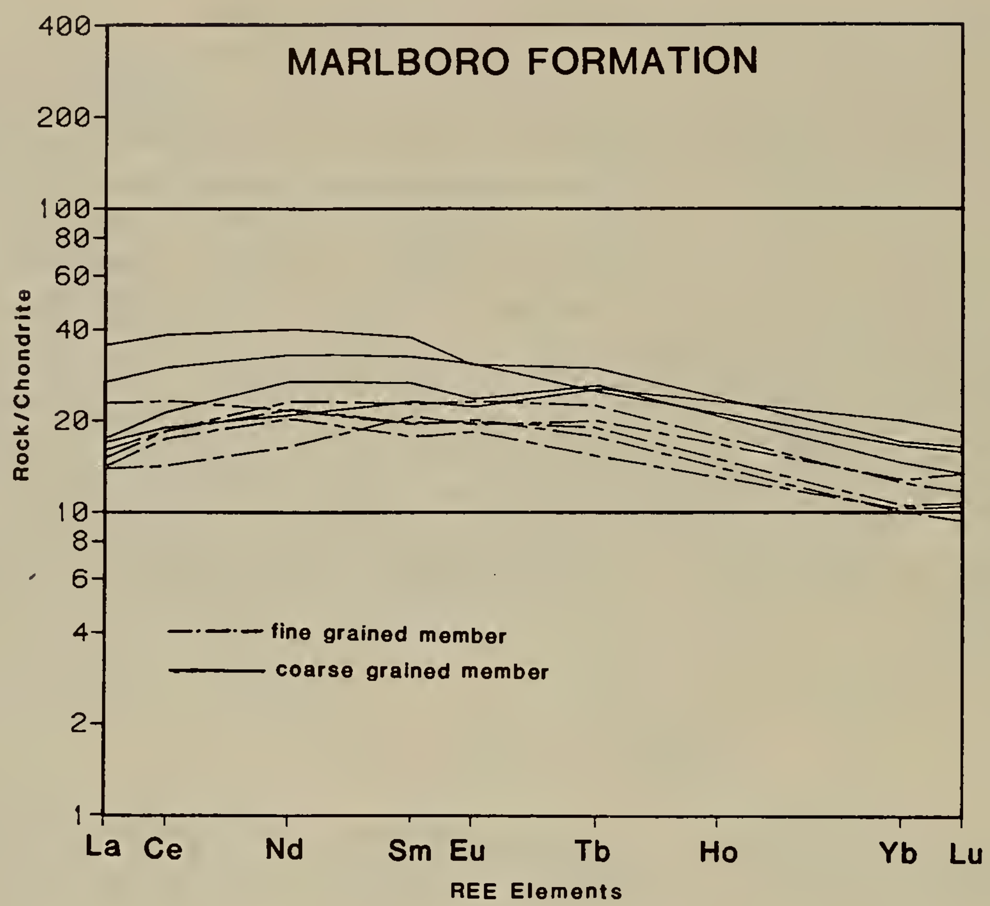


FIGURE 6

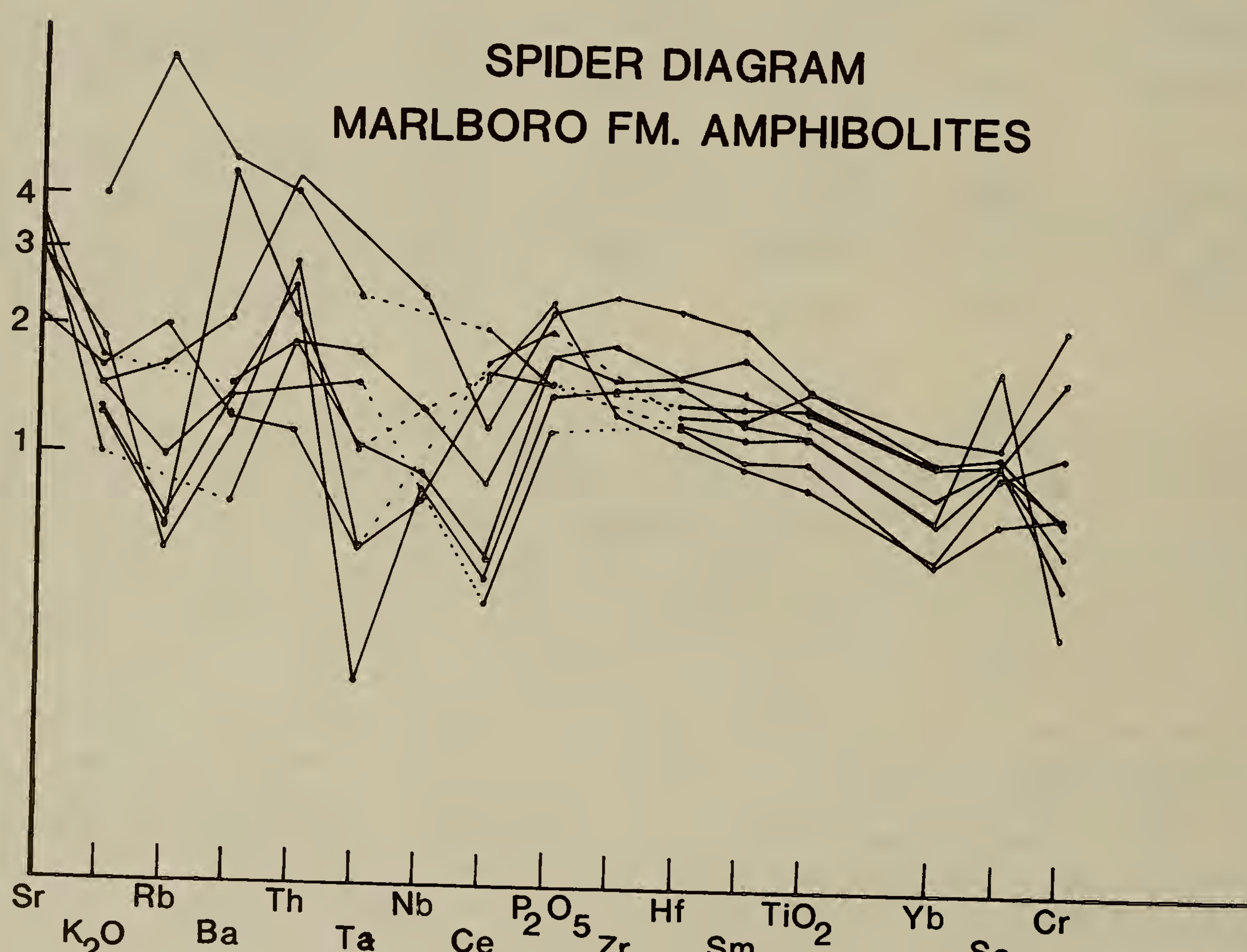
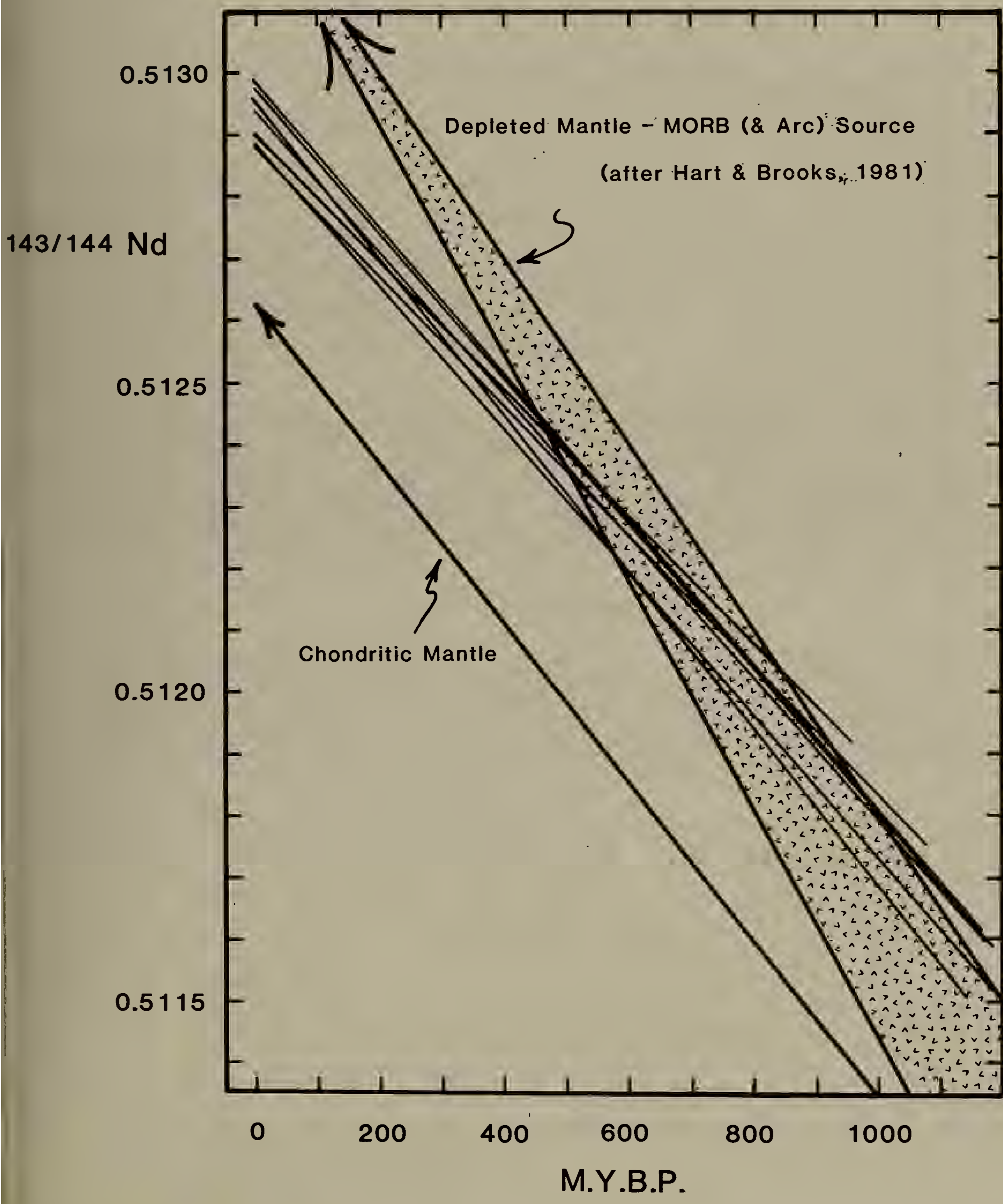


FIGURE 7:
ND EVOLUTION DIAGRAM
MARLBORO FM. AMPHIBOLITES



(1981); this source should also apply to a sub-arc mantle. Model ages (the times of intersection of the rock 143/144Nd growth curves with this depleted mantle range) confirm a pre-450 m.y. age for the Marlboro. The most likely age lies in the range 450 to 550 m.y. because older intercepts are more likely to reflect contamination with LREE enriched crust. However, Late Proterozoic ages cannot be ruled out based on the present data.

Correlations and Ages

Locally, the Marlboro Formation can be correlated with rocks mapped elsewhere in the Nashoba Block (Figure 2). The Sandy Pond Amphibolite Member along strike to the northeast is correlative with Hansen's (1956) and Nelson's (1975) Marlboro Formations, and is correlative with Bell and Alvord's (1976) Sandy Pond Member. The Gneiss Member of the Marlboro Formation, which is east of the Sandy Pond Amphibolite Member is correlative with the lower unnamed member of the Marlboro of Bell and Alvord (1976). The three members of the Marlboro Formation west of the Sandy Pond Member have no clearly identifiable correlatives. To the southwest, the Marlboro Formation is cut out of the section by the Bloddy Bluff Fault Zone in the Grafton quadrangle. Regionally, the Marlboro Formation is correlative with the Quinnebaug Formation of Connecticut (Dixon, 1968; Zartman and Naylor, 1984).

The age of the Marlboro Formation is considered to be pre-Silurian on the basis of radiometric ages from the Andover Granite that intrudes the Marlboro. The Andover has yielded dates of 408 to 450 m.y. (Zartman and Naylor, 1984), with a probable minimum age of 430 $\pm$ 5 m.y. Geochemical data, discussed above, also suggests a pre-Silurian age, but not a late Proterozoic Age. The Fishbrook Gneiss has been radiometrically dated at 730 $\pm$ 26 m.y. (Olszewski, 1980) and probably represents a maximum age for the Nashoba Block units.

Acknowledgements

DiNitto (1983) mapped the northern part of the Marlborough Quadrangle as part of an M.S. thesis study in the Department of Geology & Geophysics, Boston College. He wishes to acknowledge support by the Rev. Daniel Linehan, S.J. Fund. Hepburn's and Hill's work on the Nashoba Block is supported by NSF grants EAR-8212760 and EAR-8212761, respectively.

Road Log

Meet at the Holiday Inn parking lot, Marlborough, Massachusetts promptly at 9:00 a.m. It is located at the intersection of U.S. Route 20 and Interstate 495. To reach the Holiday Inn from Danvers, Massachusetts, take Interstate 95 (State Route 128) south to the Massachusetts Turnpike. Travel west on the turnpike, exiting at I-495 north. Proceed north on I-495 until you reach the Marlborough, Route 20 exit. Take the Route 20, east exit ramp. The Holiday Inn is directly across Route 20 from where you exit I-495.

Mileage	
0	Field excursion starts at the side of the Holiday Inn parking lot. Exit Holiday Inn parking lot. Take a right onto Route 20. Continue to first set of lights.
0.40	Turn left at lights onto Glenn Street.
0.95	Cross Forest Street and continue straight.
2.05	Take left onto Cedar Hill Street. This road becomes Northborough Road at the Town boundary.
2.70	Park on the right shoulder of the road.
	<u>Stop 1.</u> Exposures of the gneiss member are exposed on both sides of the road; the best being on the south side. Seen here are dark gray to black, lighter weathering, medium- to coarse-grained, moderately to highly-foliated amphibolite gneiss. Segregational layering varies from approximately 5 mm to 15 cm. The leucocratic layers are commonly pegmatitic to dioritic in composition, while the darker mafic layers are well-foliated amphibolites. Locally, the amphibolite dominates the exposure without any obvious gneissosity. Return to cars. Turn around and head west on Northboro Road to Cedar Hill Street
4.30	Turn right onto Bartlett Street
4.45	Turn left onto Hayes Memorial Drive. Head north.
4.90	Pull vehicles over to the right hand side of the road, as close to the curb as possible. Proceed to the sourthern most exposure at this stop.
	<u>Stop 2.</u> Sandy Pond Amphibolite Member. Exposures of this member are along the east side of the road. As you walk north along the road from the southern most exposure the amphibolites grade from medium-grained to coarse-grained and then to fine-grained. The medium and fine-grained amphibolites are black, well-foliated as a result of thin (1 mm) plagioclase stringers, and rarely contain thin interlayers of "epicule" (Skehan and Abu-moustafa, 1976). The coarse-grained variety exhibits only a mild foliation, and a mottled texture due to the weathering of plagioclase around hornblende crystals that range in size from 0.5 to 2 cm). This mottled appearance may be a relict texture of a gabbric intrusion. Note the existance of a thin (1 m) fault zone in the coarser amphibolite exposure. A flinty mylonite can be seen in the fault. The amphibolites at this exposure trend approximately N75E and dip 55 NW. Return to cars. Continue north on Hayes Memorial Drive.

5.20 Pull vehicles over to the right hand side of the road.

Stop 3. The rusty schist and granulite member exposed here consists of thinly interlayered, medium-gray, lighter gray weathering, fine- to medium-grained granulite; medium gray, rusty-weathering, fine- to medium-grained schist; and black, poorly foliated, fine- to medium-grained amphibolite. The dominant foliation seen here results from the interlayering and differential weathering of the three rock types. Also, note the conspicuous garnet crystals within the schist. Prior to the covering of portions of this outcrop garnet megacrysts up to 5 cm in size could be observed. Return to cars. Continue north on Hayes Memorial Drive.

5.30 Exposure on both sides of the road are amphibolites of the amphibolite and schist member of the Marlboro Formation.

5.35 Turn right onto Route 20, eastbound.

6.85 Cross over I-495 and continue straight.

8.35 Entering Marlborough Center. Exposure on the left hand side of Route 20, behind the storefront buildings, is Emerson's (1917) type locality for the Marlboro Formation.

8.65 Turn left into parking lot located just past City Hall, which is on the right hand side.

Stop 4. Along the north side of Main Street is the type locality of the Marlboro Formation defined by Emerson (1917). Please view only those rocks exposed at the edge of the parking lot, as local store owners have not given permission to walk behind the buildings. Additionally, the town has asked that no hammers be used because of landscaping activities the town has performed recently. Exposed here are interlayered rusty-weathering fine- to medium-grained, muscovite-biotite schist; medium gray, fine-grained, moderately-foliated granulite; and medium-grained, black to bluish-dark gray amphibolite. The rocks here trend nearly east-west (N80E) with a steep dip (80°N). The exposure also contains numerous small open folds whose axial planes trend nearly east-west and have shallow dips to the north. One thin layer (10 cm) of granulite contains conspicuous porphyroblasts of hornblende that reach 2 cm in length. Return to cars. Take a left when exiting the parking lot and continue east along Route 20.

8.85 Turn left following Route 20, east. Exposures on the left hand side are amphibolites of the Sandy Pond Amphibolite Member of the Marlboro Formation.

9.15 Turn right at the lights following Route 20, east.

9.55 Continue through the lights along Route 20, east.

10.35 Turn left into Fire Station driveway. Please park only along the left hand side of the driveway.

Stop 5. Exposures behind the Fire Station are medium gray to black, medium- to coarse-grained, well-foliated epicule-bearing amphibolites of the Sandy Pond Amphibolite Member. These rocks are typical of the Sandy Pond and contain large conspicuous lenses and pods (up to 60 cm in length) of light green epicule, parallel to the foliation. Return to cars. Exit left from the fire station driveway.

10.45 Lunch stop. Turn left into McDonald's. Exit left from the parking lot, and continue along Route 20, east.

10.90 Turn right onto Farm Road.

11.10 Pass Marlboro Airport on left.

12.05 Pull vehicles off of the right hand side of the road. BE CAREFUL, watch for cars coming around the corner.

Stop 6. Indian Head Hill Granite. The granite here is an equigranular, medium-grained, bluish-medium gray granite that weathers to a ligher gray or a rusty beige color. This exposure is very typical of the Indian Head Hill exposed in the Marlborough area. Return to cars. Continue straight on Farm Road.

12.75 Turn right onto Framingham Road.

13.30 Turn right onto Maple Street and continue north back towards Marlborough Center.

13.90 Turn left following Maple Street. Continue straight on Route 20, west.

15.90 Take first ramp onto I-495 northbound, from Route 20.

16.70 Park on right-hand side of highway on the grass. Make sure your vehicle is off the breakdown lane.

Stop 7. This large roadcut on both sides of the highway belongs to the schist submember of the amphibolite and schist member of the Marlboro Formation, and represents the type locality for the schist submember. Exposed are highly-weathered, rusty-sillimanitic schists that are interlayered with minor amounts of dark gray to black, poor- to moderately-foliated, medium-grained amphibolite, quartzite and light gray, fine-grained granulite. Andover Granite intrusions can be seen throughout the roadcut, but are primarily exposed at the extreme northern end. Return to cars and continue northward on Route I-495.

17.00 Take exit 25A from I-495 to I-290, eastbound.

17.30 Pull vehicles off the exit ramp to the left and park on the grass. Lock your vehicles. Walk back along the exit ramp towards I-495, just south of the exit ramp.

Stop 8. At the southern end of this exposure is the foliated phase of the Andover Granite. The rock here is a light gray to white weathered, porphyritic, medium- to coarse-grained biotite-muscovite-garnet granite. The foliation seen here is produced by the wrapping of fine-grained micas around augens of plagioclase. The texture is cataclastic in origin, and would be classified as a protomylonite.

Proceeding northward, approximately 60 feet along the right side of I-495, towards the exit ramp to I-290, are exposures of the Millham Reservoir member of the Marlboro Formation. The Millham here, consists of a medium bluish-gray, light gray weathering, medium-grained, poorly to moderately-foliated granite. Many stringers and porphyroclasts of quartz and plagioclase define the foliation. These rocks are also considered to be protomylonites on the basis of cataclastic testures.

Continuing northward along the east side of the exit ramp the rocks in the roadcut change to a rusty-weathering sillimanitic schist that is highly weathered in places. The schist forms two rather large lenses within the Millham granulites at this I-495 and I-290 roadcut (Figure 3). These schists are very simlar to those of the schists of the amphibolite and schist member of the Marlboro Formation. Proceeding northeasterly towards the merger of the exit ramp from I-495 with the eastbound lane of I-290 the rocks become very sheared and represent a mylonitic submember of the Millham Reservoir member. These rocks grade (southwest to northeast) from protomylonite to ultramylonite. The rocks vary in textures and colors from medium bluish-grays and porphyroclastic protomylonite to fine-grained darker mylonite to very fine-grained, medium to light whitish-gray ultramylonite. All of the rocks are highly foliated and represent the presumed trace of the Assabett River Fault Zone. Return to cars. Continue on exit ramp towards I-290, eastbound. Follow I-290 to the end.

18.40 Turn right onto Fitchburg Street.

20.10 Continue through the intersection at the lights, crossing Elm Street.

20.30 Turn right onto Lincoln Street.

20.75 Merge with Route 20, west.

21.40 Turn right into Holiday Inn parking lot. End of road log.

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GEOLOGY, PETROLOGY AND ORIGIN OF THE PRECAMBRIAN IGNEOUS ROCKS LOCATED IN THE AREA NORTH OF BOSTON

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The terrane lying east of the Bloody Bluff Fault zone is variously identified as Dedham - Milford Zone (Zen, ed., 1983, Zartman and Naylor, 1984), Avalon Zone (O'Brien et al, 1983). The adopted designation here is after Hepburn, Hon and Hill (in review) who suggested the term Boston Platform in recognition that tis terrane is largely underlain by igneous and metamorphic rocks of pre-Appalacian age (>585 m.y.) with a thin veneer of the late Precambrian to Early Paleozoic sediments (Protorozoic Z to Cambrian). This field trip guide is designed to visit localities in a fault-bounded block north of Boston which show some of the complex relationships among various plutonic and volcanic units emplaced during the maximum igneous activity period (585 - 630 m.y.) and during the period immediately preceeding it.

Located north of Boston on the Boston Platform, this fault - bounded dominantly igneous Precambrian terrane (Fig. 1) abuts along its southern border against the Boston Basin and is seperated from the Boston Basin by the Northern Boundary Fault. The Walden Pond Fault seperates the block from the Salem Gabrodiorite and the Peabody Granite located to the north on the Boston Platform. The dominant rock type of this terrane is the Dedham Granodiorite, which is a fine to coarse grained calc-alkaline plutonic ranging from granodiorite to tonalite to diorite. Lying above as well as intruded by the Dedham are felsic volcanics. Current thinking (Zen, ed., 1983, Kaye, 1980) designates all these felsic volcanics as the Lynn Volcanic Complex with assigned ages ranging from Silurian to Precambrian (Protorozoic Z). Through detailed field mapping, we are able to show that there are actually two different felsic volcanic units. One of these units is in fact the Lynn Volcanic complex proper which through field criteria has been subdivided into four members, three extrusive and one intrusive. The other unit has a diffent structural setting, petrography and geochemistry than the Lynn. This unit is spatially related to the Middlesex Fells Volcanic complex and the Westboro Formation. It is pervasively intruded by the Dedham, is recrystallized and occurs in a rapidly changing sequence with the Middlesex Fells basalts. Because of this relationship, it is apparent that the Middlesex Fells Volcanic complex consists of two members, a previoulsy identified basaltic unit (here called the Middlesex Fells mafic member) and a felsic unit previously identified as the Lynn (here named the Middlesex Fells felsic member). The Middlesex Fells and the Westboro are pervasively intruded by the Dedham and sit essentially as large roof pendents in the Dedham. Through detailed field mapping, petrographic and geochemical studies, we have concluded that the Lynn and the Dedham form a co-magmatic volcano-plutonic complex.

FIELD AND PETROGRAPHIC STUDIES

The Dedham Granodiorite is a Precambrian, 600 - 630 m.y. old, granitic body that underlies most to the Boston Platform. North of Boston it occurs mostly as a granodiorite and diorite consisting of quartz, plagioclase, hornblende and small, varying amounts of microcline. Texturally the Dedham ranges from subvolcanic to coarsly crystalline varieties. The subvolcanic part contains a

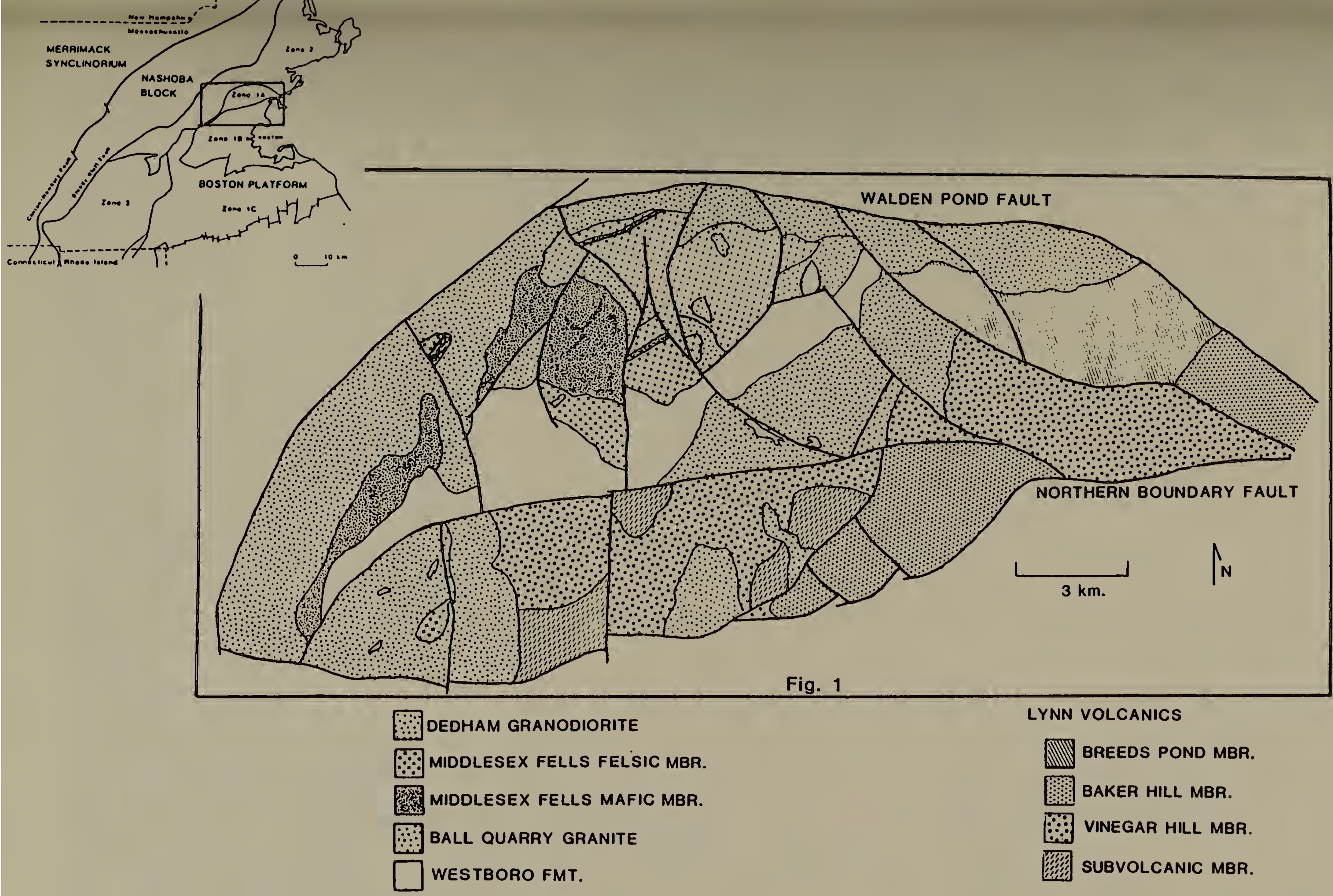

MERRIMACK SYNCLINORIUM
NASHOBA BLOCK
BOSTON PLATFORM
WALDEN POND FAULT
NORTHERN BOUNDARY FAULT
3 km.
N
DEDHAM GRANODIORITE
MIDDLESEX FELLS FELSIC MBR.
MIDDLESEX FELLS MAFIC MBR.
BALL QUARRY GRANITE
WESTBORO FMT.
LYNN VOLCANICS
BREEDS POND MBR.
BAKER HILL MBR.
VINEGAR HILL MBR.
SUBVOLCANIC MBR.

Fig. 1

bimodal distribution of mineral sizes with large plagioclase and hornblende grains in a granophyric groundmass. The coarse crystalline phase consists of equal sized, euhedral grains of plagioclase and hornblende with quartz and microcline forming interstitial grains. In places the Dedham contains xenoliths of Westboro quartzite and basalt probably derived from the Middlesex Fells volcanics.

The Lynn Volcanic complex is composed of dacites and rhyolites that form a thick extrusive and subvolcanic sequence. The phenocryst content of all the Lynn members is made up of plagioclase and hornblende with occasional microcline phenocrysts. The Lynn totally lacks quartz as a phenocryst phase. The phenocrysts lie in a glassy groundmass. The Lynn has been divided, based on field criteria, into four members; Breeds Pond, Baker Hill, Vinegar Hill and subvolcanic members. The Breeds Pond member is a red, gray and black porphyritic ash fall and tuff with sedimentary and volcanic xenoliths. Individual eruption events can be distinguished through an upward fining sequence in each event. This is especially evident in the Lynn Woods where the base of an event is marked by large volcanic xenoliths that range up to 2 meters in size with the size of the xenoliths decreasing upward in the sequence. The Baker Hill is a red porphyritic ash fall and tuff containing only small volcanic rock fragments. This is a very homogeneous member with bedding difficult to determine in the field. Both the Breeds Pond and the Baker Hill show bedding with a preferred alignment of the phenocrysts and glass shards. The most extensive member of the Lynn is the Vinegar Hill member which is a typical lahar and volcanic breccia containg crystal/lithic tuff beds. This member is massive with no bedding evident in the lahars. The many xenoliths in the Vinegar Hill are composed of the Westboro quartzite and other sediments, various felsic and mafic volcanics and granitic fragments which range in size up to 1 meter. These granite xenoliths are not of the Dedham but fragments of an older granite that outcrops in the Ball Quarry in Saugus and here termed the Ball Quarry Granite. The subvolcanic member of the Lynn has a bimodal size distribution of mineral grains. Plagioclase and hornblende phenocrysts are distributed in a fine grained groundmass of quartz and feldspar. The distinction made here between subvolcanic and extrusive is that the subvolcanic rocks have a coarser grained groundmass where the individual mineral grains in the groundmass are visible under the microscope.

The Ball Quarry Granite is a coarse grained plutonic rock containing large subhedral grains of kspar and plagioclase with anhedral quartz grains. It is distinguished from the Dedham because of the kspar and the excess of K_2O over Na_2O in the Ball Quarry.

The Middlesex Fells Volcanic complex as here defined is a bimodal suite of basalt and rhyolite. Both the felsic and mafic members are entirely thermally recrystallized and pervasively intruded by the Dedham. The felsic member contains phenocrysts of plagioclase, quartz and hornblende with inclusions of Westboro quartzite. These volcanics form thin pyroclastic sheets that are spatially related and intercalated with the Westboro quartzite and the mafic member of the Middlesex Fells. The mafic member is an alkali olivine basalt containing phenocrysts of pyroxene, hornblende and plagioclase. Both members of the Middlesex Fells are pervasively intruded by the Dedham.

The Middlesex Fells felsic and the Lynn Volcanics can be distinguished from one another in the field in several ways. The Lynn volcanic units hve a very fine grained glassy groundmass and suffers from no recrystallization. The Middlesex Fells, due to intrusion of the Dedham, is entirely recrystallized. The Lynn and the Dedham have a texturally gradational contact which is evidence for their co-genetic nature. The Middlesex Fells has a very complex relationship with the Dedham. Throughout the terrane, the Dedham is pervasively injected into the Middlesex Fells and in places becomes fine grained near the contacts. The

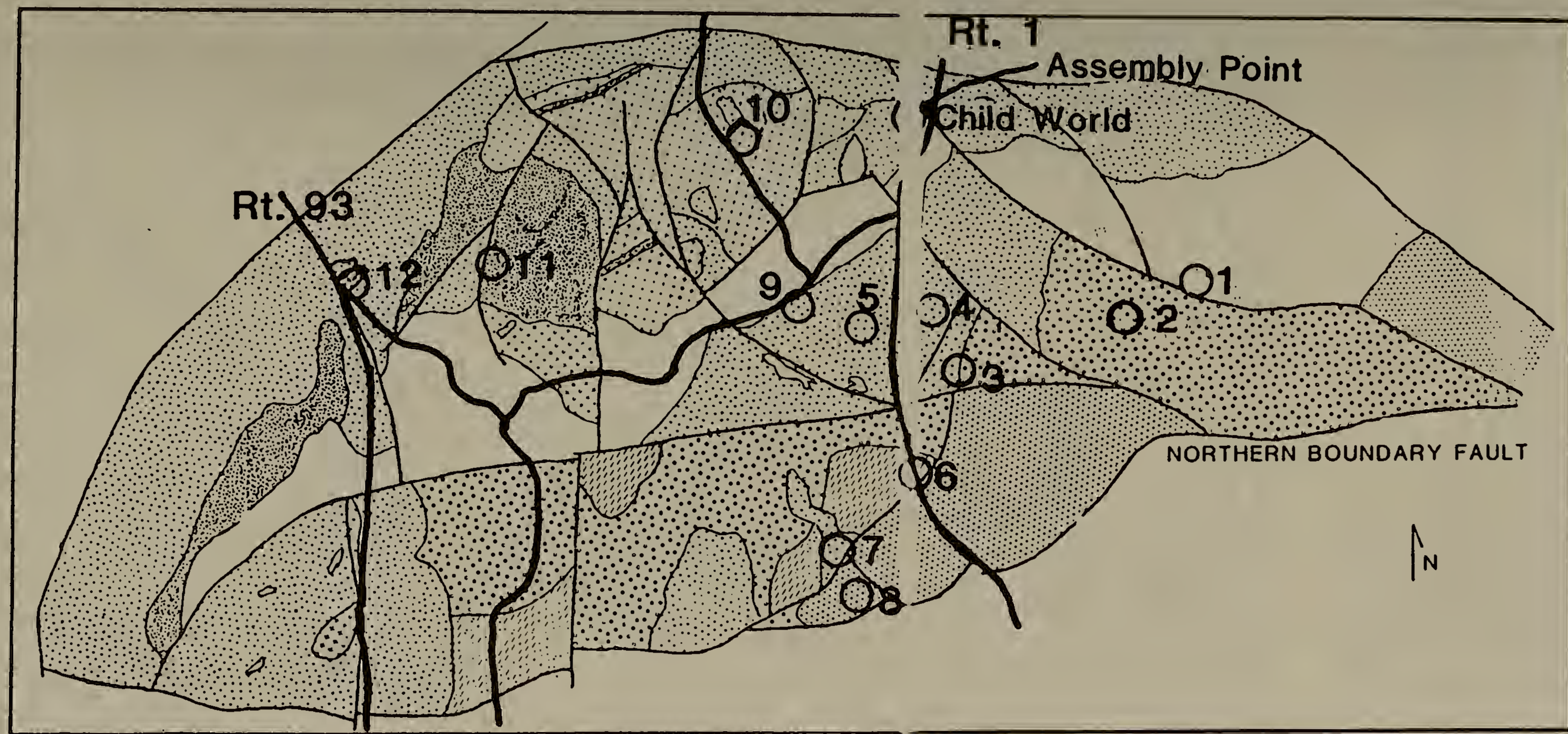


Fig. 2

	STOPS
LYNN VOLCANICS	1,2,3,5,7,8
DEDHAM PLUTONICS	4,5,7,9,10,11,12
BALL QUARRY GRANITE	2,3,5
MIDDLESEX FELLS FELSIC	10,12
MIDDLESEX FELLS MAFIC	10,11,12
WESTBORO FM.	4

Lynn contains no phenocrysts of quartz whereas the Middlesex Fells felsic member has quartz phenocrysts.

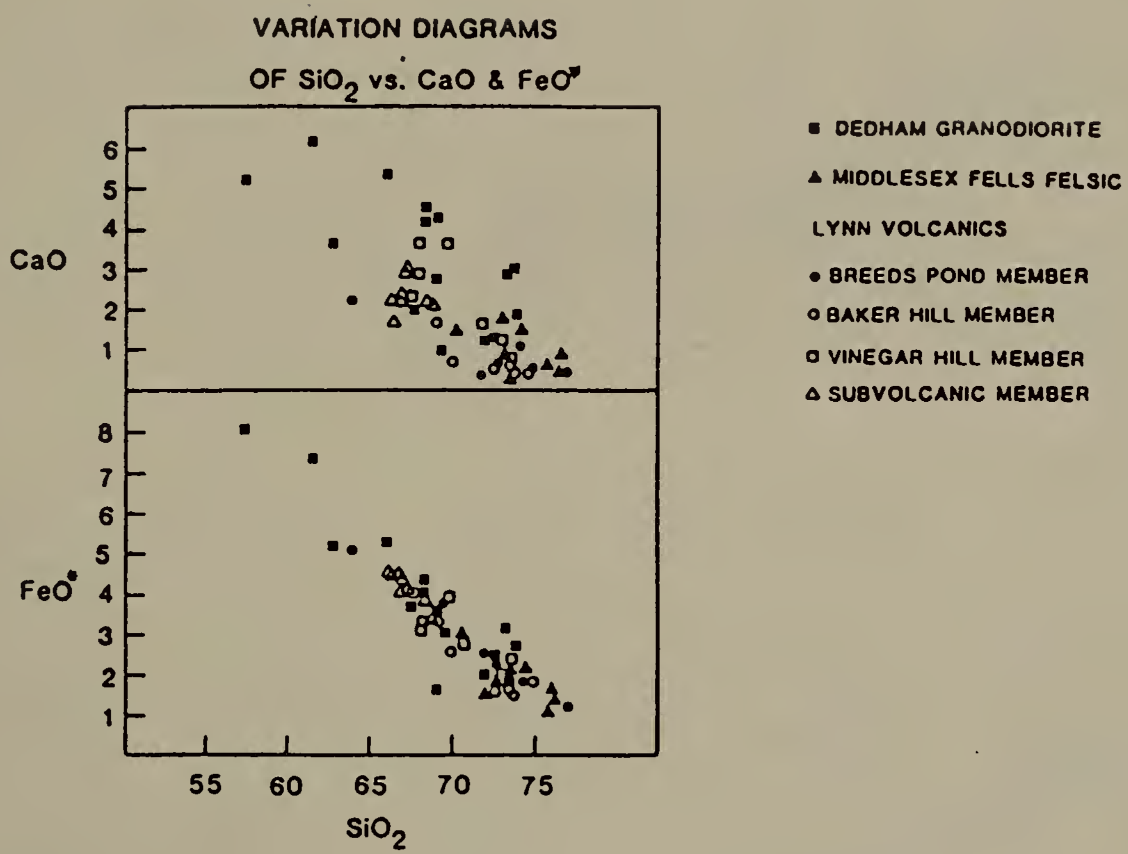
Another difference between the Lynn and the Middlesex Fells felsic member is that the Lynn is co-genetic with the Dedham. In the field this cogenetic nature is displayed through a gradual textural transition of the subvolcanic member between the volcanic textures of the Lynn and the plutonic textures of the Dedham. The major phenocryst phases of the Lynn, plagioclase and hornblede, are also the dominant mineral phases of the Dedham. Some hornblende grains in the Dedham exhibit cumulitic textures. The mineralogy of the Lynn and the Dedham sugest that the Lynn and the Dedham are partly related through the fractionation of plagioclase and hornblende.

GEOCHEMISTRY

Samples of the Lynn and Middlesex Fells Volcanics and Dedham Granodiorite have been analyzed for both major and trace elements as well as Sm-Nd isotopes. Major and trace elements show no major geochemical difference between the Lynn and the Dedham. Both the Lynn and the Dedham have large variations of major element abundences with a complete overlap between the two units. There is also no difference in major element chemistry between the different members of the Lynn, all having a complete overlap. The Dedham contains the least evolved rock types in the area and has a large variation in major elements (eg. SiO_2= 57% - 74% and $Fe_2O_3^*$= 1% - 8%). The Lynn has less of a variation of the major elements than the Dedham, yet there is still a large range (eg. SiO_2= 66% - 77% and $Fe_2O_3^*$= 1.5% - 5%). The subvolcanic unit has a very small range of major elements and as a group are the least evolved of the Lynn members. The overlap and trend of major element compositions can best be seen on variation diagrams of SiO_2 vs. CaO and $Fe_2O_3^*$, as well as on an AFM diagram (Fig. 3). On an AFM diagram, the Dedham and Lynn plot on a calc-alkaline fractionation trend where the Dedham has the most evolved of the rock types. This major element similarity lends further support to the co-genetic nature of the Dedham and the Lynn. The Middlesex Fells felsic member has a very small range of major element concentrations (eg. SiO_2= 73% - 76% and $Fe_2O_3^*$ = 1% - 2%). This tight grouping of major element concentrations as opposed to the large range of the Lynn is another method of distinguishing between the Lynn and the Middlesex Fells felsic.

Consistant with the major elements are the trace elements. The Dedham and the Lynn have large variations in trace elements abundences, again with a complete overlap between the two groups. The Dedham has a large variation in trace elements (eg. Sc= 3 - 22 ppm) and the samples that are the least evolved in major elements are also the least evolved in the trace elements. The Lynn again shows a smaller spread in trace element abudences than the Dedham (Sc= 2.7 - 12 ppm). The subvolcanic Lynn has a very small spread in the trace elements abundences (Sc= 9 -12 ppm). The Middlesex Fells felsic has an even smaller spread (Sc= 4 - 6 ppm). The rare earth elements (REE) exhibit a range of compositions for all the rock types (Fig. 4). The spread of REE compositions in the Dedham is identical to the spread in the Lynn. The Dedham has lower values of the light REE (La-Sm) suggesting that it is the least evolved of all the rock types. The high heavy REE (Tb-Lu) contents (10 - 20 times chondrite) point to a lack of garnet in the residue suggestting a shallow level of partial fusion. From the trace elements, especially the REE, the fractionation trend and the evolutionary sequence of the Dedham Granodiorite and the Lynn Volcanics can be deduced. The subvolcanic member is the least evolved member, followed by, in order of least to most evolved, Vinegar Hill, Breeds Pond and Baker Hill members. The overlap of trace element abundences provides further support to the co-genetic nature of the Dedham and the Lynn.

As stated before, the Dedham and the Lynn may be related through the


VARIATION DIAGRAMS
OF SiO2 vs. CaO & FeO*
CaO
FeO*
SiO2
55
60
65
70
75
DEDHAM GRANODIORITE
MIDDLESEX FELLS FELSIC
LYNN VOLCANICS
BREEDS POND MEMBER
BAKER HILL MEMBER
VINEGAR HILL MEMBER
SUBVOLCANIC MEMBER

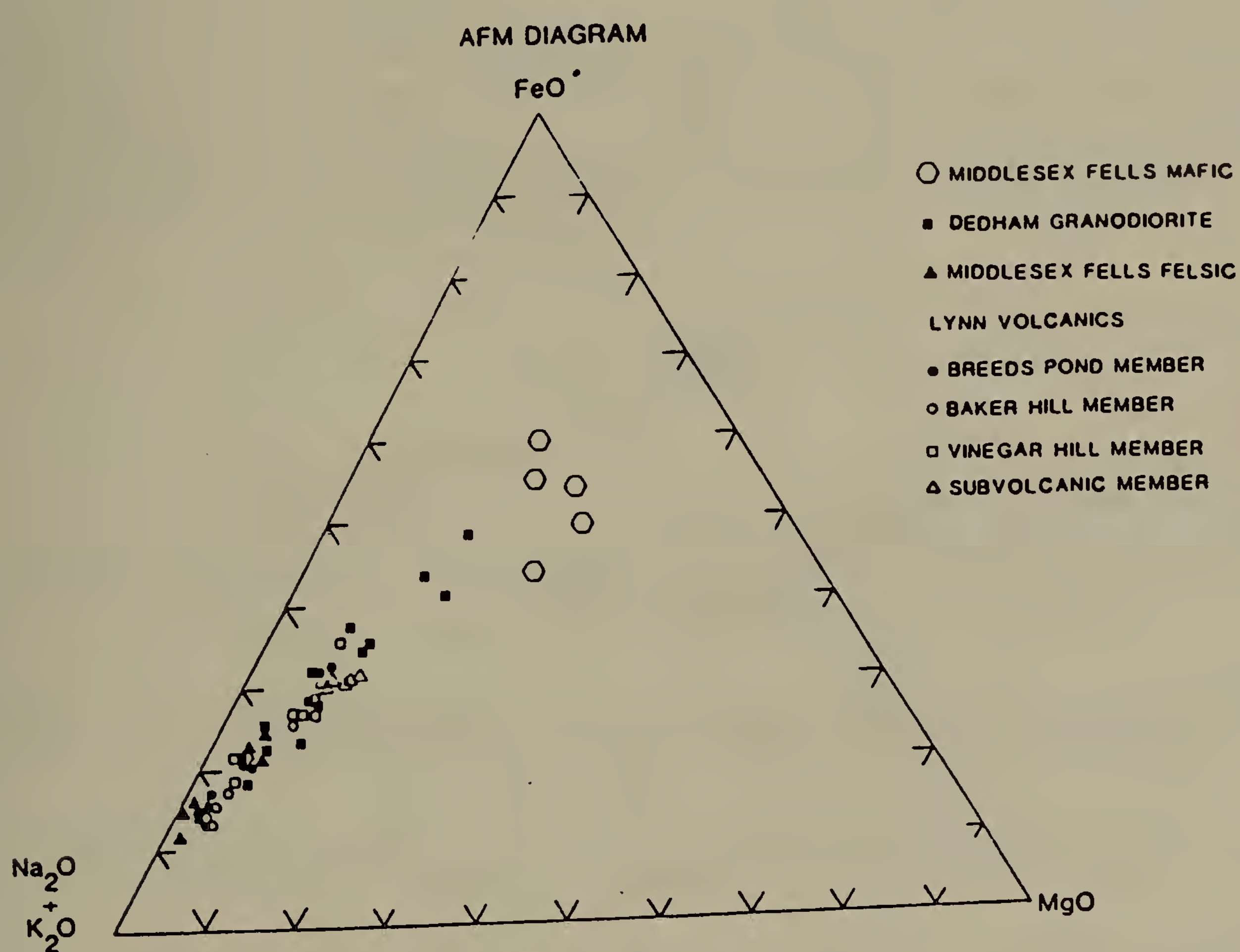

AFM DIAGRAM
FeO*
Na2O + K2O
MgO
MIDDLESEX FELLS MAFIC
DEDHAM GRANODIORITE
MIDDLESEX FELLS FELSIC
LYNN VOLCANICS
BREEDS POND MEMBER
BAKER HILL MEMBER
VINEGAR HILL MEMBER
SUBVOLCANIC MEMBER

Fig. 3

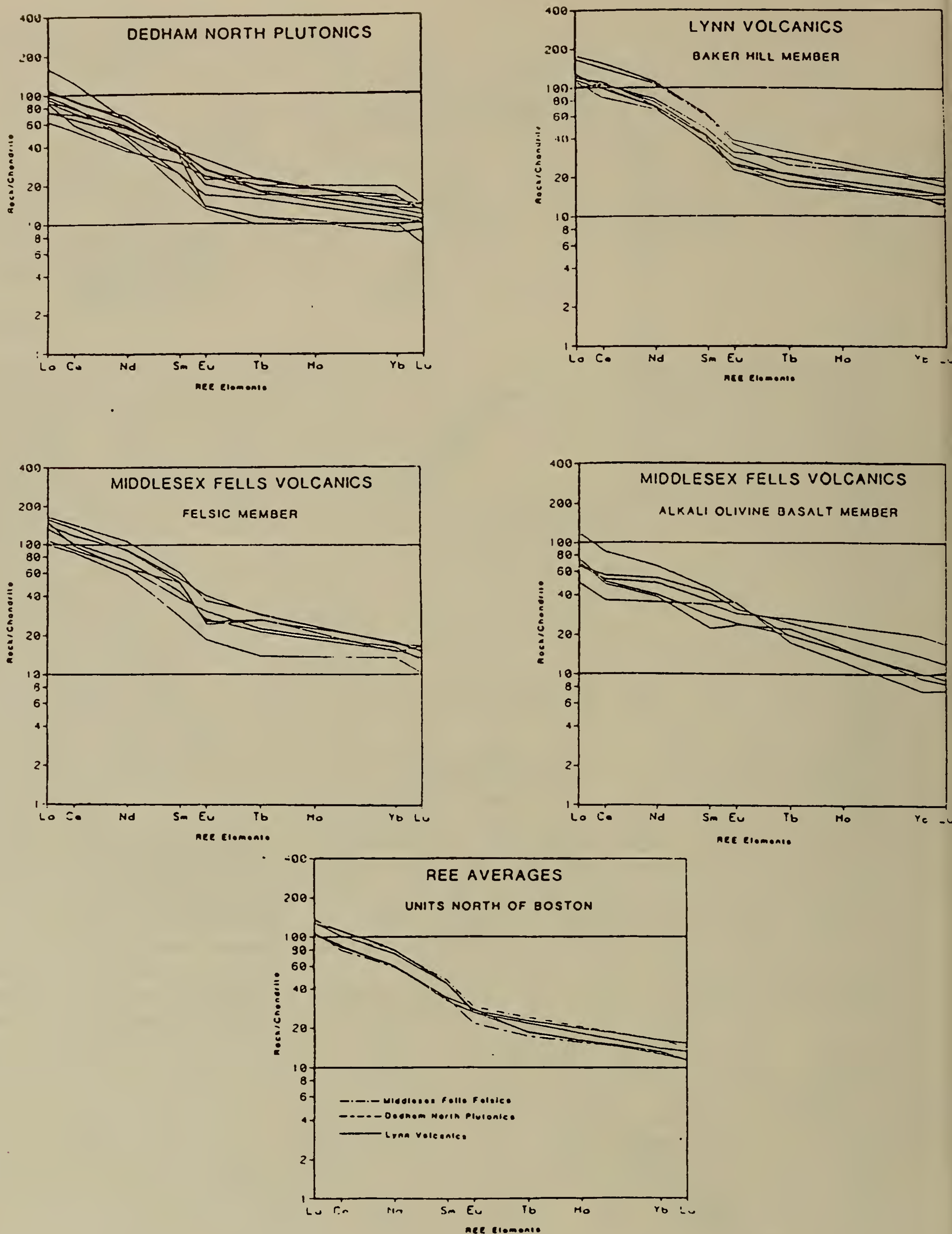

DEDHAM NORTH PLUTONICS
LYNN VOLCANICS
BAKER HILL MEMBER
MIDDLESEX FELLS VOLCANICS
FELSIC MEMBER
MIDDLESEX FELLS VOLCANICS
ALKALI OLIVINE BASALT MEMBER
REE AVERAGES
UNITS NORTH OF BOSTON
Middlesex Fells Felsics
Dedham North Plutonics
Lynn Volcanics
Rock/Chondrite
REE Elements
La Ce Nd Sm Eu Tb Ho Yb Lu

Fig. 4

fractionation of plagioclase and hornblende. Smith et al (1984) applied crystal fractionation models to explain the chemical differences between the REE of the various members of the Lynn and the Dedham (Fig. 5). Using the Dedham as the most primitive rock, the subvolcanic member of the Lynn can be derived. The rest of the Lynn members can be modelled using the subvolcanic member as the parent by fractionation of varying amounts of plagioclase and hornblende phenocrysts.

The Sm-Nd data (M. Hill, personal communication) suggets that the Lynn and Dedham were isotopically heterogeneous, inplying heterogeneous sources. This data is interpreted as two processes that worked in conjunction to form the Dedham/Lynn magma (Fig. 6). Variable mixing of two shallow sources, basaltic and continental crust sources followed by fractional crystallization of hornblende and plagioclase to form the Dedham and the various Lynn members.

SUMMARY

From the field, petrographic and geochemical evidence, it is concluded here that the Dedham and the Lynn are the intrusive and the extrusive equivalents of the same parent magma. The two are related partially through magma mixing but mostly by a fractional crystallization process through the fractionation of hornblende and plagioclase. Thus the Lynn must be of a similar Late Precambrian age as the Dedham. The Middlesex Fells Volcanics are here divided into two members, a mafic member and a felsic member. The felsic member had been previously undefined because of its inclusion in the Lynn Volcanics. The field and geochemical evidence conclusively show that there are two felsic volcanics in the area. The Middlesex Fells forms an earlies volcanic pulse in the magmatic episode that formed the Dedham and the Lynn.

This field trip is designed to visit localities which show relationships dealing with one of the following points (Fig. 2):

1. Textural variations within the Lynn Volcanic complex are evidence for the gradational textural changes from the Lynn into the Dedham granodiorite
2. The Middlesex Fells consists of two members, felsic and mafic
3. Evidence for the pre-Dedham granitic terrane.

It is adviseable to have copies of the Lynn and the Dedham North U.S.G.S 7 1/2 minute quadrangles as a reference. Funding for this work was provided through NSF grants EAR-8212760 awarded to J. C. Hepburn and R. Hon and EAR-8212761 awarded to M. Hill.

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aye, C.A., 1980, Bedrock geologic maps of the Boston North, Boston South, and Newton Quadrangles, Massachusettes; U.S.G.S. Misc. Field Studies Map MF-1241.

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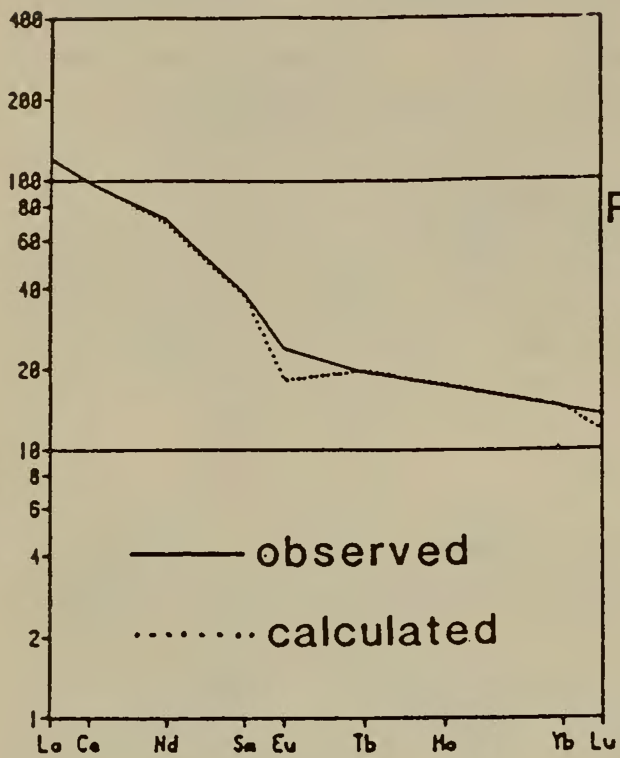
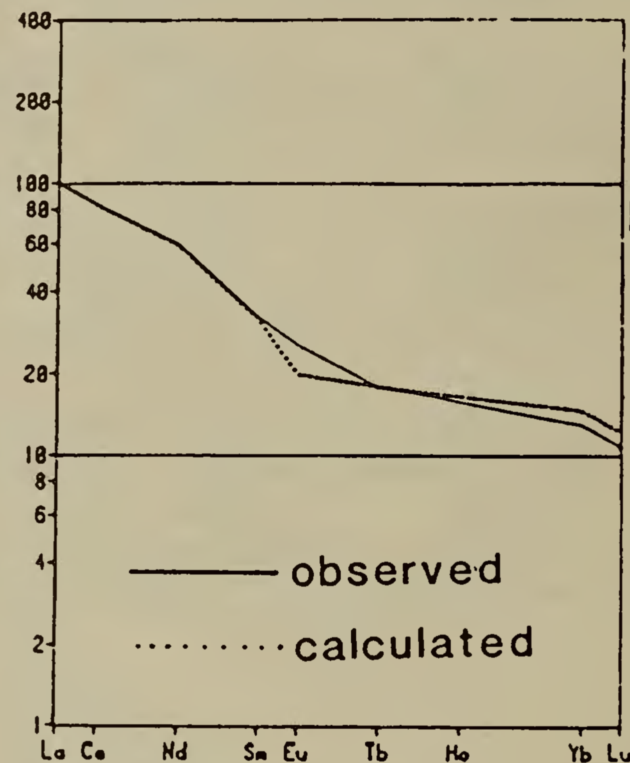
mith, C., Hon, R. and Hill, M., 1984, Precambrian cogenetic volcano-plutonic association in the Avalonian terrane north of Boston, eastern Massachusetts; Geol. Soc. Amer. Abs. w. Programs, v. 16, p. 64.

MODEL REE DIAGRAM
FOR
BAKER HILL MEMBER



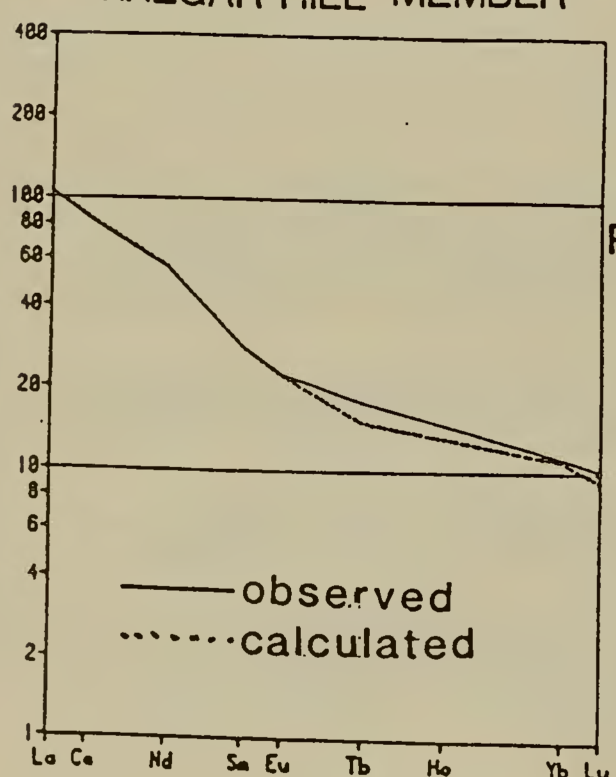
PARENT:
AVG. SUBVOLCANIC LYNN
24% CRYSTAL REMOVAL
85% PLAGIOCLASE
15% HORNBLENDE

MODEL REE DIAGRAM
FOR
SUBVOLCANIC MEMBER



PARENT:
AVG. DEDHAM NORTH
22% CRYSTAL REMOV
75% PLAGIOCLASE
25% HORNBLENDE

MODEL REE DIAGRAM
FOR
VINEGAR HILL MEMBER



PARENT:
AVG. SUBVOLCANIC LYNN
6% CRYSTAL REMOVAL
25% PLAGIOCLASE
75% HORNBLENDE

Fig. 5

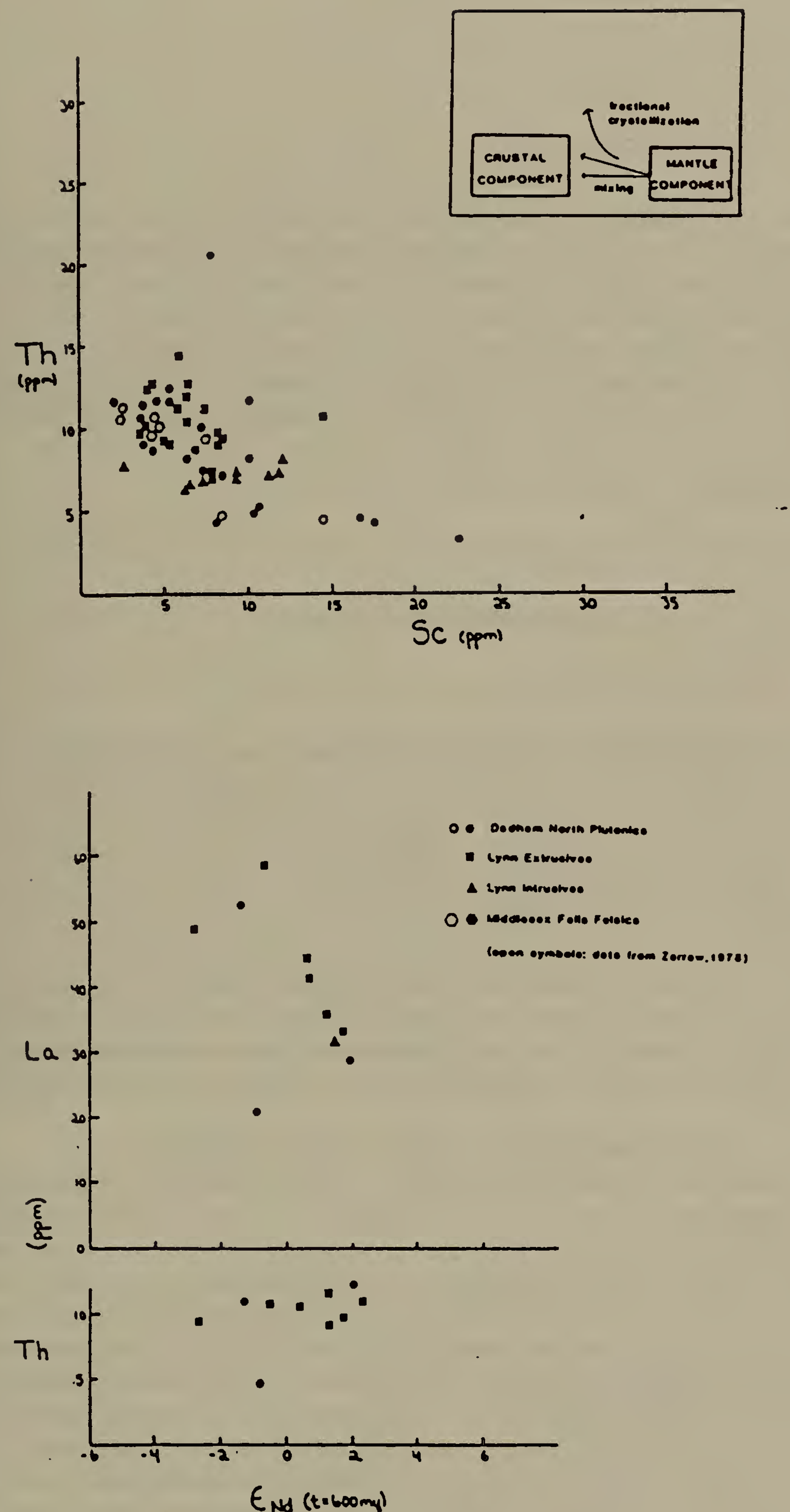

CRUSTAL COMPONENT
MANTLE COMPONENT
mixing
Th (ppm)
Sc (ppm)
La
(ppm)
Th
ε Nd (t=600my)

Figure 6.

Zartman, R.E. and Naylor, R.S., 1984, Structural implications of some radiometric ages of igneous rocks in southeastern New England; Bull. Geol. Soc. Amer., v. 95, p. 522-539.

Zen, E-an, ed., 1983a, Bedrock Geologic Map of Massachusetts; U.S.G.S., scale 1/250,000.

The starting point for the trip is in the parking lot next to Child World in the Saugus Plaza off of Rt. 1 south. The Saugus Plaza is 6.5 miles south from the Inn at Danvers and approximately the same distance north from the Tobin Bridge in Boston. The Plaza is also accessed from Walnut St., Rt. 129. Please note that many of the stops are on private properties. We would appreciate courtesy in respecting the owner's privilages and rights by securing permission before viewing the outcrops; also no hammers please in these places. Thank you.

Assemble at Child World at 8:30.

BEGIN ROAD LOG

0.0	0.0	Exit the Saugus Plaza and turn right onto Walnut St. and proceed over Rt. 1.
0.6	0.6	Keep left at the traffic lights.
0.4	1.0	Birch Pond on the left.
2.5	1.5	Park on the left (north) side of the road in the designated parking spaces. From the parking lot, walk straight across the ballfields toward the spillway for Breeds Pond. Continue northwest along the shore for 750 feet. Meet at an abandoned quarry.

STOP 1. LYNN VOLCANICS: BREEDS POND MEMBER
The abandoned quarry displays a rapid change in a variety of rock types with irregular contacts and possible dome-like intrusions. There are four rock types here; welded tuffs, crystal tuffs, breccias and intrusive dome facies. The volcanics are typical of the Breeds Pond with phenocrysts composed of plagioclase and hornblende with the plagioclase dominant. The matrix is glassy and exhibits flow textures. The inclusions in the breccia are composed of fragments of other members of the Lynn. The dome facies here intrudes the volcanics in many places with a very irregular contact between the two facies. This intrusive style is common in the Lynn terrane where domes of Lynn magma intrude into its own volcanic pile. Proceed back to the park and examine the exposure just beside the spillway. Note: there are no granitic inclusions found in this member.

Turn around and proceed west on Walnut St.

0.3 2.8 At the blinking traffic light turn left onto O'Callahan Way.

0.3 3.1 Turn right into the King's Lynne apartments on King's Hill Dr. and proceed up the hill.

0.6 3.7 At the top of the hill, park on either side in parking areas. Walk to the outcrops on the east side of the road.

STOP 2. LYNN VOLCANICS: VINEGAR HILL MEMBER
Please, no hammers at this stop!
This exposure is of a boulder breccia that is massive with no observable structure. The clasts in this rock weather in relief and are composed of other volcanics, Westboro quartzite, granite (Ball Quarry Granite) and basalts (Middlesex Fells mafic). Note at the top of the outcrop, an inclusion of basalt that had, prior to eruption, intruded the Ball Quarry Granite. Chemically the Vinegar Hill is the most primitive unit of the Lynn. The volcanic vent must be close to this exposure and the Vinegar Hill covered the preexisting granitic terrane.

Proceed north (same direction) on King's Hill Dr.

0.2 3.9 Turn left on Sherman St. and proceed to the end of the street.

0.1 4.0 Turn left on Fairmount St. and proceed south.

0.2 4.2 Turn right into Forest Highlands development. Turn immediately right and drive to the end of the street.
0.1 4.3 Park and walk to the freshly blasted outcrops.

STOP 2A. LYNN VOLCANICS: VINEGAR HILL MEMBER
These outcrops are exactlly identical to stop 2 but contain a greater amount of granite xenoliths. Note: these outcrops are temporary and possibly may not be accessible in the future.

0.3 4.6 Leave Forest Highlands and turn right on Fairmount St. and continue to the end of the street.

0.5 5.1 Turn right (at Pvt. J. W. Pace Square) onto Pace Rd. and proceed west.

0.2 5.3 Join (toward right) onto Hamilton St. and proceed west.

0.6 5.9 Cross the Saugus River.

0.3 6.2 Continue straight through Saugus Center on Main St.

0.5 6.7 At the traffic lights, turn left onto Vine St.

0.1 6.8 Turn left onto Talbot St. and continue to the end of the road.

0.3 7.1 Park at the Waybright School. Walk across the ballfield toward the railroad tracks. Walk south along the railroad tracks to outcrops along the tracks.

STOP 3. SUBVOLCANIC LYNN AND BALL QUARRY GRANITE
Along the railroad tracks is the subvolcanic facies of the Lynn. This is a shallow intrusion of Lynn magma containing phenocrysts of plagioclase and hornblede. The matrix is slightly coarser grained than the volcanic units. Walk around the north wall toward the abandoned quarry. Exposed along the north wall is a largely brecciated Ball Quarry Granite (BQG) cemented by matrix identical to the Vinegar Hill. Note: The BQG is also the one which is identified in the numerous inclusions in the Vinegar Hill. As you proceed from the railroad tracks, the brecciated facies is gradually replaced by massive BQG. This granite is coarse grained having large euhedral kspar and plagioclase grains. This granite is the only rock in this terrane with K_2O in excess of Na_2O. We interpret this quarry as a volcanic vent in which the Vinegar Hill extruded and covered an older granitic terrane of the BQG. Return to the parking lot by proceeding over the walls of the quarry. Note again the increasing occurance of the brecciated facies.

Return to Main St.

0.3 7.4 At the traffic lights, turn right onto Main St.

0.1 7.5 Turn left into Saugus Common.

0.2 7.7 Park here on the right.

STOP 4. DEDHAM NORTH GRANODIORITE AND INCLUSIONS OF WESTBORO FORMATION
Outcrops in the middle of the loop exposes Dedham containing numerous inclusions of the Westboro Formation. This is an example of the typical granodiorite phase of the Dedham with coarse euhedral plagioclase and hornblende grains surrounded by quartz and minor kspar. These outcrops show how the Westboro sits in the Dedham as a roof pendant.

Continue on the road and complete the loop and return to Main St.

0.5 8.2 Turn right on Main St. and proceed west.

0.4 8.6 Cross over Rt. 1 and continue on Main St. for approximately 1000 feet.

0.2 8.8 Turn left into entrance of K-Mart Shopping area and continue to the rear of K-Mart.

0.1 8.9 Park here at outcrops behind K-Mart and Stop & Shop.

STOP 5. DEDHAM NORTH GRANODIORITE
Typical outcrops of the Dedham. This is a coarse grained rock containing hornblende, biotite, plagioclase, quartz and kspar. The Dedham is a typical calc-alkaline granodiorite with normative diopside up to 8%. There are two rhyolite dikes that intrude through the Dedham. Note the contorted contacts between

the dikes and the Dedham. These "soft" contacts imply that the Dedham was not wholly crystalline but partly plastic when the dikes intruded. On the south side of the exposure in the middle is a 2 - 3 feet large granitic inclusion mineralogically and chemically identical to the BQG, which serves as additional evidence that the BQG is a pre-Dedham granite. Note the numerous mafic dikes of unknown age that intrude the Dedham.

Return to Rt. 1 south.

0.4 9.3 Turn right on Rt. 1 and proceed south.

0.6 9.9 Bear right onto Rt. 99 south.

0.3 10.2 Turn left into parking lot of Vogue restaurant. Walk north approximately 300 feet along Rt. 99 to exposures on southeast side of the road.

STOP 6. SUBVOLCANIC LYNN MEMBER
A typical red-green outcrop of the subvolcanic member of the Lynn. This outcrop is very homogeneous containing phenocrysts of plagioclase and hornblende in a ground mass of quartz and feldspar in which the individual mineral grains are visible in thin section. This unit is geochemically very homogeneous with a very limited range in both major and trace element compositions. This unit acted as the shallow magma chamber from which the other members fractionated and which also grades texturally into the Dedham.

Return to Rt. 99 south by turning left out of the parking lot.

0.8 11.0 At the traffic light, turn right onto Elwell St.

0.1 11.1 Turn left into the Malden Moose and park by the large exposure behind the building.

STOP 7. SUBVOLCANIC DEDHAM WITH RHYOLITIC DIKES
Here the fine grained facies of the Dedham is intruded by rhyolitic dikes of the same age. The subvolcanic facies of the Dedham contains phenocrysts of hornblende and plagioclase in a granophyric groundmass of hornblende, plagioclase, biotite, kspar and quartz. The rhyolitic dikes are of the Lynn and are of a similar age as the granite due to the "soft" contacts between the dikes and the granite, similar to the dikes at stop 5. Here the dikes show flow lineations and contain phenocrysts of plagioclase. This was a very shallow area quite near a volcanic opening. The dikes may have acted as feeder dikes to a volcanic vent.

0.2 11.3 Return to Elwell St. and proceed straight across Rt. 99 onto Central Ave. Continue up the hill.

0.2 11.5 Turn left on Kennedy Dr. across from the Granada Apartments entrance.

0.1 11.6 Park on the left side at the red roadcuts.

STOP 8. LYNN VOLCANICS: BAKER HILL MEMBER
These exposures of the Baker Hill are homogeneous and massive red and green volcanics. This unit is an ash fall and welded tuff that contains plagioclase and minor hornblende phenocrysts, lacking or containing very small xenoliths. Note the dike at the end of the northern exposure which contains an inclusion of gray Lynn. This member contains some of the most evolved (least primitive) units of all the members.

0.1 11.7 Turn left at the stop sign onto Salem St.

0.3 12.0 Turn left at the stop sign. Go under Rt. 1 and turn left onto the entrance ramp for Rt. 1 north. Proceed north on Rt. 1.

2.3 14.3 Exit Rt. 1 immediatly after the overpass onto Main St. Proceed west on Main St.

0.9 15.2 At the traffic lights and the Village Park Shopping Plaza, turn left onto the Lynn Fells Parkway and proceed west.

0.3 15.5 Turn left into the Sheffield Heights Condominiums.

0.2 15.7 Park on the right by the roadcut.

STOP 9. DEDHAM NORTH GRANODIORITE
Here exposed is the quartz dioritic facies of the Dedham. This rock is slightly darker than the exposures at stop 5, somewhat finer grained and more homogeneous. The plagioclase and hornblende grains are euhedral and make up greater than 70% of the rock. Many of the hornblende grains exhibit cumulitic textures. Sphene is a very common phase and occurs as euhedral prisms that are evident in hand sample. The Sheffield area contains some of the most primitive rocks in the area with low SiO_2 (61% - 66%), high FeO^* (5% - 7%) and high Sc (10.5 - 17 ppm) compositions. The rocks in this area were some of the first to form, being where some fractionation occured causing cumulate and "quenched" textures.

Exit Sheffield condominiums by turning right onto the Lynn Fells Parkway.

0.4 16.1 Turn left on Main St. at the traffic lights and Village Park Shopping Plaza and proceed north on Main St.

0.9 17.0 Pull off the right side of the road by the exposures on the northeast.

STOP 10A. MIDDLESEX FELLS BASALT AND DEDHAM NORTH GRANODIORITE
At the south end of the outcrop is a fine grained dioritic phase of the Dedham. Further north on the outcrop, this Dedham intrudes into the Middlesex Fells basalt. This basalt has been thermally metamorphosed and recrystallized. Intercalated with this basalt is a small bed of the Middlesex Fells felsic member. The Dedham is quenched, fine grained and dark containing plag-

ioclase and hornblende grains and small pods of coarser material. The basalts were originally fine grained containing plagioclase phenocrysts. It has since been recrystallized due to the injection of the Dedham. The small felsic bed is very fine grained and also recrystallized.

Either walk or drive north along Main St.

0.3 17.3 Stop at the outcrops along the right of the road.

STOP 10B. MIDDLESEX FELLS FELSIC MEMBER.
Outcropping here is a light, somewhat pinkish, recrystallized, porphyritic rhyolite that is intruded by a dioritic dike possibly related to the Dedham magma source. Time permitting, a walk up Castle Hill is possible to further view the variety of types of the felsic member. This is a good exposure to contrast the Lynn and the felsic Middlesex Fells. The rocks here have a lower phenocryst content than the Lynn although the visible phenocryst phases are the same, plagioclase and hornblende. While the Lynn has a very fine grained glassy grounmass, the rock here is totally recrystallized. The different structural setting between the two volcanics is also evident where the Dedham is injected into the Middlesex Fells. Chemically, the Middlesex Fells felsic member has overall higher SiO_2 (74% - 76%) than the Lynn (67% - 75%) and a much narrower range of composi tions. These series of outcrops show the rapidly changing nature of the Middlesex Fells where the basalts and rhyolites are intercalated and are closely related to the Westboro quartzite just south along Main St. This whole sequence is then pervasively intruded by the Dedham. (Good exposures showing the pervasive intrusive character of the Dedham can be seen under the power line about one half mile from the road in both directions.)

Proceed north on Main St. Upon entering Wakefield, Main St. changes to Farm St.

1.1 18.4 Turn left onto Water St. and continue west to Wakefield Center.

0.9 19.3 Railroad crossing

0.2 19.5 Turn left onto Main St. (in Wakefield) and proceed south on Main St.

1.4 20.9 Continue on Main St. through the traffic lights.

1.1 22.0 Turn left at the traffic light onto Franklin St. (in Melrose) and proceed northwest on Franklin St.

0.4 22.4 Railroad crossing.

0.3 22.7 Note small outcrop of the Middlesex Fells felsic to the right. This is a typical, rapidly changing sequence of different units. The rock types here are similar to the types at stop 10B.

0.9 23.6 Turn left on Franklin Place into Stoneham Jr. High.

0.2 23.8 Proceed and park at the tennis courts and walk to the outcrops behind the tennis courts.

STOP 11. MIDDLESEX FELLS BASALT
At the extreme south end of the outcrop, the dioritic facies of the Dedham is intruding the Middlesex Fells basalt. Further behind the tennis courts, there are aplitic dikes intruding the basalts. The basalts are thermally recrystallized due to the Dedham, Strike is N 30° E which is evident from the rapid succession of differing basalt flows. In the northeast corner, a porphyritic rhyolite dike or volcanic of the Middlesex Fells felsic member is exposed in the basalt. On the north face of the outcrop, there are structures that are interpreted to be pillows. The lack of full development of the pillows is due to the low viscosity of alkaline olivine basalts. Observed pillow structures are rare in the Middlesex Fells. The basalts are nepheline normative alkali olivine basalts that sit as roof pendants in the Dedham.

0.2 24.0 Return to Franklin St. and turn left. Proceed left.

0.7 24.7 At the traffic light, turn left onto Rt. 28. Proceed south.

0.4 25.1 At the traffic light, turn right onto Marble St. and go west .

0.3 25.4 At the lights, turn left onto Park St. and go south.

0.2 25.6 Turn left into Mosely Park Townhouse Condominiums and park. Assemble at southern outcrops along Park St.

STOP 12. MIDDLESEX FELLS VOCANICS AND DEDHAM GRANODIORITE
Walk south along Park St. to Orchard St. Walk north examining outcrops of Dedham North diorite - granodiorite. These are sheared medium grained grading northward to fine grained. The Dedham then intrudes the mafic Middlesex Fells. These basalts are recrystallized to greenschist facies having the original structures totally obliterated due to intrusion of the Dedham. Carefully cross Park St. and walk up the northern entrance ramp of Rt. 93 to the outcrops on the right. Here the Middlesex Fells felsic member is also intruded by the Dedham. These exposures nicely exhibit the the rapidly changing sequence in the Middlesex Fells and the intrusive nature of the Dedham. New foundation construction exposed a marble lens 15 feet wide in the Middlesex Fells mafic member. We attach a particular importance to the lens since the observed lithologies make it comparable to the Blackstone Series in Rhode Island or possibly the Greenhead Group in New Brunswick.

END OF TRIP

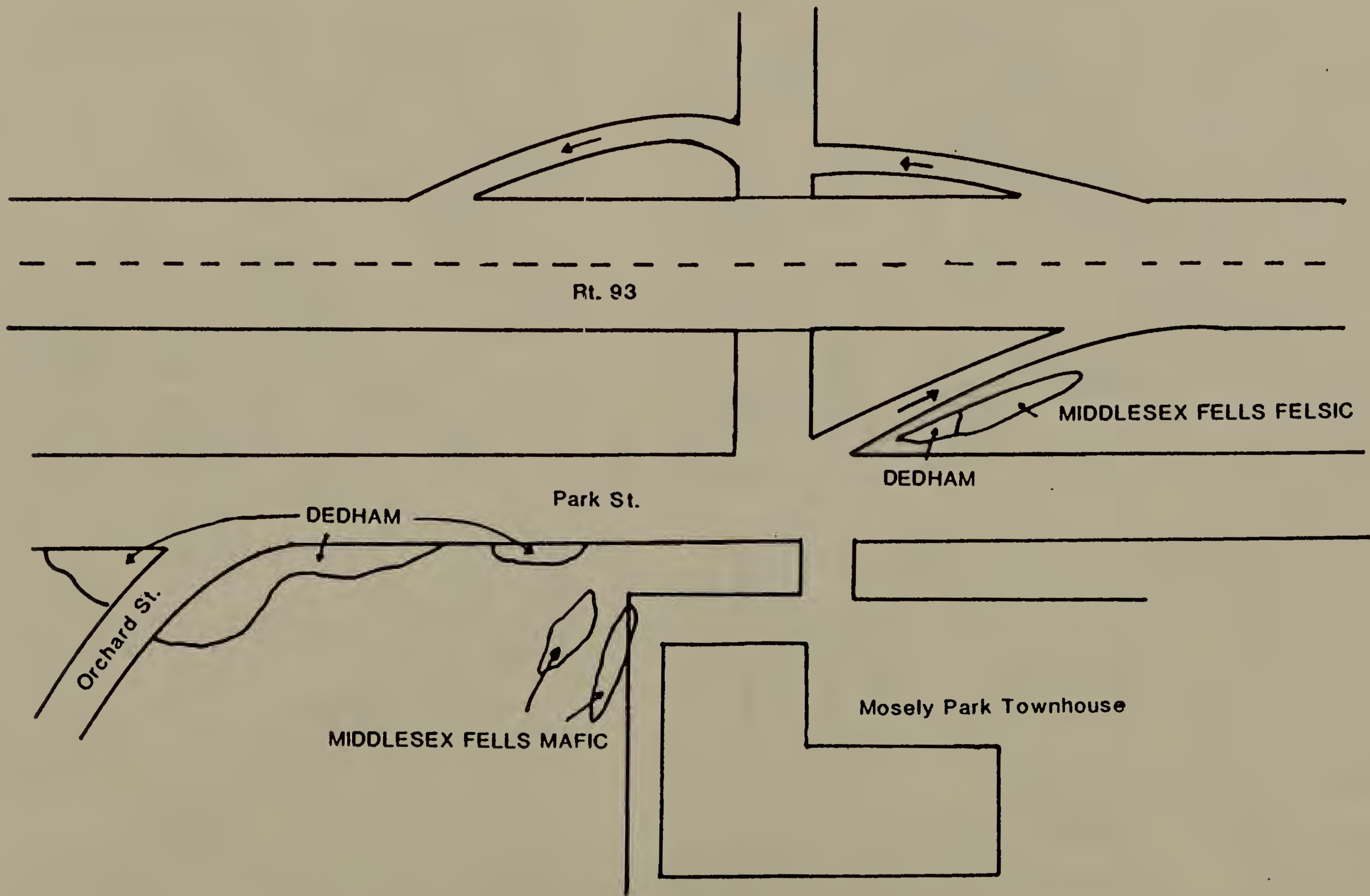


Fig. 7

Map of STOP 12

THE BLOODY BLUFF FAULT SYSTEM

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35 Potter St., Concord, MA 01742

Introduction

The Bloody Bluff fault system is part of one of the most important structural zones in New England and has had a very long and varied history of movement. The delineation of the fault system and recognition of its importance has gradually developed over the past 20 years. The mapping of it was primarily done as part of detailed 1:24,000 scale quadrangle studies. The principle geologists involved and references to a few summary reports are Shride (1976a and b) and Dennen (1981) in the northeast portion, Cupples (1764) and Bell and Alvord (1976) in the central portion and Barosh (1977 and 1981) in the southwest. The great significance of fault emerged in the early 1970's as regional stratigraphic studies indicated the magnitude of the change across the fault (Bell and Alvord, 1976, Barosh and others, 1977), interpretation of aeromagnetic data indicated its prominance and continuity (Alvord and others, 1976; Barosh and others, 1974) and it was demonstrated to correlate with the Lake Char fault in eastern Connecticut (Barosh 1977). New features continue to be found, such as the recent discovery of Late Triassic-Early Jurassic fossils in a fault sliver along it (Kaye 1983), that confirmed an earlier supposition (Oldale, 1962), and many important problems remain to be solved.

General Description

The Bloody Bluff fault system, along with the Lake Char and Honey Hill fault zones in Connecticut, form the structural boundary separating the Nashoba Thrust Belt from the Southeast New England Platform (Fig. 1 and Fig. 1 in Introduction). No pre-Pennsylvanian stratigraphic units are known to correlate across this boundary and it probably represents part of a zone of continental plate collision (see Introduction to guidebook). The fault system is a composite fault zone consisting of several large faults and is herein collectively referred to as the Bloody Bluff fault system. The fault system is irregular in pattern with numerous splays extending southwest from the zone into the Southeast New England Platform. The faults generally dip steeply to the northwest, but may flatten at depth. Movement along the fault system was apparently right-lateral with west over east during the pre-Mesozoic and probably left-lateral and normal during the Mesozoic. The lateral movement apparently created pull-apart basins both within and along the zone at different times (Fig. in Introduction). The total pre-Mesozoic movement may measure in the 100s of kilometers. The Mesozoic left-lateral movement can be deduced from fault patterns (Ballard and Uchupi, 1975), but appears relatively minor. Earthquake studies do not indicate any general movement along the system in the past 300 years. It is probably cut locally by younger small northwest-trending faults.

The relations of Precambrian to Mesozoic rock at various places along the fault system demonstrate a long history of repeated movement. Caution must be used to avoid attributing the locally perceived type and time of movement as the only movement along the system. The strong flow-foliation, mylonite and folds of the Late Precambrian granitic rock along the fault indicate the main period of deformation was prior to the

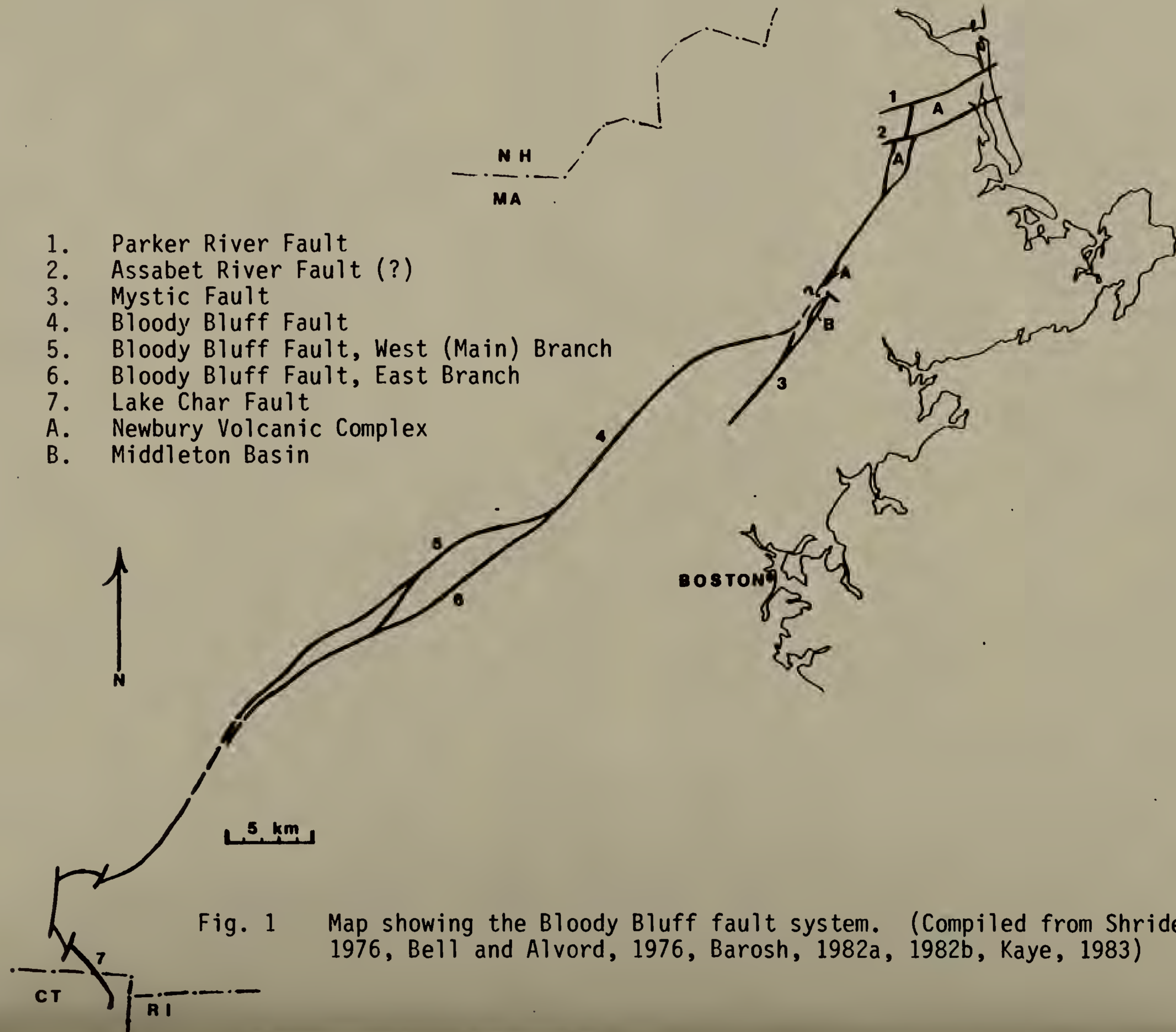

Fig. 1 Map showing the Bloody Bluff fault system. (Compiled from Shride, 1976, Bell and Alvord, 1976, Barosh, 1982a, 1982b, Kaye, 1983)

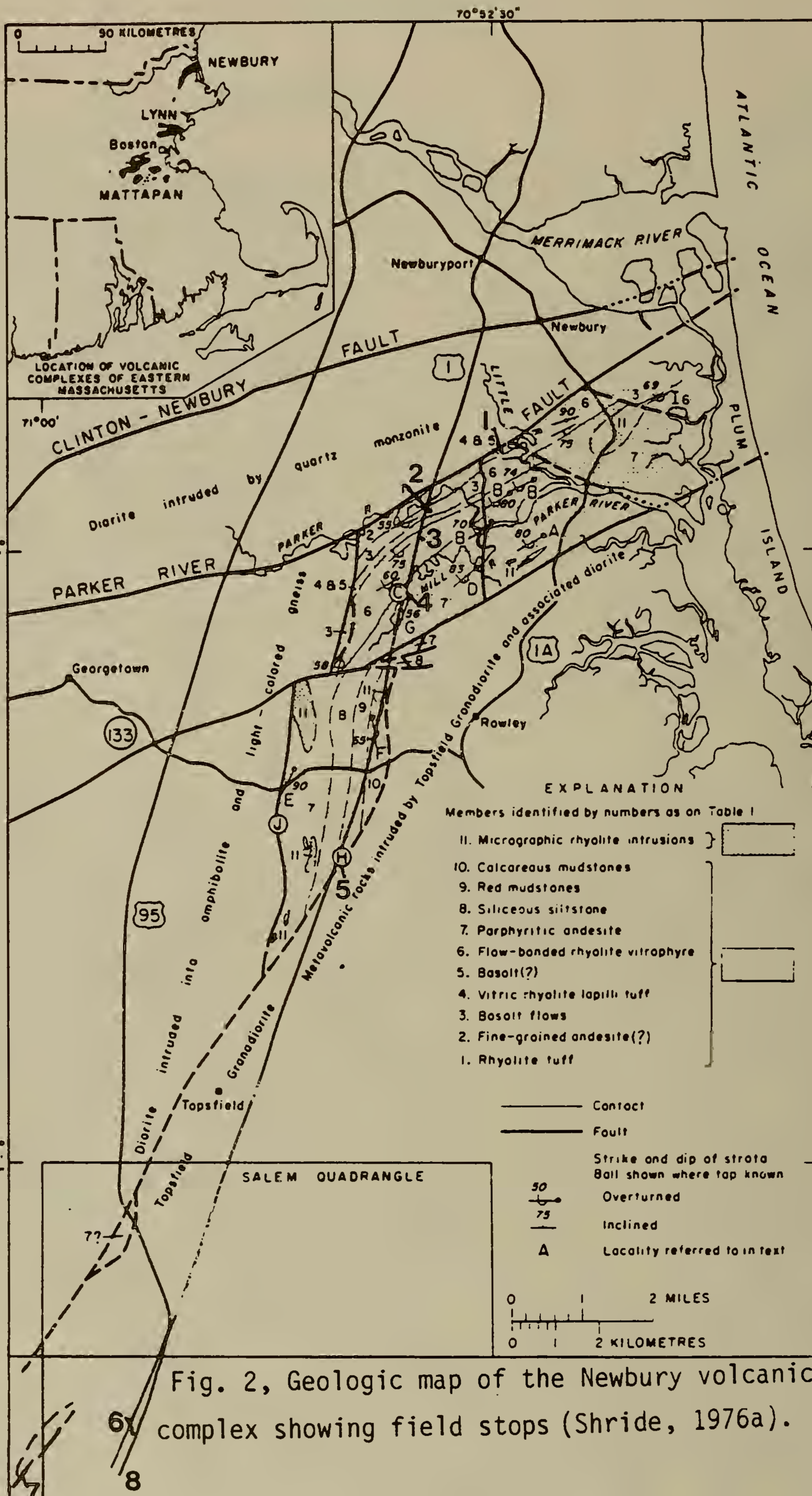

Fig. 2, Geologic map of the Newbury volcanic complex showing field stops (Shride, 1976a).

much lessor deformed Ordovician intrusive rock. Furthermore, the spatial relations of the syntectonic structures in the Late Precambrian intrusive rock to the fault system shows movement along the system at that time. These rocks were deeply buried then and the movement produced ductile features accompanied by intrusions. The Ordovician and younger rock are effected by more brittle structures. The Ordovician intrusive rock is highly fractured locally along the fault system; far more than the fault slivers of unmetamorphosed Late Silurian-Early Devonian and Late Triassic-Early Jurassic volcanic and sedimentary rock in the zone.

The purpose of the field trip is to show some of the variety of types and ages of rock along the Bloody Bluff fault system and different types and amounts of deformation that effect them that illustrate both the antiquity of the fault system and some of its numerous reactivations (Figs. 1, 2 and 3 and 4). The first part of this trip follows part of a previous one by Shride (1976b) and those wishing to know more about the Newbury volcanic complex should follow the rest of his trip. Those interested in features farther southwest along the Bloody Bluff can follow an earlier guide to the area (Barosh, 1982).

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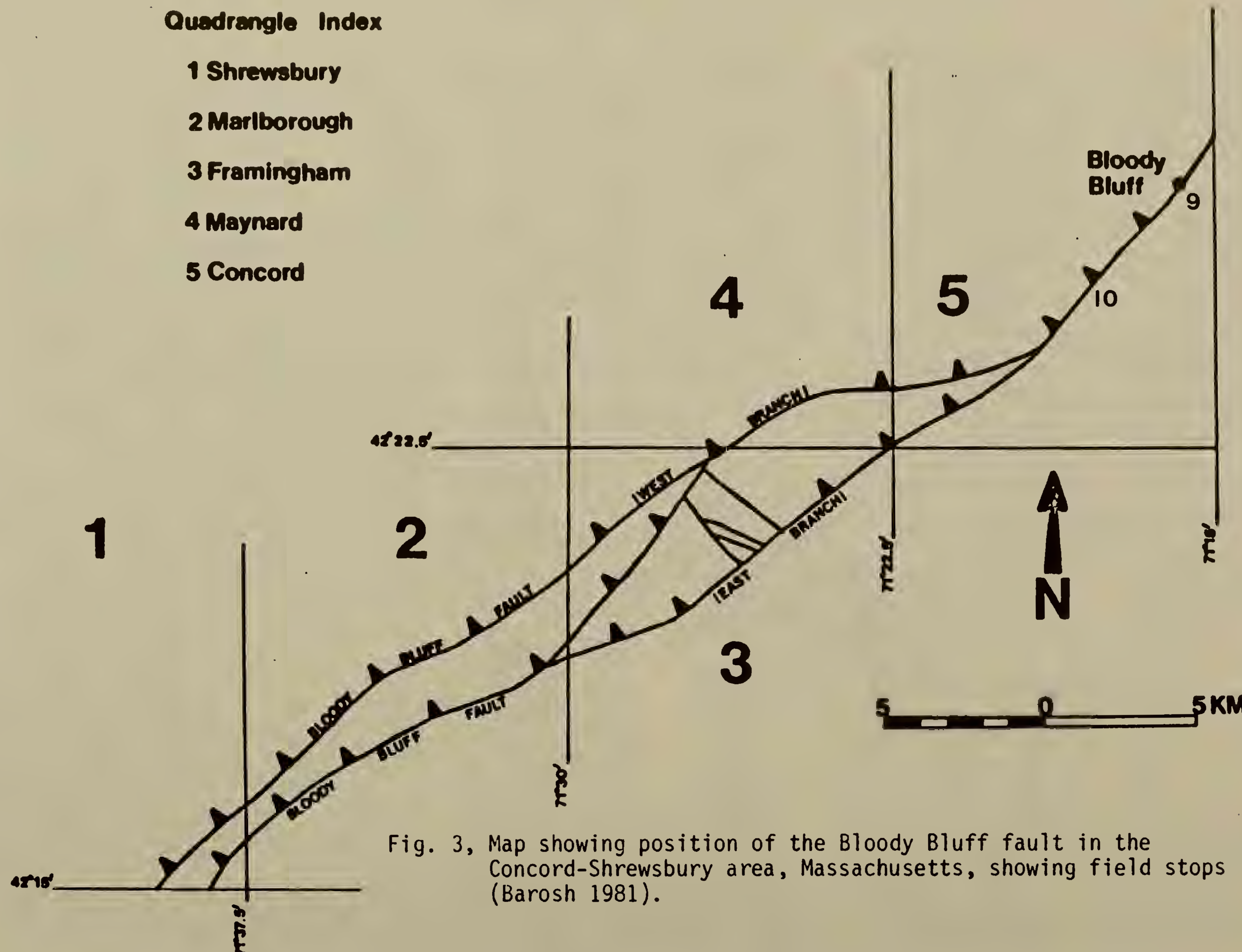

Fig. 3, Map showing position of the Bloody Bluff fault in the Concord-Shrewsbury area, Massachusetts, showing field stops (Barosh 1981).

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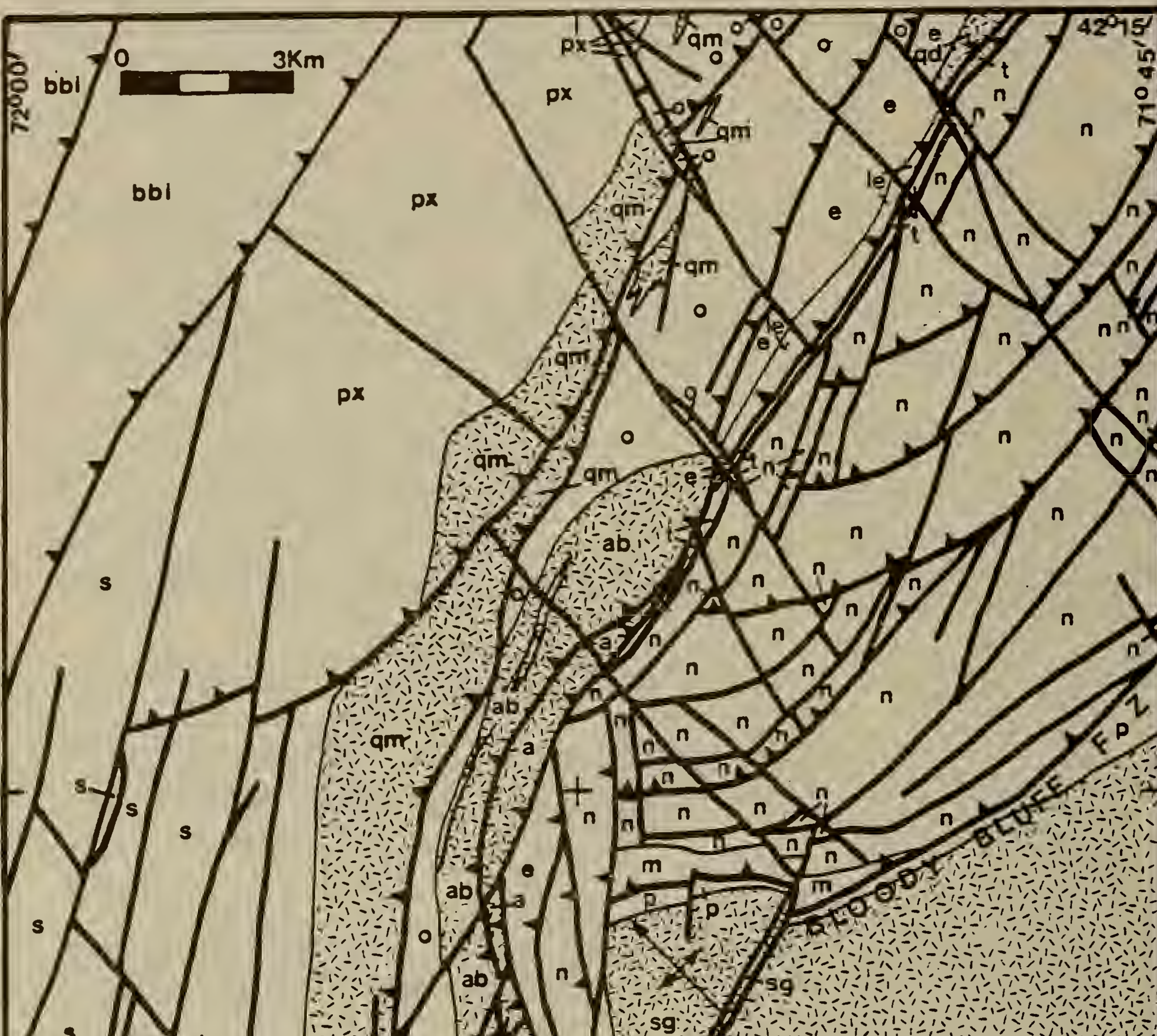

Fig. 4, Geologic map of the Webster-Oxford area, MA, CT, and R.I. showing relation of the Bloody Bluff and Lake Char fault zones (Barosh, 1982).

EXPLANATION FOR GEOLOGIC MAP OF THE WEBSTER - OXFORD AREA, MA, CT, and RI

METASEDIMENTARY ROCK

Late-Sil.-Early Dev.	e	Eliot Fm.
	le	lower Eliot Fm.
		fault
		Brimfield Group
pre-Ord.	bbl	Bigelow Brook Fm., lower mb.
	px	Paxton Group
	o	Oakdale Fm.
		fault
pre-Ord. Pre-€ (?)	t	Tadmuck Brook Schist
	n	Neshoba Fm.
	m	Marlboro Fm.
		fault
Pre-€	p	Plainfield Fm.

INTRUSIVE ROCK

Dev.	c	Canterbury Gneiss
	qd	Quartz Diorit
	qm	Muscovite Quartz Monzonite
Sil.-Ord.		Ayre Intrusive Complex
	a	Undivided
	ab	Biotite Quartz Monzonite
Pre-€	sg	Sterling Plutonic Group

SYMBOLS

Contact

Fault, thrust where teeth

Anticlinal axis showing plunge

SOURCES OF DATA: Barosh 1974, 1976, and 1977 and Barosh and Johnson 1976.

Road Log

MAPS - 7.5-minute quadrangle maps covered in this road log are the Newburyport East, Newburyport West, Georgetown, Salem, Reading and Concord.

<u>MILEAGE</u>

0 Rt. 1 (Newburyport Turnpike) and Boston Rd. (1.1 mi. north of Rt. 1 and Parker River and 13.5 mi. north of Rt. 1 and Rt. 95). Travel east on Boston Rd.

0.3 Veer right at fork onto Hay St.

1.0 Turn around at railroad crossing and go west on Hay St.

1.1 Park just beyond outcrops along north side of road.

STOP 1, Parker River fault. This east-northeast-trending fault zone along the Parker River marks the boundary between the Newbury volcanic complex to the south and intrusive rock in the Nashoba Thrust Belt to the north. This breccia of quartz monzonite marks the north border of the fault zone (Shride, 1976b). The rock shows the effects of the brittle fracturing and has undergone some alteration resulting in local silicification and iron coatings and staining. Several fault surfaces are exposed. They trend N65-90E and dip 60-90N and have prominent slickensides that plunge 30NE and some steeper plunging faint striations.

Continue west on Hay St.

2.0 Turn left, south, onto Rt. 1 after stopping.

3.1 Cross Parker River.

3.2 Turn left, east, at south end of bridge into parking area just to east and park. <u>CAUTION POISON IVY AND POISON SUMAC</u> everywhere.

STOP 2, Structural wedge of Newbury Volcanic complex along the Bloody Bluff fault system. Note the lack of foliation and metamorphisum in this Late Silurian or Early Devonian rock. "Exposures of the lowest two members of the Newbury are in the salt marshes to the N. The rhyolite tuff outcrop (member 1), 1,000 feet NW of the Parker River bridge, is well preserved considering that it borders the Parker River fault zone. Along the SE margin and on top of the island 500 feet NW of the bridge, note the variety of textures and structures in the fine-grained andesites (member 2).

From the parking site, members 3,4,5, and 6 can be crossed in succession by walking 1,500 feet NE along the south bank of the marsh and then wending a route southward across the more prominent outcrops that will come into view. In the basalt flows (member 3) note the uniform fine texture, lack of detrital materials, and difficulty of recognizing flow boundaries--which are usually in the swales. Outcrops of members 4 and 5 are few and border the marshy meadow S of the basalt ridge. Flow-banding, which becomes wider spaced upward, and intricate slump structures are readily seen in the rhyolite vitrophyre (member 6); zones of spherulites and lithic tuffs may be crossed. After reaching the bluffs that overlook the marshes of the Mill River, a traverse along the S margins of the islands that extend E to the junction of the Parker and Mill Rivers provides good views of the flows that dominate the basal 600 feet of the porphyritic andesite member (member 7). Volcaniclastic strata will be dominant higher in the member. On the W end of the island, 500 feet SW of the river junction, two conglomerate beds contain water-worn rhyolite pebbles derived from member 6, providing one item of the evidence for an overturned section. This conglomerate can be traced NE to Little River." (Shride, 1976b).

Return to highway and turn left, south, onto Rt. 1.

3.4 Low roadcuts on both sides of road are of the vitric rhyolite lapilli tuff member of the Newbury volcanic complex, in which many pumice fragments are tabular, as though flattened, but shards of the matrix are only slightly compacted (Shride, 1976b).

3.6 Outcrop at entrance of Old Newbury Golf Club on left is the dense basalt (?) that intervenes between the vitric rhyolite lapilli tuff and flow-banded rhyolite vitrophyre members (Shride, 1976b)

3.7 Park on right opposite prominent road cut.

STOP 3, Newbury volcanic complex. Transitional zone at the top of the tuff that here locally constitutes the basal 400 feet of the flow-banded rhyolite vitrophyre member (member 6) (Shride, 1976b). Few near vertical fault surfaces roughly parallel the road; some with 30N-plunging slickensides. Note laminated tuff at south end of west roadcut; similar Precambrian volcanic rock is present at STOP 10.

Continue south on Rt. 1.

4.8 Cross Central St. (east side) and Glen St. (west side), park just beyond intersection and walk to large roadcuts on Central St.

STOP 4, Newbury volcanic complex. "Intercalated flows and water-laid ash-fall(?) tuffs of the porphyritic andesite (member 7). The first reported fossils ("marine types" comprising "one or more species of brachiopods, a species of gastropod, fragments of crinoids, and probably a pelecypod"--see Emerson, 1917, p. 163) from the Newbury were collected here, apparently at the road intersection in rocks now deeply buried. Similar thin fossil zones exist in the vicinity, and additional collections were made in the 1960's; the sites may not be accessible now. Note zones of alteration and the differences between flows and tuffs." (Shride 1976b).

Very dark gray to greenish gray nonfoliated fine-grain, porphyritic and amygdaloidal rock present. Several small faults of various trends and bearing vertical to horizontal slickensides are exposed. One prominent near vertical fault trends N70E and has 30N plunging slickensides.

Continue south on Rt. 1.

7.1 Cross Rt. 133 at signal.

Large boulders of the underlying red mudstone member of Newbury volcanic complex on south side of Rt. 133 50 m east of intersection. The rock is Maroon mudstone to siltstone with rare to common pale yellow green spheres to irregular lenses, that appear to be slightly more siliceous concretions, fossil and burrow fillings.

7.9 Cross Ipswich town border.

8.4 Turn left into parking lot at Omni Electrical Supply, park and walk 70 m to low bluff behind two-storied house.

STOP 5, Newbury volcanic complex. This"is the only known outcrop of the member. Elsewhere, the red micaceous mudstone seen here occurs as concentrations of shaly of flaggy detritus in the soil or of friable erratics distributed to suggest that the member may be several hundred feet thick. Chips of the yellowish-brown, medium- to coarse-grained sandstone, found here as lenses, can be seen in soils elsewhere; but the andesitic(?) pebbles, cobbles, and flow rock seen toward the north end of this outcrop have not been recognized elsewhere." (Shride, 1976b). Most of the outcrop is gray to purplish gray arkose, that is laminated in part, and maroon mudstone and siltstone, that strikes N3SE and is overturned to dip 65NW. Many very delicate small-scale sedimentary structures are preserved. The southeast border fault lies a short distance to the southeast (Shride, 1976b).

Turn left, south, back onto Rt. 1.

11.1 Cross Ipswich Rd. at signal

12.2 Cross Rt. 97.

15.1 Cross Rt. 95.

15.3 Turn left, south, onto Rt. 95 south.

17.1 Turn right onto exit for Centre St., Danvers.

17.3 Turn right, west, onto Centre St. and park near Rt. 1 overpass. Roadcuts both sides.

STOP 6, Salem Gabbro-Diorite (?). This dark gray rock is characteristically variable in both texture and composition. It is very similar to the Salem Gabbro-Diorite southeast of the fault system, although it was mapped as the Marlboro (?) formation by Toulmin (1964). Generally nonfoliated although some flow-banding (N65-75E, 70NW) is present on the north side of the road near the overpass for Rt. 1. Few small faults and numerous fractures present. The widely distributed rock is only found southeast of the Bloody Bluff fault system and its mafic nature produces relatively high magnetic response. The termination of magnetic highs over this unit is responsible for the Bloody Bluff being so well displayed on the aeromagnetic maps.

Continue under overpass.

17.4 Turn right, north,

17.5 Turn right, east,

17.6 Turn right, south, onto Rt. 1.

19.1 Turn right, west, onto exit for Lowell St., Peabody

19.3 Turn left, west, onto Lowell St.

19.6 Veer right at fork.

20.0 Veer right onto Russell St. at fork just beyond Crystal Lake.

20.7 Turn right, north into entrance for Essex Bituminous Concrete Corporation, stop at office for permission to enter and proceed to quarry beyond processing equipment. Break in road log.

STOP 7, Mystic Fault and Late Triassic-Early Jurassic rock of the Middleton Basin. The quarry exposes highly faulted, sheared and altered, generally finer-grained varieties of probable Salem Gabbro-Diorite. The faults generally trend northeasterly and are near vertical, but trends of all directions are present. The faults are brittle and the surfaces are slickensided. This is the Mystic fault zone of Bell and Alvord (1976), a part of the Bloody Bluff fault system. A few large faults apparently come together in this area and this probably accounts for the unusual fracturing here.

A fault wedge of reddish gray pebble to cobble conglomerate and arkose and maroon siltstone to mudstone is exposed at the northwest edge of the quarry. This rock has not undergone the alteration and deformation of the rock in the quarry and is separated from it by a fault trending N43E, and dipping 65NW. The rock is similar to that at STOP 5, but recently plant fossils of the Newark Group were found and the block of rock has been named the Middleton basin (Kaye, 1983).

Return to entrance on Russell St. and resume road log. Turn left, east, and retrace route towards Rt. 1.

22.2 Pass under Rt. 1 on Lowell St.

22.3 Pass under Rt. 95.

22.4 Turn left, north, on Bourbon St.

22.6 Park on right before entrance to Ledgewood Condominiums.

STOP 8, Salem Gabbro-Diorite. Glacial-polished pavement to the northeast and old quarry walls around the tennis court show a highly variable Salem Gabbro-Diorite and some possible dikes of Cape Ann Granite. The rock is much less deformed than at the last spot, but displays a very interesting fracture pattern of closely intersecting small faults of various trends and , at least, slightly differing ages.

Turn around and return to Lowell St.

22.9 Turn right, west, onto Lowell St.

23.2 Turn left, south, into entrance ramp for Rt. 1 south.

23.4 Turn right, south, onto Rt. 1

24.9 Turn right into connecting road for Rts. 95 and 128.

25.3 Veer right into entrance for Rt. 95 south.

25.8 Enter Rt. 95 (128) south at Lynnfield town line.

33.2 Pass under Rt. 93.

38.9 Pass under Rt. 3 north.

42.3 Turn right at exit 45B for Concord and Hanscom A.F.B.

42.5 Turn right, west, onto Rt. 2A. Outcrops just to west and to south along entrance ramp to 128 south are nonfoliated Ordovician Salem Gabbro-Diorite with pendants, xenoliths and xenocrysts of sheared to mylonitized Precambrian Dedham Granodiorite.

42.8 Cross intersection with signals.

42.9 Right turn into parking lot for Minuteman National Park and park.

STOP 9, Bloody Bluff fault zone. Bloody Bluff, that lies across the road to the west, beneath the power lines, exposes shattered Dedham Granodiorite that has undergone limonitic and chloritic alteration. Many fault surfaces are exposed, but highly sheared or mylonitized rock is uncommon here. The fault zone is at least 0.5 mile wide here. The bluff is a block bounded by a fault along the northwest side (which marks the contact between the Dedham and the Andover Granite to the west) and another fault along its southeast side. The latter fault passes through the southeast corner of the electric substation seen to the south. East of this fault, the Dedham is sheared to mylonitized and is crosscut by gabbro that postdates the shearing. The gabbro is well exposed along the powerline just south of the substation. Return to parking lot and walk east on footpath 50 m beyond footbridge, turn right into woods and go about 80 m, paralleling Rt. 2A, to roadcuts along a side road. Highly sheared Dedham crosscut by gabbro, part of Salem Gabbro-Diorite, which although not foliated is cut by later faults.

Turn right, west, from parking lot and continue along Rt. 2A.

43.7 Pass monument commemorating site of Paul Revere's capture on right.

45.3 Veer left at fork in road.

45.9 Continue straight upon joining Rt. 2 at signal.

47.0 Turn left, south, onto Rt. 126.

47.4 Pass Walden Pond on right.

49.0 Pass Codman Rd. on left.

49.4 Turn left, east, on Rt. 117 at signal.

49.9 Cross Lincoln Rd.

50.4 Behind red barn 100m to right and along ridge on right to railroad are scattered outcrops of nonfoliated medium=grained Ordovician (?) granodiorite, that only occurs northwest of the Bloody Bluff fault.

50.7 Cross railroad.

50.8 Cross swale in topography marking center of Bloody Bluff fault.

50.9 Turn right into Bowles Terrace and park. Walk back across Rt. 117 to outcrop and continue along 117 across Tower Rd. (0.1 mi.) and outcrops for 0.1 mi. farther.

STOP 10, Bloody Bluff fault zone. Very detailed studies here have revealed much about the fault history. Southeast side of fault is a complex of foliated and locally mylonitic Precambrian Dedham Granodiorite and older rhyolitic and mafic tuff extensively intruded by Salem Gabbro-Diorite and possibly younger mafic dikes. The tuffs exposed in the roadcuts are felsic, similar to those at STOP 3, more mafic ones crop out in the woods. These are commonly mistaken for mylonite. The outcrops across from Bowles Terrace are mainly Dedham cut by Salem. Those beyond Tower Rd. are largely laminated tuff cut by Salem.

Foliation in the Dedham and volcanic rock describe a ductile drag fold that formed and probably rotated slightly prior to the intrusion of the nonfoliated Salem and the more brittle deformation after the Salem. The main deformation along the fault is pre-Salem and probably Late Precambrian in age.

End of field guide.

Return to Rt. 2 for points to north or northwest. Continue southeast along Rt. 117 for junction with Rt. 95 (128) at Rt. 20 for points south and west (via Rt. 90 - the Mass. Pike.)

OPTIONAL STOP For those traveling west or southwest, an excellent example of ductile shearing and drag folding within fault slices along faults associated with the Bloody Bluff fault and the high-grade metamorphic rock east of it can be seen to the southwest at Oxford, MA. Take Rt. 90 (Mass Pike) west to the Auburn exit, thence south on Rt. 52. Stop opposite a high roadcut on the left, east, side of the roadway 4.3 mi. after passing under Rt. 20 and 2.4 mi. after passing the Oxford town line (see Barosh, 1982).

STOP 11, Thrust-fault complex cutting the Nashoba formation. The Bloody Bluff fault zone crosses beneath the highway at the pond just south of Sutton Ave. overpass 2.0 mi. south of here. The thrusting here is related to movement on the Bloody Bluff and show earlier and deeper structural features than generally seen to the north. The offsets and drag folds here demonstrate north over south movement here in the Nashoba formation, a unit that only occurs to the northwest of the Bloody Bluff and is of Pre-Ordovician and probably Precambrian age. Note the pegmatite along the earlier structures here.

Complex of moderately north-dipping shears, thrust faults, and overturned folds with north-dipping axial planes. Consistently north over south transport, which is a local deviation from the regional northwest over southeast transport. The foliation and mylonitization parallels relict bedding. The observed layering is thus a result of a combination of causes. Many of the thrust faults have pegmatite, both foliated and nonfoliated, along them. Field evidence suggests that many of these developed during thrusting; porphyroblasts of feldspar and quartz form in blastomylonites, increase in amount and coalesce, forming pegmatites. The geochemistry of this type of mineral growth was studies by Wintsch (1975). In this area, the Marlboro-Nashoba sequence has undergone radical tectonic thinning by omission along both the Bloody Bluff and Clinton-Newbury fault zones and a series of internal faults. The exposed thickness decreases from about 18,000 m northwest of Boston (Bell and Alvord, 1976) to less than 1,000 m south of Oxford center.

SILURIAN AND DEVONIAN ROCKS IN THE ALTON AND BERWICK QUADRANGLES NEW HAMPSHIRE AND MAINE

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Introduction

This trip will examine stratigraphic, structural and metamorphic relations in staurolite + andalusite to sillimanite + white mica grade, polydeformed metasedimentary rocks on the extreme southeast limb of the Kearsarge-Central Maine Synclinorium (KCMS; Lyons and others, 1982; the former Merrimack Synclinorium of Billings, 1956). We will primarily be examining new outcrops along the Spaulding Turnpike north of Rochester, New Hampshire to establish the stratigraphy and some of the more critical structural and metamorphic relationships between the Maine-New Hampshire border and the southwestern end of the Blue Hills Range (Parker Mountain, Strafford, New Hampshire) (see Figure 1B).

Geologic Setting

The metasedimentary rocks of interest are primarily a mix of rusty, sulfidic and nonrusty pelitic schists and metaquartzites with subordinate calcsilicate schists. This stratigraphic package is intruded by elongate syntectonic (?) diorite bodies. They are similar to the Spaulding quartz diorite, dated by Lyons and Livingston (1977) by Rb-Sr whole rock techniques at 393 ± 5 Ma. (age was recalculated using the presently accepted decay constant, $\lambda = 1.42 \times 10^{-11}\ yr^{-1}$, Steiger and Jager, 1977), and Winnipesaukee quartz diorite, both members of the New Hampshire Plutonic Series. Post-tectonic two-mica granites and quartz monzonites also cross cut the sequence. They resemble lithically the nearby Sebago, Effingham and Lyman plutons dated by Hayward and others (1984) and Gaudette and others (1982) at 322 ± 12 Ma. (Rb-Sr whole rock).

The Nonesuch River/Campbell Hill/Hall Mountain strike slip fault system, with strongly debated relative motion (see Zen 1983, p. 71-73 and Lyons and others, 1982, p. 55-57 for details), forms the southeastern boundary of the KCMS in this area. Juxtaposed along the fault to the southeast are rocks of the Merrimack Trough considered Pre-Silurian (?) by Hussey (1984), Late (?) Ordovician to Precambrian by Lyons and others (1982, p. 54) and Precambrian (?) by Olszewski and others (1984). The Berwick Formation, a calcsilicate bearing, quartz + biotite granofels, crops out southeast of the fault in the trip area and is part of the Merrimack Group.

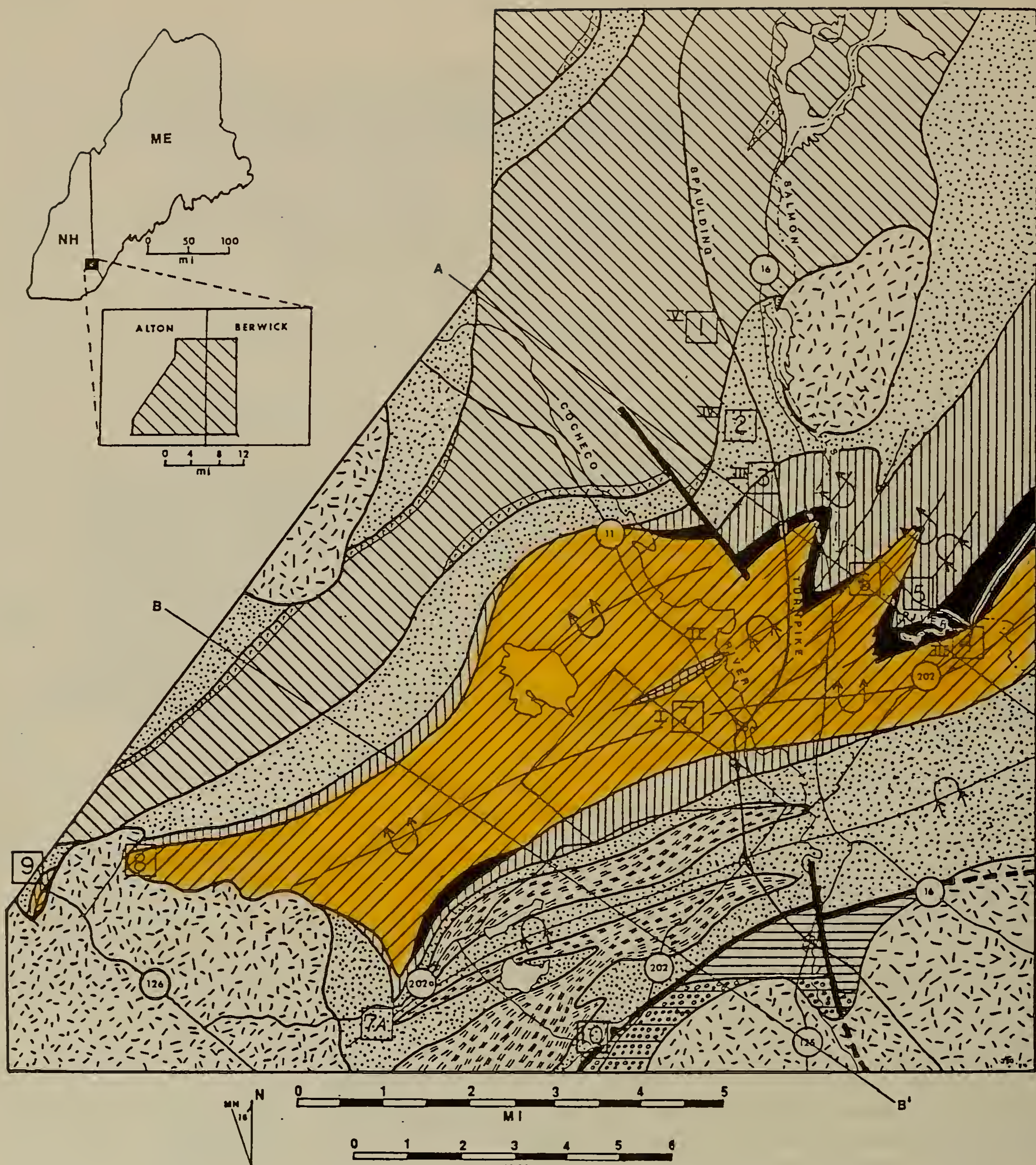
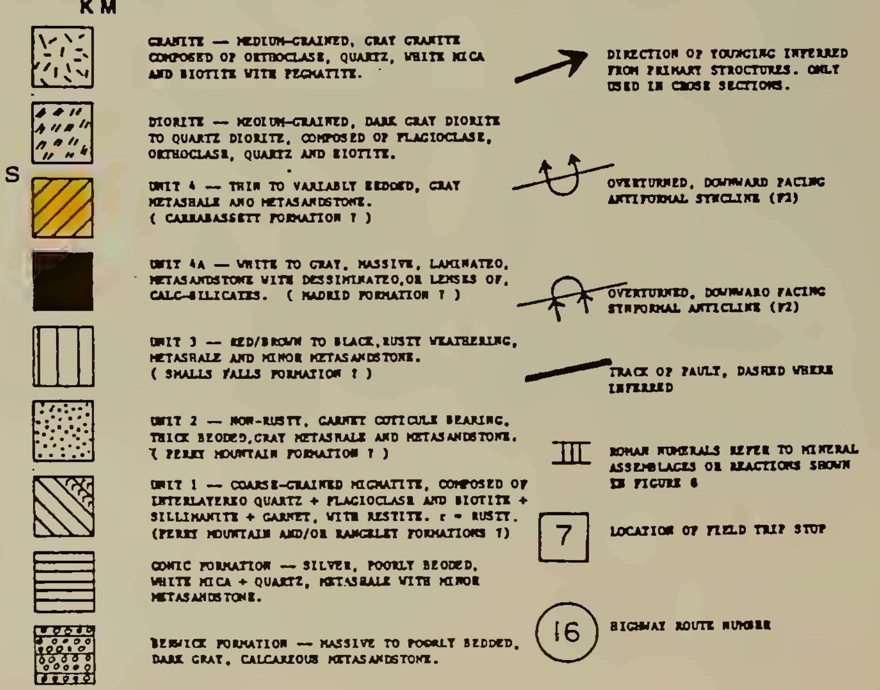
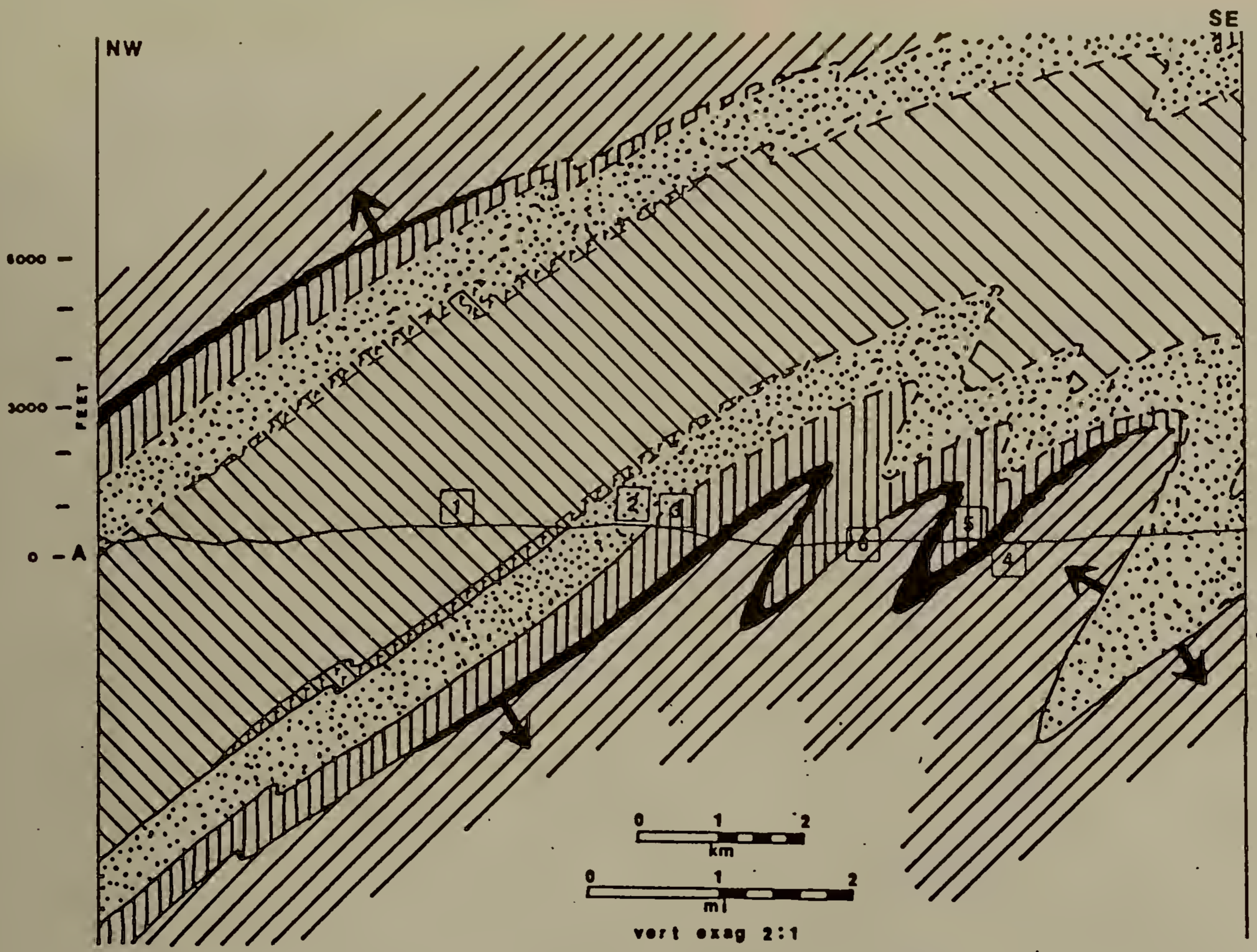


FIGURE 1A Geologic map and cross sections of the field trip area. Field trip stops shown on the cross sections are in places projected to the line of section.

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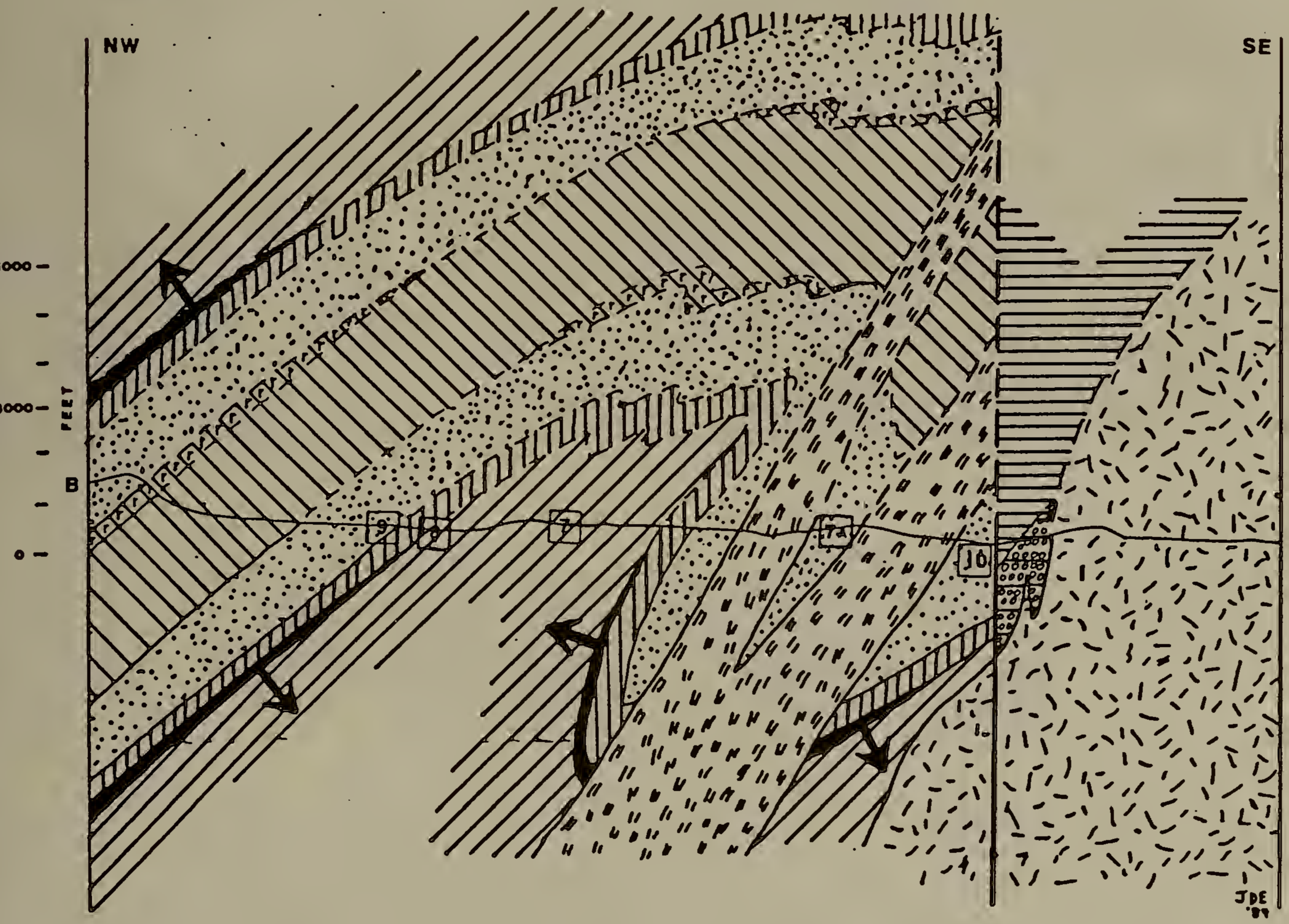
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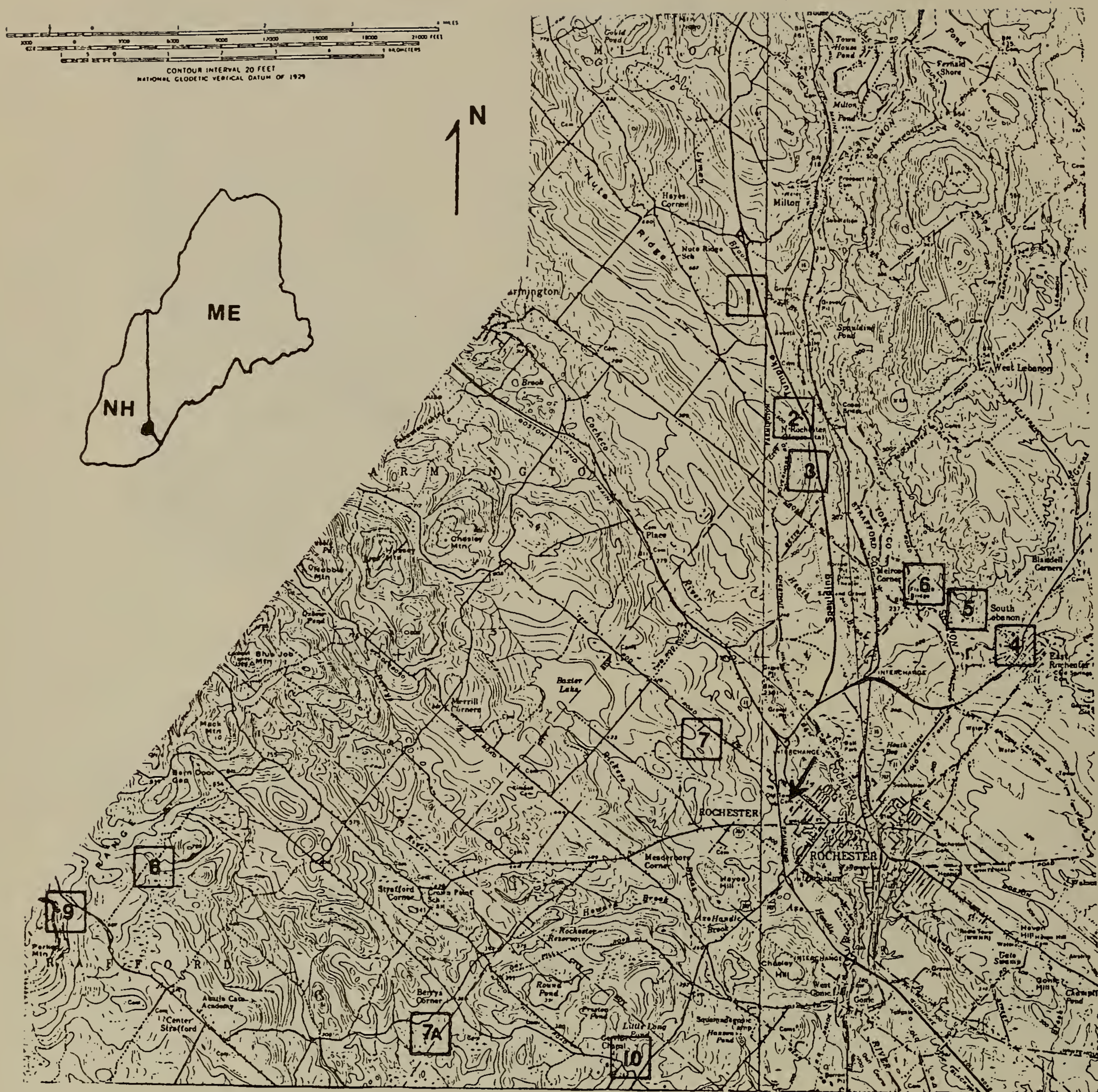

FIGURE 1B Geographic map of the field trip area. Numbers enclosed by square outlines are stop locations reffered to in the text. The thin vertical black line crossing the right center of the figure is the boundary of the Alton, NH (west) and Berwick, ME (east) 15' quadrangles. Tip of arrow east of STOP 7 marks the starting point of the trip.

STRATIGRAPHY

All of the metasedimentary rocks in the trip area were originally shown as the Devonian Littleton Formation by Billings (1956) in New Hampshire and the Rindgmere and Towow Formations by Katz (1917) in Maine. Hussey (1962, 1968) included the Towow and Rindgmere Formations in the Shapleigh Group which he correlated with the Littleton Formation. Gilman (1977, 1978) mapped the Shapleigh Group in the Newfield and Kezar Falls, Maine quadrangles. In New Hamsphire the Littleton Formation was subdivided into a number of distinct lithologic members (Heald, 1955; Stewart, 1961 and Carnein, 1976). All of the metasedimentary rocks were thought to be Devonian in age based on a lithic similarity to the fossiliferous Littleton Formation across the KCMS near Littleton, New Hampshire (Billings and Cleaves, 1934). Figure 2 outlines the stratigraphic correlations prior to this study.

Detailed mapping from this study has redefined what was previously a simple stratigraphy, into a more complex one based on lithic differences and good topping control from primary sedimentary structures, principally graded bedding and to a less extent, cross bedding. The newly recognized stratigraphy is remarkably similar to the Siluro-Devonian section described by Moench and others (1970) in a continuous part of the KCMS near Rangeley, Maine (see Figure 3). We propose to correlate, equivocally the Rangeley Formation, and with greater certainty the Perry Mountain, Smalls Falls, Madrid and Carrabasset Formations, defined by Moench and others (1970) and Boone (1973) to the lithic units of this field trip area. One of us (AMH) considers a correlation of the units of southwestern Maine and adjacent New Hampshire with the Carrabasset-Seboomook sequence of West-central Maine as another alternative (Hussey, 1984) (see Figure 3).

On a small scale map, the units near Rangeley, Maine, do not appear to be on strike with the rocks seen on this field trip. We suggest, however, that indeed the northeast trending, redefined stratigraphic package in southeastern New Hampshire and southwestern Maine, trends north, then northwest through southwestern Maine, recrosses into New Hampshire, where it trends back to the northeast near Conway, New Hampshire, and emerges on the north side of the White Mountains batholith on strike with the Rangeley, Maine, belt (Eusden, 1984). The preliminary compilation map shows the proposed stratigraphic connection between the field trip area and Rangeley, Maine (Figure 4).

The major impact of this correlation is a further extension of an older sequence into southwestern Maine and southeastern New Hampshire. The extension supports ongoing work in central and northeastern New Hampshire that has similarly extended the Rangeley stratigraphy on strike (Hatch and others, 1983; Thompson, 1984). Figure 3 summarizes the proposed correlations.

Because lithic similiarity is the basis of this correlation, the description, bedding character and facies relations of each unit within the southeastern New Hampshire sequence is presented below. Following the reasoning of Moench and Boudette (1970, p. A-1, 1), the usage of the terms metashale, metasandstone, etc. is preferred instead of their metamorphic equivalents. More detailed petrographic descriptions of each unit are

AGE	NH STATE MAP BILLINGS (1956)	SOUTHEASTERN NH HEALD (1955)		SOUTHWESTERN ME HUSSEY (1968)	
DEVONIAN	LITTLETON FM^F	LITTLETON FM	not recognized	SHAPLEIGH GROUP	Towow Fm
			Jeness Pond Mbr		Upper Rindgmere Fm
			Pittsfield Mbr		Lower Rindgmere Fm
LOWER			not recognized		Gonic Fm

FIGURE 2 Stratigraphic correlations of units prior to this study. All units were thought to be Devonian in age

AGE	Southeastern NH THIS STUDY	Rangeley ME Hatch et al (1983)	Central ME Pankiwskyj et al (1976)	Central NH Hatch et al (1983)	Western NH Billings (1956)	AGE	ALTERNATIVE THIS STUDY HUSSEY (1984)	
LOWER DEV	not exposed Unit 4	Seboomook^F Hildreths Carrabassett	not exposed Carrabassett	Kearsarge Mbr of Littleton Littleton	Littleton^F	DEVONIAN	SEBOOMOOK^F	Day Mtn Mbr.
UPPER SILURIAN	Unit 4A	Madrid	Fall Brook	Warner	Fitch^F			
	Unit 3	Smalls Falls	Parkman Hill^F	Francestown				Temple Stream Mbr.
MIDDLE SILURIAN	Unit 2	Perry Mtn		Crotched Mtn				
LOWER SILURIAN	Unit 1 (?)	Rangeley^F	Sangerville^F	Rangeley	Clough^F			Mt. Blue Mbr.
	not exposed	Greenvale Cv	not exposed	Greenvale Cv				
PRE-CAMBRIAN?	Gonic Berwick						Carrabassett	

FIGURE 3 Revised stratigraphic correlations of Siluro-Devonian units in the Kearsarge Central Maine Synclinorium. F = fossil control.

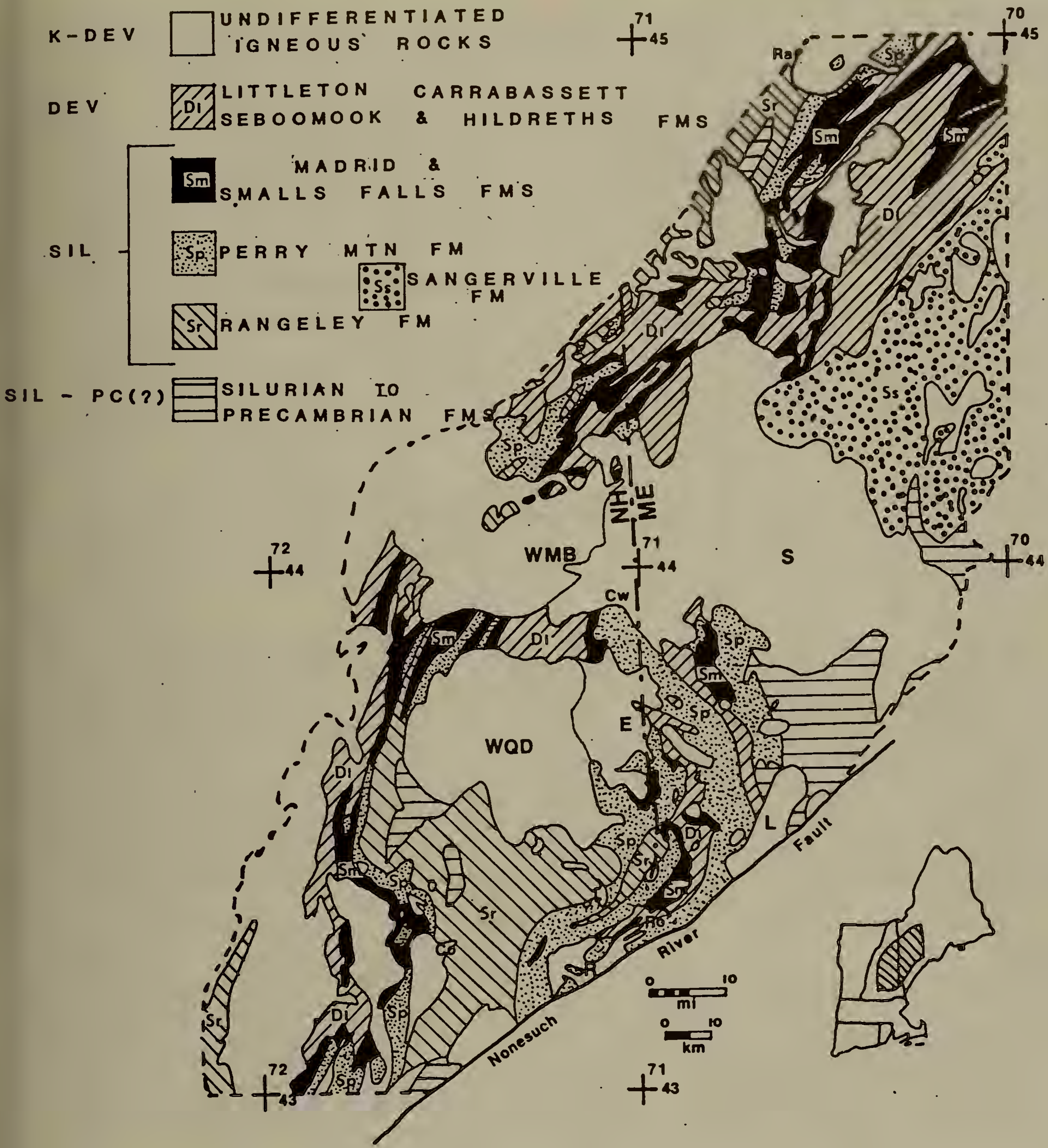

FIGURE 4 Small scale map showing the extent of Siluro-Devonian metasediments in part of the Kearsarge-Central Maine Synclinorium. The distribution of the metasedimentary rocks between the Effingham (E), Sebago (S) and Lyman (L) plutons is preliminary. This map was compiled using data from Hatch and others (1983), Gilman (1978 and 1977), Hussey (1968), Osberg and others (1984) and this study. WMB = White Mtn Batholith, WQD = Winnepesauke quartz diorite, R = diorite exposed around Rochester, NH, Co = Concord, Ro = Rochester, Cw = Conway, Ra = Rangeley.

summarized in Table 1. In brackets next to the subheading of each unit is our proposed correlation to the central-western Maine sequence of Moench and Boudette (1970).

	SAMPLE #	SH	SS	C	M	CS	Q	PLG	KSP	WM	BIO	CH	GA	ST	AD	SI	AM	EP	CA	SP	AP	OP	TM	ZR	RT	COMMENTS
UNIT 1	92 Oc	X					15			15	10	3F	6	1	50							Tr	Tr			Fresh andalusite
	Mb 54	X					10	5		15	20	1F	10		35	3						1				II Fig. 6
	Mb 63	X					10	3		20	15	10F	10		30					Tr		2		Tr		Sensitized andalusite
	14 Oc		X				55			25	14		5	1	1							2	Tr			Fresh andalusite
	96 Ic		X				70			10	7	5F								1		5	Tr			
UNIT 4A	128 Sm					X	45	20			10	1M	5				10	5	3	1				Tr		Calcsilicate lense within metasandstone
UNIT 3	97 Ssr	X					20	15		30	10	15F	Tr			7					Tr	3				IV Fig. 6
	27 Ssr^A	X					17			65	10			3		Tr						5				I→V Fig. 6
	25^A Ssr		X				70	6		2	15	Tr	2	2							Tr	3			Tr	
	25^B Ssr		X				80			2	15		Tr									3				
	26^A Ssr		X				60			23	15	Tr										1	1	Tr		Sensitized andalusite
	26 Ssr		X				50	4		20	15	2F	5									4	Tr			
UNIT 2	Mb 51	X					25	10		20	20	3F	10			10							2			Sensitized andalusite
	30 Spm	X					30			30	25	9F	5			Tr						1				
	109 Spm	X					30	10		20	20	2F	5			10		Tr				2	1.			SI from M2
	77 Spm	X					20	7		50	10	Tr	3								2	5	3			
	28 Spm		X				55	5		25	10	Tr	1	1								3	Tr			
	32 Spm		X				55	3		10	15		3		3	Tr						10	Tr	Tr		IV Fig. 6
	104 Spm		X				40	Tr		20	20	5F	10	?								5				
	106 Spm		X				55	3		5	10	15M	5								Tr	5	1		Tr	
	28 Spm			X			40			3	6	Tr	50								Tr	1	Tr.		Tr	ZONED GARNETS FIG. 7
	93 Spm			X			20			7	1		60		10?							1	1			
UNIT 1	Mb 63				X		30	15		30	5		3			10						1	3	Tr		All I Fig. 6 ↓
	Mb 61				X		17	10		25	20		10			15						3		Tr		
	37 Sa				X		50	Tr		15	15		3			15						2				
	38 Sa				X		40	3		10	15		5			15						12				
	40 Sa				X	X	60	10					15				10	Tr		1		3				
	83 Sa				X		25	5	Tr?	35	3		3			20				2		7				SI + KSP ???
	GONIC FM	X					25			25	20	5F	10	10								4		Tr	Tr	
	BERWICK FORMATION		X				40	25			20	10F						Tr	Tr			4				
						X	40	40			5						10	Tr	1	3		1				

TABLE 1 Table of estimated modes and other petrographic data for the metasedimentary rocks of the field trip area. Roman numerals in comments section are explained in Figure 6. Abbreviations: SH-metashale, SS-metasandstone, C-garnet coticule, M-migmatite, CS-calcsilicate rock, Q-quartz, PLG-plagioclase, KSP-potassium feldspar, WM-white mica, BIO-biotite, CH-chlorite(F-Fe rich, M-Mg rich as determined using the method of Albee,1962), GA-garnet, ST-staurolite, AD-andalusite, SI-sillimanite, AM-amphibole, EP-epidote, CA-carbonate, SP-sphene, AP-apatite, OP-opaques, TM-tourmaline, ZR-zircon, RT-rutile, Tr-trace amount.

Unit 1 (Perry Mountain Formation and part B of The Rangeley Formation?)

Unit 1 is a coarse-grained migmatite composed of intercalated quartzo-feldspathic (plagioclase) and biotite + garnet + sillimanite schistose layers with moderate to abundant secondary white mica. Layering is swirly with a great range in attitude. Sillimanite is locally abundant and has been observed in bundles 8 to 10 cm long and 5 cm in diameter. In places, outcrops have a distinctive yellow stain, probably an alteration product of sulfur-bearing phases. Elongate lenses of zoned calcareous and minor quartzitic concretions are preserved as restite within the migmatite. A mappable, but subordinate rusty weathering migmatite is found close to the transition from migmatite to non-migmatized, well-bedded, Unit 2. This is coarse-grained, dark brown rusty weathering migmatite composed of white mica + sillimanite + quartz + minor tourmaline (see Figure 1A).

Unit 2 (Perry Mountain Formation?)

Unit 2 is a metaturbidite composed of well bedded metasandstone and metashale. Bedding generally thins up section. The thickest beds are approximately 0.5 m and have about four times more metasandstone than metashale. Near the contact with Unit 3, the beds thin to 5-10 cm with equal amounts of metasandstone and metashale.

Bedding is well preserved and is emphasized by the development of abundant, large (≈5 cm) knobby, black weathering, porphyroblasts of andalusite partially or completely pseudomorphed by white mica. These are colloquially referred to as "andalumps" (P. Robinson in Hatch and others 1983).

Despite the well preserved bedding it is at times difficult to tell topping direction in this unit. This is due in part to the "fast graded" character of the beds marked by an abrupt gradation between the metasandstone and metashale within a single bed. In the classic Bouma series, only divisions A and E are observed. Rip up clasts of metashale are infrequently preserved in the metasandstone. These sedimentologic observations suggest that the basal metasandstone represents an environment where all of the turbidity currents during deposition were fast flowing and during quiescence were covered by a relatively thin layer of pelagic mud. Walker (1979) suggests a proximal source for similar turbidities elsewhere.

Distinctive pink colored garnet + quartz coticules occur in the metasandstone as discontinuous stringers and pods that parallel original bedding and often outline folds beautifully. These may represent maganiferous chert layers, but the origin of such a sediment is unclear (Docka, 1984; Hatch and others, 1983) as is also the relation of this peculiar feature to the better constrained paleo-environment previously mentioned. The coticules are in part one of the distinguishing criteria used in separating units 2 and 4, or the Perry Mountain and Carrabasset Formations, two very similar turbidites in this sequence.

The contact between Units 1 and 2 is defined by the first appearance of migmatite. This is an indication of change in metamorphic conditions rather than an exact stratigraphic contact. In fact, recognizable restite, particularly the coticules characteristic of Unit 2 is found in the migmatite near the inferred contact. Elsewhere within the migmatite is a distinctive rusty weathering zone that is stratigraphically below Unit 2, based on facing directions nearest the contact. This rusty migmatite, presumably older than Unit 2, may correlate with part B of the Rangeley Formation as described by Moench and Boudette (1970). Because of these observations we propose to correlate Unit 1, in essence a metamorphic unit, to the lowermost portion of the Perry Mountain Formation and to at least part B of the Rangeley Formation in central-western Maine. All of Unit 2 is correlated to the Perry Mountain Formation.

Unit 3 (Smalls Falls Formation?)

Unit 3 is a distinctive rusty weathering, crumbly, well foliated poorly bedded metashale and metasandstone. At and away from the contact with Unit 2 the bedding character remains generally the same as that described for Unit 2. The abrupt change to a distinctive rusty color marks the contact between the two units. Up section the unit becomes much more shaley with only scattered pods and stringers of metasandstone remaining. Unfortunately, this bedding style change is not systematically well defined.

At STOP 6, unit 3 has lenses of 2-10 m thick, rusty weathering quartz pebble grit with angular to subrounded vitreous quartz clasts and gray lithic fragments up to 0.5 cm in size, often tectonically elongate, within a fine-grained quartz + biotite + minor white mica matrix. Hussey distinguishes between this grit and a grit at the base of the Towow Formation (Unit 3 equivalent) as seen outside the field trip area in the Berwick, Maine quadrangle. An increase in blue, vitreous quartz clasts and the lack of lithic fragments distinguishes the latter from the former for him.

Graded bedding is poorly preserved or absent due to the more homogeneous shaley nature of this unit. Hence stratigraphic position has been determined by careful examination at the contacts with Units 2 and 4 (STOPS 3, 6, 8 and 9) that have well preserved graded bedding and/or cross bedding. The bedding style, rusty weathering, graphite and pyrrhotite suggest that this unit may represent more distal turbidites deposited in an oxygen restricted basin (see also Moench and Boudette, 1970).

The outcrop pattern of unit 3 is quite variable (see Figure 1A). In places it thins sharply to the point that only a few meters of Unit 3 intervene between Units 2 and 4. This is unlikely a function of fold geometry alone but at least partly an expression of the original shape of the euxinic basin which was probably geometrically irregular. Unit 3 is correlated to the Smalls Fall Formation of the central-western Maine section.

Unit 4 A (Madrid Formation?)

Unit 4A is a white to gray, laminated to massive metasandstone with numerous zoned calcsilicate pods or concretions. The lamination is rhythmic in places and is defined by black to dark gray biotite rich layers at about 1-3 cm intervals within the metasandstone. Where massive, the color is much whiter with 10-20 cm long calcsilicate pods intercalated with the metasandstone. The calcareous pods are zoned with biotite rims and amphibole + plagioclase + garnet cores. Fractures and joints in this unit are commonly silica-filled and stand out as resistant mini-ridges on the outcrop surface.

Only four exposures of Unit 4A have been observed in the field trip area and no definitive topping features are known. It is more common to see the contact between units 4 and 3 without any intervening 4A.

This unit is thought to represent an influx of more proximal, higher energy clastics perhaps associated with the waning front of a slightly calcareous sandstone delta similar to the ensimatic carbonate flysch recognized by Roy (1984) in the Maine Slate Belt. Unit 4A and 4 mark a significant change in the depositional regime. Whereas units 2 and 3 may have been deeper water proximal to distal turbidities with variable oxygen supply, Unit 4A may be a more proximal sediment reflecting influx from a new source or a drop in sea level or both. Unfortunately, very few paleocurrent indicators have been found in any of the units to establish a source direction. This is in part due to the high metamorphic grade and deformation observed in this area.

Unit 4 - (Carrabassett Formation?)

Unit 4 is a well bedded turbidite, in places quite similar to Unit 2. This unit is typically rhythmically bedded with thicknesses on the order of 3-5 cm, with slightly more metasandstone than metashale. The metashale is often graphitic and has well developed andalusite (chiastolite) and staurolite that are fresh or only partially pseudomorphed by white mica and chlorite.

When not rhythmic, the bedding style is quite variable. Beds of up to 3 m thick with roughly five to six times more metasandstone than metashale are commonly observed as are thick beds of metashale. This type of bedding character is most typical near the contact with Units 3 and 4A, but can be quite random.

Well preserved graded beds and cross beds give excellent control on topping direction in this unit. Unlike Unit 2 which is "fast graded" Unit 4 is "slow graded" or continuously graded. There is a smooth transition within each bed between metashale and metasandstone. As in Unit 2, the graded bedding is enhanced by large porphyroblasts of andalusite except in areas of low metamorphic grade where normal graded beds still persist. Near East Rochester, New Hampshire, some of the beds in this unit are "fast graded". These have well developed refracted cleavage only in the metashale.

The contact with Unit 4A is marked by the disappearance of calcsilicate rocks and the first appearance of shaley interbeds. When in contact with Unit 3, the transition between the two is marked by the abrupt loss of rusty weathering and appearance of non-rusty, variably bedded turbidite.

Unit 4 is sedimentologically similar to Unit 2. It may, however, represent a more distal turbidite deposited in a lower energy environment as suggested by the "slow grading" of beds. The proximity and depositional energy has decreased significantly in unit 4 with respect to unit 4A.

In summary, the stratigraphy presented here consists of five units, four of which are part of a stratigraphic succession. The fifth and lowermost, Unit 1, is a migmatite. Some changes in the local stratigraphy with respect to the older and better known Shapleigh Group have been made. Unit 1 is approximately equivalent to the Lower Rindgmere Formation. Unit 2 is a part of the Upper Rindgmere Formation. Unit 3 is equivalent to the Towow Formation. The Towow Formation was originally recognized in Maine and not across the state line in southestern New Hampshire (Hussey, 1968).

However, as shown in Figure 1A, we have mapped the Towow Formation or Unit 3 well into southeastern New Hamsphrie. Units 4 and 4A were previously mapped as either the Upper Rindgmere or Towow Formations.

STRUCTURE

The metasedimentary rocks are polydeformed. Three folding events are observed, all possibly Acadian or a combination of both Acadian and Alleghanian in age (Lyons and others, 1982). Table 2 explains the order and timing of both structural and metamorphic events.

TABLE 2

Timing	Deformation	Metamorphism	Comments
Based on Rangeley, ME fossils; Silurian to early Devonian	S_o, soft sediment deformation	Diagenesis	original bedding
Early Devonian?	F1, S1	M1?	fragmented andalusite porphyroblasts, nappe stage folding, inverted sections
Early Devonian based on Spaulding Qtz Diorite Rb/Sr age.	F2, S2	M1? M2	intrusion of diorite near Rochester, foliation, rolled porhyroblasts, NE-SW subhorizontal fold axes.
Middle to Late Devonian? or Carboniferous?	F3?	M3?	no cleavage, small scale bending of metasediments, secondary white mica and chlorite
Carboniferous?	F3? strike slip faulting?	M3?	small scale bending of metasediments, secondary white mica and chlorite, crenulation cleavage near fault trace.

F1 is characterized by large recumbent folds or nappes of uncertain vergence. The main schistosity (S1) is generally parallel to original bedding (So) and consequently is difficult to see. In places though, well developed S1 cleavage is beautifully preserved in what were probably parasitic F1 isoclinal folds on the limbs of the large folds. This cleavage and some of the associated early F1 isoclines are refolded about F2 folds axes.

The most convincing evidence of early recumbent folding is the recognition of extensive inverted sections within the stratigraphic sequence, much of which will be seen on this trip. The inverted limbs are refolded into open, asymmetric, downward facing antiformal synclines and synformal anticlines. This alone is evidence that an early folding event had to have occurred prior to refolding about F2 fold axes.

F2 folds refold large recumbent F1 folds about northeast to southwest gently plunging (up to 20°) axes. These folds are open, asymmetric, and verge to the southeast. Andalusite porphyroblasts are occassionally aligned parallel to F2 axes. Figure 5 is a plot of the F2 fold axes.

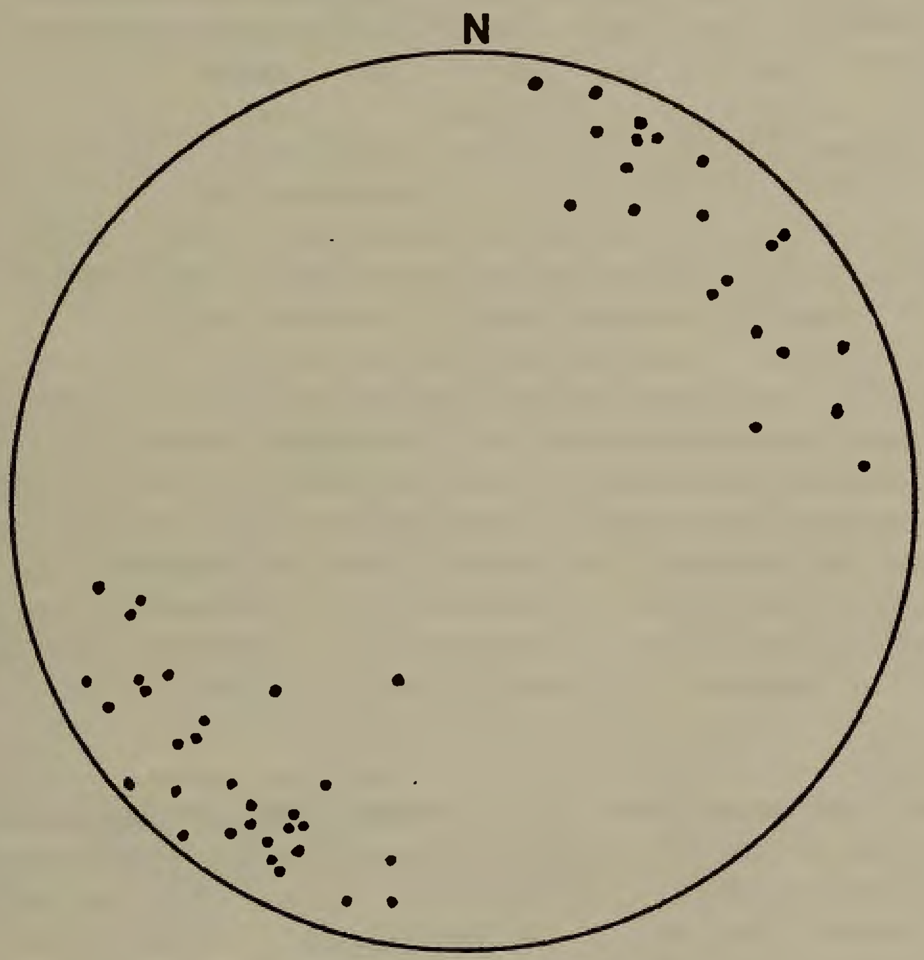

FIGURE 5 Equal area plot of F2 fold axes in the field trip area. Axes trend northeast-southwest with a shallow plunge in either direction

There is a well developed S2 axial plane cleavage. The biotite in the metasandstone of unit 4 is reoriented parallel to S2. Lineations resulting from S2 and S_o intersections are well preserved throughout the area.

The third folding event, F3, is defined by a change in attitude of the metasedimentary map pattern from east-northeast near Parker Mountain, New

Hampshire to north-northeast near West Lebanon, Maine. (see Figure 1A). No observable cleavage associated with F3 is seen in the field trip area.

On a smaller scale map (see compilation map, Figure 4) F3 is more pronounced. The northeast striking metasediments within the field trip area trend north and then northwest between a corridor defined by the Winnepesuake quartz diorite and the Sebago batholith. It is speculated that the change in regional strike may be a result of lateral "shouldering" by the emplacement of these igneous bodies or alternatively as a consequence of major strike slip faulting along the Nonesuch River/Campbell Hill/Hall Mountain fault system.

The Nonesuch River fault defines the southeastern boundary of the KCMS and may be part of a master fault system that includes the Norumbega fault in Maine and the Clinton-Newbury/Wekepeke/Flint Hill/Silver Lake faults in New Hampshire and Massachusetts (see Gaudette and others 1982, p. 1351 and Lyons and others 1982, p. 55-56 for details). A portion of the fault extends through the southeast corner of this area (see Figure 1A). It separates Unit 2 from the Precambrian (?) (Olszewski and others, 1984) Berwick Formation of the Merrimack Group. There is a distinct topographic lineament occupied by a series of deep ponds, abundant silicifed zones and minor slickensided surfaces ornamented by serpentine.

The Gonic Formation, a member of the Shapleigh Group (see correlation chart, Figure 2 and 3) is also separated by the fault from units within the KCMS. Although redefined by this study, the majority of the Shapleigh Group (Upper and Lower Rindgemere and Towow Formations) remain within the KCMS. We support the hypothesis proposed by Hussey (personal communication), that the Gonic Formation should be included within the Merrimack Group and agree with Bothner and others (1984) that it correlates to the Gove member of the Berwick Formation recognized near Raymond, New Hampshire.

The interpretation of the structure as presented above is different from that discussed by previous workers. Hussey (1962) recognized the Lebanon syncline, a major doubly plunging, upward facing syncline in the Berwick, Maine quadrangle. Stewart (1961) proposed a similiar structure for the metasedimentary rocks in adjacent New Hampshire. Gilman (1977) reported that overturned isoclinal folds are refolded into the major structures (Pequawket synform and Hiram antiform) seen in the Kezar Falls, Maine quadrangle.

We elaborate on Gilman's observations and suggest that large, early, F1 recumbent folds or nappes are refolded by open, inclined, F2 folds with west dipping axial surfaces. The inverted limbs of F1 folds are commonly refolded into downward facing F2 folds. Furthermore, we suggest that the Lebanon syncline is in fact better described as a downward facing antiformal syncline with Unit 4 and not Unit 3 or the Towow Formation preserved as the youngest unit in the fold core.

METAMORPHISM

Three events are recognized within the polymetamorphic rocks of the field trip area. (see Table 2). The early event, M1, is a progressive low pressure facies series regional event. In the metashales of all units, M1 reaches sillimanite + white mica grade and is characterized by well developed andalusite and staurolite in Units 4 and 2.

Locally abundant sillimanite and a coarsening of mineral grain size in the metasedimentary rocks near and at the contacts of igneous bodies is thought to represent a contact metamorphic event, herein called M2. It is difficult to distinguish between M1 and M2 in the field trip area. It is possible that these are not separate events, but represent a metamorphic continuum. We have not clearly established the timing of each event at this point. This uncertainty is expressed in Table 2.

The latest event, M3, is characterized by non-foliate decussate white mica and chlorite. The white mica is pseudomorphous after the alumnino-silicates, and chlorite pseudomorphs biotite and garnet. No foliation is associated with M3.

The increase in metamorphic grade to the northwest towards the "sillimanite plateau" of Thompson and Norton (1968), that is associated with M1, is beautifully preserved. The mineral assemblages listed in Table 1 for quartz + white mica bearing rocks rather tightly constrain the pressure-temperature path of M1 metamorphism (Figure 6). Andalusite is common indicating low pressure facies series metamorphism. Garnet + chlorite is not stable whereas staurolite + biotite, andalusite + biotite and sillimanite + biotite are. Therefore, the maximum temperature in all rocks must have been above reaction 1 in Figure 6. Staurolite is most commonly associated with andalusite + garnet + biotite and seldom with sillimanite + garnet + biotite. Consequently, the progressive metamorphic path must have passed close to the intersection of reactions 2 and 3 with the andalusite to sillimanite polymorphic transition (Figure 6). Sillimanite + potassium feldspar assemblages have not been recognized. The maximum observed grade is sillimanite + white mica in areas of abundant migmatites.

The position of the reaction boundaries in Figure 6 will change if X_{H2O} does not equal unity. The occurence of graphite and sulfides in equilibrium with the metashales (pelitic schists) indicate the presence of carbonaceous and sulfide components in the metamorphic fluid phase. Thus the activity of H_2O varied within the study area and may have affected mineral paragenesis (see discussion by Neilson, 1981; French, 1966; Thompson, 1972; Tyler and Ashworth, 1982 on similar rocks). Chamberlain and Lyons (1983) suggest that "graphitic pelites may contain higher-grade assemblages than adjacent non-graphitic pelites" (Chamberlain and Lyons, 1983, p. 536-537). Recent experimental work by Dutrow and Holdaway (1983) supports an extension of the staurolite stability field to lower pressures. These are some of the reasons why in Figure 6 we did not give values for the pressure and temperature axes. We attempt only to show the observed ensemble of mineral assemblages and important reactions and not to quantify the reaction boundaries.

The presence of randomly oriented biotite included in porphyroblastic garnet and andalusite suggest that these phases grew prior or during the development of S1. The prophyroblasts have been subsequently rotated during the development of S2 which is defined by aligned sheet silicates (see Figure 7). Garnet may have continued to grow after the development of S2. As shown in Figure 7, garnet cores contain abundant inclusions of quartz and biotite, followed by a zone of ilmenite. From this textural unconformity to

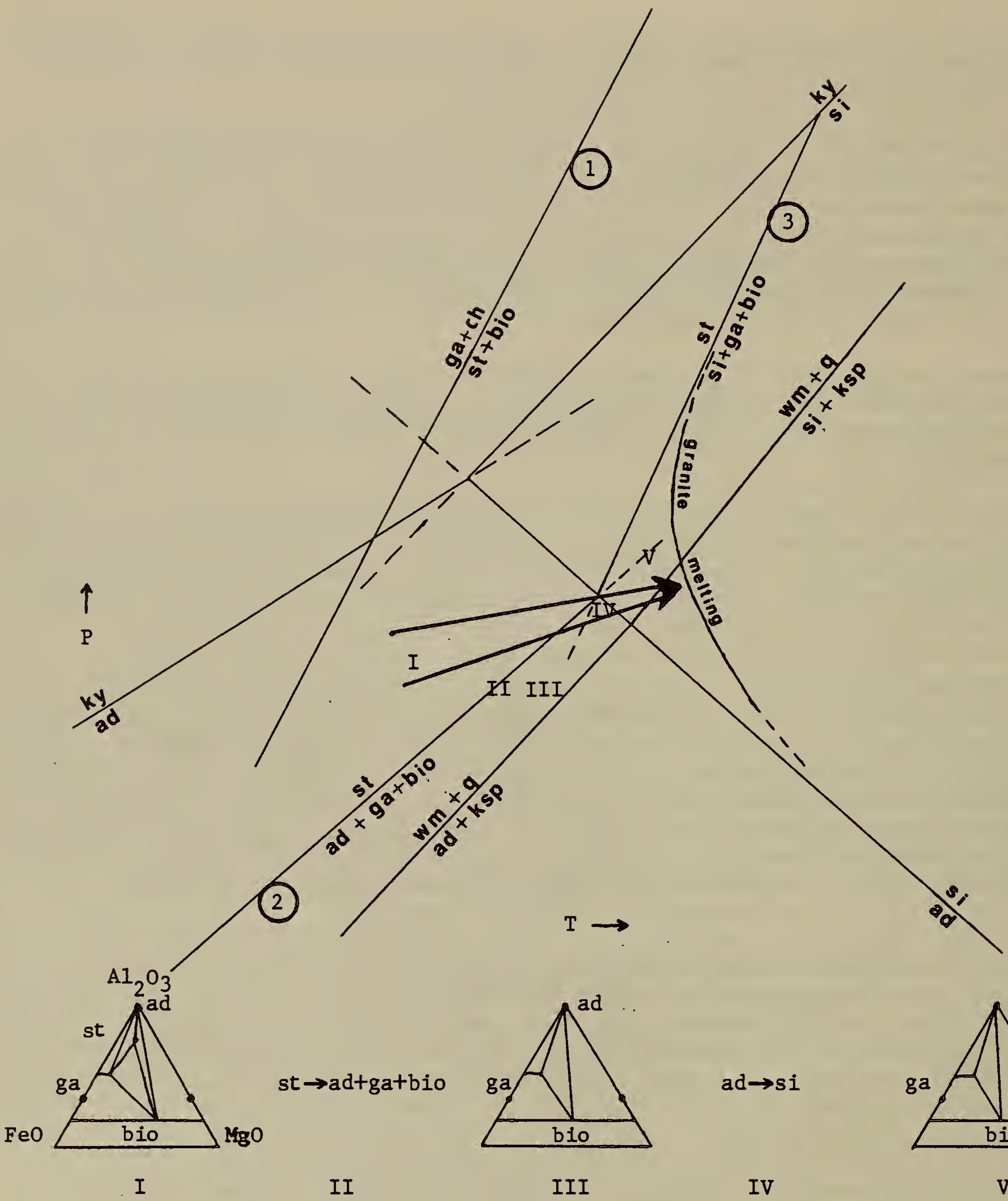

FIGURE 6 Path of low pressure facies series metamorphism associated with M1. Arrow defines path, Roman numerals refer to pertinent ensembles of assemblages or reactions observed in quartz + white mica metashales (pelitic schists) in the field trip area. Petrogenetic grid after Labotka (1981), granite melting curve (X_{H_2O} = 1) after Kerrick (1972) and Thompson projections showing pelitic assemblages after Thompson (1957). Quartz, white mica and H_2O are involved in the reactions as necessary. Abbreviations the same as in Table 1.

the rim, garnet is idioblastic and has grown over the biotite foliation (S2).

Andalusite is commonly attenuated, fragmented and aligned parallel to F2 fold axes. Elsewhere it is randomly oriented within pelitic layers as radiating clusters.

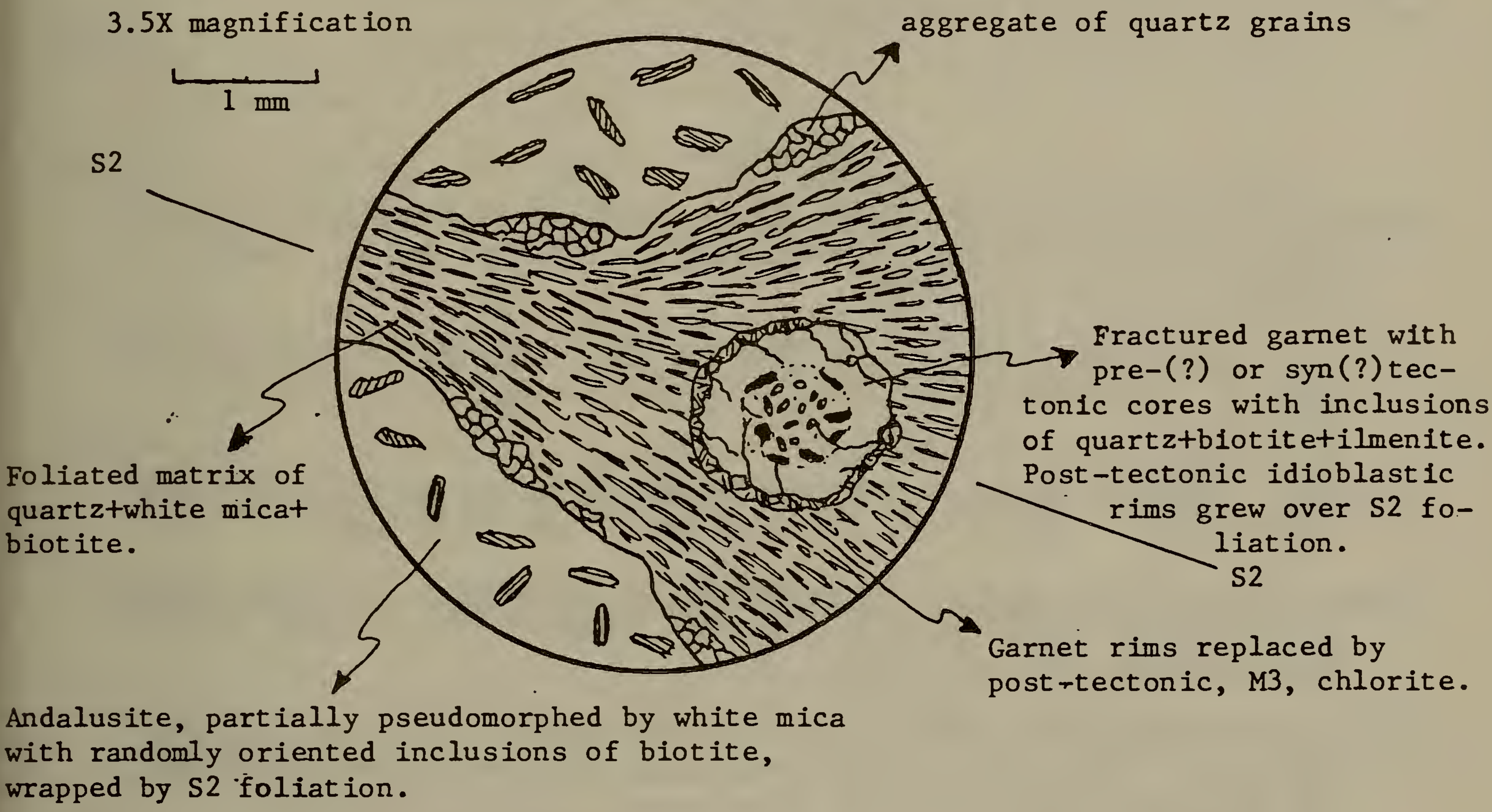

FIGURE 7 Textural sketch of thin section from Unit 4. Sample from East Rochester about .5 mile downstream along the Salmon Falls river from STOP 4.

Conclusions

A revised stratigraphy for the southeastern New Hampshire and southwestern Maine part of the KCMS has been presented. This stratigraphy was developed through detailed mapping of lithologic units with good control on succession from primary sedimentary structures. We have recognized Unit 4 above what was previously considered the top of the exposed section, Unit 3 or the Towow Formation, of the Shapleigh Group.

The remarkable similarity of both the order and lithology of this stratigraphy to the fossiliferous section near Rangeley, Maine suggests that they may be correlative. If so, the age of the metasedimentary rocks in southeastern New Hampshire and adjacent Maine is Siluro-Devonian. The possibility exists, however, that the units in the field trip area correlate to the Devonian Carrabassett and Seboomook Formations of Maine.

The Lebanon syncline, previously described as an upward facing, asymmetric syncline, is better described as a major downward facing, antiformal syncline that has formed by the refolding of large, early recumbent folds.

The staurolite + andalusite to sillimanite + white mica grade metasedimentary rocks of this part of the KCMS are truncated to the southeast by the Nonesuch River Fault. Juxtaposed along the fault trace is an exotic ensemble of metasedimentary rocks, the Merrimack Group, of at least pre-Silurian and probable Precambrian age.

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ROAD LOG FOR TRIP C5

Assembly Point: Rochester Mall parking lot, Ten Rod Road, Exit 14 off the Spaulding Turnpike, Rochester, NH [Berwick, ME 15' Quadrangle]. See Figure 1B for location.

Time: 9:00 a.m.

Mileage

0.0 Proceed N on Spaulding Turnpike

6.7 Take Exit 17, Milton exchange, turn around and proceed S on Spaulding Turnpike to STOP 1.

7.5 STOP 1

Unit 1 (Rangeley and Perry Mtn. Formations?)
Stay on S bound side of turnpike. Park well off breakdown lane on grass shoulder. Watch traffic. This applies to all STOPS on turnpike*

Unit 1, a migmatite, is exposed in a long outcrop. It is composed of swirly, randomly oriented, layers of intercalated quartz + plagioclase and biotite + garnet + sillimanite which are overgrown by secondary white mica. Elongate restite, or mini-mesosomes, are quite common in this outcrop. They are zoned with garnet + amphibole + quartz + plagioclase cores and biotite + quartz + plagioclase rims. The long axes of the restite are oriented NE-SW, approximately parallel to F2 fold axes and lineations.

9.2 STOP 2

Unit 2 (Perry Mtn. Formation?)

A tremendous outcrop of gray, thick bedded, metasandstone and metashale with discontinuous stringers of pink, garnet + quartz, coticules is exposed here. "Fast graded" beds are well preserved and are emphasized by the metshale layer which is entirely made of black weathering porphyroblasts of andalusite (andalumps). Keep an eye out for infrequent crossbeds (or is it early F1 cleavage that has been refolded?) and ripup clasts of metashale. Tops here suggest that the sediments face down. We are within the inverted limb of a F1 recumbent fold and on the NW limb of the major downward facing, F2, antiformal syncline as depicted in section A-A. (see Figure 1A of text).

Pumpellyian F2 folds are sparse here, however, a warp associated with F2 may be seen on the cliff face near the S end of the W outcrop. A beautiful, isoclinally folded, (F1?) garnet coticule can be viewed on top of the cliff face. Minor sillimanite in addition to garnet + biotite + andalusite (fresh to completely sericitized) are observed in the metashale. We are therefore close to Roman numeral IV of Figure 6 in the text.

9.6 ✓ STOP 3

Unit 2/Unit 3 contact (Perry Mtn and Small Falls Formations?)

Exposed in this long, but low outcrop is the transition, N to S, from non-rusty, gray, garnet coticule bearing Unit 2, whose bedding thickness has, as it typically does in this area, thinned up section considerably, to rusty weathering metashale and minor metasandstone of Unit 3. The contact is defined by the first appearance of rusty weathering. Because of this, the bedding style of Unit 3 is much the same as Unit 2 near the contact. However, if you walk S down the length of the outcrop a gradual increase in metashale is apparent in Unit 3.

Tops here indicate that the section is inverted. Unit 2, the older unit, structurally overlies Unit 3. We are in the same part of the structure as described for STOP 2. Numerous F2 downward facing minor folds, with fold axes trending SW and plunging about 15°, are seen in both Units. We are also at the same metamorphic grade as described for STOP 2.

Proceed S along turnpike.

10.0 Cross bridge #30.

11.8 Take Exit 16 to Routes 202, 11 and 16, east toward Rochester, NH and Sanford, ME.

12.1 At end of ramp take a left.

12.7 Cross bridge #26.

12.8 Bear right on Rtes. 202 and 11, heading east.

14.9 In E. Rochester center turn left at double flashing red light.

14.95 ✓ STOP 4

At end of road, park on right side. Walk down to the edge of the Salmon Falls river. Outcrop is just downstream from an old, rusty, iron beam bridge.

Unit 4A/4 contact (Madrid and Carrabassett Formations?)

Along the riverbank is exposed massive, white to gray weathering, purple on fresh surfaces, quartz + plagioclase + biotite metasandstone with biotite rich partings or laminations spaced about 2-6 cm apart. The calcareous portion of the metasandstone is evenly dessiminated with no observable calcsilicate lenses. Raised ridges, formerly fractures which have annealed , criss-cross the outcrop surface. There are no visible folds in this outcrop.

If the water level is low enough we can walk downstream to the contact with Unit 4 (Carrabasset Formation?). Unit 4 is composed of massive metashale with abundant well developed chiastolitic andalusite pseudomorphed by white mica. The contact is a few tens of yards downstream from the highway bridge. Although no topping information is seen at this STOP, just downstream, behind a factory in E. Rochester, are several outcrops of inverted Unit 4 (see location on cross section A-A' for relative structural position). Thin sections from behind the forementioned factory show pervasive M3 chlorite and white mica alteration of garnet, biotite and andalusite of the metashale in Unit 4.

Lunch Stop. Turn vehicles around.

15.0 Turn left onto 202 east at flashing red lights, proceed across bridge over Salmon Falls river into Maine.

15.1 Turn left on River Road, first left after bridge. Proceed NW along River Road.

15.9 STOP 5

After crest of small hill with outcrop on the right, park on the left in dirt lot.

Unit 3 (Smalls Falls Formation?)

Exposed here in the core of a minor, F2, downward facing synformal anticline (see cross section A-A' for location) is the rusty weathering, crumbly metashale of Unit 3. This is the Towow Formation as mapped by Hussey (1968) (see correlation chart, Figure 2). The major purpose of this STOP is to assess whether this is the same unit as exposed at STOP 3 or a different unit altogether. Based on stratigraphic position and lithology, two of us (JDE and WAB) feel that they are the same, one does not (AMH). In addition to the outcrop along the road, a large cliff outcrop is located about 100 yards due NE from the road, into the woods. Steeply inclined nearly recumbent, F2 folds with subhorizontal NE-SW trending fold axes are seen in both places.

Continue NW on River Road.

16.5 Turn left towards Kings Court Campground and Water Slide.

16.6 Turn left into the Water Slide parking lot. Park.

STOP 6

Unit 4/3 contact (Carrabassett and Smalls Falls Formations?)

In constructing the artificial hill for the water slide, the glacially polished bedrock was stripped of its Quaternary cover and the contact between Units 4 and 3 was exposed. Unit 4A does not appear here. It has presumably "pinched out" as indicated in Figure 1 of the text.

Exposed closest to the parking lot are lenses of grit composed of vitreous and blue quartz and lithic fragments from 1 to 5 mm in size in a quartz + biotite + white mica matrix that is moderately rusty. Proceeding towards the woods to the left of the slide are the rusty metashales more typical of Unit 3. Abruptly the rusty weathering stops and gives way to thick, well bedded metasandstone with cross bedding and minor metashale all completely metamorphosed into "andalumps". This is Unit 4. Subtle graded beds concur with topping information from the crossbeds that these beds are inverted. This outcrop establishes the presence of a non-rusty well bedded unit (Unit 4) above the rusty metashales, (Unit 3 or the Towow Formation of Hussey (1968)) which was not previously recognized within the Shapleigh Group (see correlation charts, figures 2 and 3).

As we have seen elsewhere, the older rock, Unit 3, is structurally on top of the younger one, Unit 4, in this inverted section. (see location on cross section A-A').

Turn left out of Water Slide parking lot and proceed across bridge, back into New Hampshire.

17.1 At intersection, go straight, passing Mike's Auto Body on your left.

17.6 Take a left on Rt. 16 heading S.

17.9 Bear right onto Rtes. 202, 11 and Spaulding Turnpike following sign for Conway and Portsmouth.

18.7 Take a right onto the Spaulding Turnpike, southbound, following sign for Concord and Dover.

19.3 Cross bridge #27.

20.0 Take Exit 14, Ten Rod Road.

Outcrop of Unit 3 on ramp. Go left at end of ramp onto Ten Rod Road. Now in Alton, NH 15' Quad.

21.7 Park in new dirt road that goes up a hill to the left.

STOP 7

Unit 4 (Carrabassett Formation?)

Exposed in low outcrops along the dirt road are inverted, shallow dipping, 4 to 10 cm thick, beds of metasandstone and metashale. Graded beds are well preserved. Andalusite, fresh to partially altered, elongate subparallel to F2 lineations is surrounded by small euhedral staurolite. The pelitic assemblage here is equivalent to that represented by Roman numeral I in Figure 6.

This outcrop is typical of the majority of Unit 4 as mapped in the field trip area. Without such things as the lack of garnet coticules and stratigraphic position with respect to the rusty weathering metashale, Unit 3, it would be difficult to distinguish this rock, Unit 4, from the garnet coticule bearing turbidite, Unit 2, lower in the section.

Continue up Ten Rod Road, heading NW.

21.9 Take a left on Four Rod Road

23.7 Take a left at the "T" intersection.

24.7 Take a right at the four way intersection.

24.9 Bear left on 202 A, heading WSW.

26.1 Outcrop of diorite near Rochester reservoir.

27.1 OPTIONAL STOP 7A (time permitting)

Park under power lines on right side of road. Outcrops are across the road under powerlines.

Unit 2 (Perry Mtn. Formation?)

Under the power lines are several pavement outcrops of thick bedded, 4-20 cm, garnet coticule bearing metasandstones and metashales of Unit 2. The grain size is fairly coarse here reflecting M2 contact metamorphism due to the intrusion of the nearby diorite. Relict andalusite? or staurolite? are overgrown by sillimanite as seen in the thin section here. This is also a result of M2 metamorphism overprinting M1 assemblages. ?

No definitive topping information has been recognized in this exposure. This is in part due to the high metamorphic grade and the "fast graded" character of each bed. We are still within the inverted limb of the early F1 recumbent fold (nappe, if you will) but are on the upright limb of an inclined, downward facing, F2 synformal anticline (see cross section B-B'). The diorite body probably exploited the weak axial region of this F2 fold. (see cross section B-B'). Minor F2 folds are seen in the outcrops as are probable F1 isoclinal folds in the garnet coticules and refolded S1 cleavage. I disagree

Continue W along Rt. 202A

30.0 Junction of Routes 202A and 126 in Center Strafford, NH

30.2 Continue on Route 126. Route 202A leaves to the left.

31.5 Swamp on right, a landmark to check mileage!

31.8 Take a right down a dirt road at a mail box labelled "E. McCarthy, P.O. Box 82".

32.6 STOP 8

Park on continuation of dirt road just beyond where it turns right toward remote residence. Walk about 1/4 mile down same road to brook crossing, follow brook downstream to gorge.

Unit 4/3 contact (Carrabassett and Smalls Falls Formations?)

The contact between the rusty weathering metashale of Unit 3 and the non-rusty thin bedded alternating metasandstone and metashale of Unit 4 is exposed parallel to and in places in the gorge. Specifically the contact is marked by the abrupt loss of rusty weathering color.

At the downstream end of the gorge in some glacial potholes are preserved graded beds and some questionable flames. Topping indicates downward facing sediments. F2 folds are nearly recumbent and also face down.

This STOP confirms the presence of a non-rusty, well bedded, turbidite, Unit 4, above the rusty metashales of Unit 2 as presented at STOP 6. In addition this demonstrates that the stratigraphy as presented earlier in the day along the Spaulding Turnpike and in adjacent Maine extends at least this far into New Hampshire.

Turn around and go back along dirt road to Route 126.

33.4 Take a right onto Route 126 heading NW.

34.2 STOP 9

Just below the top of the hill, park in small dirt parking lot the trailhead for Parker Mountain on the left side of Route 126.

Unit 4/3/2 contact (Carrabassett, Smalls Falls and Perry Mtn. Formations?)

With the exception of Unit 1 (the migmatite) and Unit 4A, the entire stratigraphy is exposed here. About 100 yards down the road from the parking lot is an outcrop of thin bedded Unit 4, quite similiar to that described at STOP 8. Up the road and to the top of the hill are exposed Units 4, 3 and 2 respectively in sequence. Based on graded beds in Unit 4, the sediments here are again inverted. We are very close to a granitic body as evidenced by abundant pegmatite intrusions that are injected throughout the outcrops.

The distinctive rusty weathering metashales of Unit 3 separates, well bedded, garnet coticule bearing Unit 2 from Unit 4 which is well bedded below the parking lot but tends to be a non-rusty massive metashale near the contact.

In the last outcrop of metasedimentary rock along the road on the NE side, at the top of the hill are isoclinally, folded, F1?, coticules. F2 folds are best preserved near and within the rusty metashales of Unit 3.

Depending on desire and time we may walk up the trail from the parking lot to a cliff outcrop of Unit 4 in contact with Unit 3, spotty pavement exposures of which are in the trail. From there we would proceed into the woods above the trail and view outcrops of Unit 2 with some downward facing F2 folds, similar to those seen earlier.

Turn around and retrace route to 202A and Center Strafford, NH.

36.5 Stop sign, go straight ahead.

36.7 Bear left on 202A towards Rochester.

40.2 Take a right on Pond Hill Road.

40.9 Pass Round Pond on left.

42.3 STOP 10 Fault contact between Unit 2 and Berwick Formation

Park at bottom of hill on shoulder.

Unit 2/Berwick Formation/Nonesuch River Fault.

We are right along the extension of the Nonesuch River Fault. This prominent topographic low forms a linear that marks the fault separating the Precambrian (?) Berwick Formation from the metashales and metasandstone of Unit 2 (the Silurian Perry Mtn. Formation?). Silicified zones and minor slickensided surfaces adorned with serpentine have been found about 1/2 mile in the woods to the NE on the NW side of Little Long Pond.

Outcrops of Berwick Formation, are composed of calcsilicate bearing quartz + biotite + plagioclase granofels. Unit 2 is composed of well bedded metashale and metasandstone with partially altered andalusite and staurolite. No garnet coticules have been found in this outcrop.

END OF TRIP

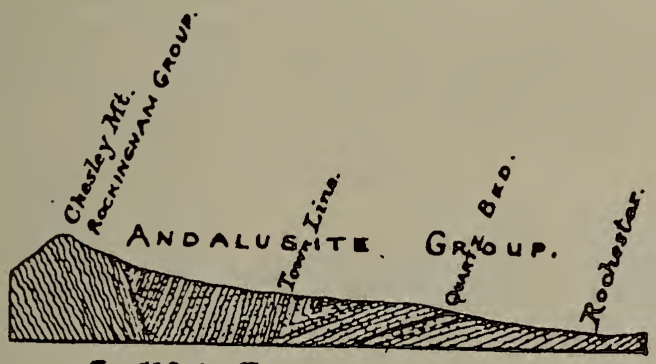

Fig. 110. In Farmington and Rochester.

C.H. Hitchcock, 1877
The Geology of New Hampshire
Volume II, 627

"In Fig. 110 is a section from Chesley mountain, Farmington to Rochester Village, lying partly in the Rockingham and partly in the andalusite group. The strata are much twisted, with a general southeast dip, and the schists contain beds of grainte".

GLACIOMARINE SEDIMENTS AND FACIES ASSOCIATIONS, SOUTHERN YORK COUNTY, MAINE

Geoffrey W. Smith
Department of Geological Sciences
Ohio University

INTRODUCTION

Late Wisconsinan ice retreat and marine submergence in southern coastal Maine is recorded in a complex succession of glacial and glaciomarine sediments. Surficial geologic mapping of these deposits has led to the development of a general stratigraphic framework for glaciogenic deposits throughout the coastal zone (Figure 1). The following general conclusions regarding ice retreat from coastal Maine can be drawn from this framework: (a) (during retreat (late Wisconsinan) ice was marine-based (e.g., ice was grounded below the prevailing sealevel; (b) the retreating ice sheet was warm-based and discharged large volumes of clastic sediment along the ice margin; (c) during retreat, the ice sheet was active (as opposed to passive/stagnant). General rapid retreat was interrupted by intervals of stillstand and minor readvance. (d) the character and pattern of ice retreat and deposition of glaciogenic sediments were controlled by a variety of factors including ice thickness, topography, and bathymetry.

Information gained from surficial geologic mapping has been supplemented by data acquired during the course of aquifer studies conducted by the Maine Geological Survey and the U.S. Geological Survey (Tolman, et al., 1983). This composite body of information permits the development of a provisional sedimentary facies model for glaciomarine and associated glacial and glaciofluvial deposits in the Maine coastal zone.

Original mapping of the Kennebunk quadrangle was conducted by Bloom (1960) at a reconnaissance level (1:62,500; 1:250,000). The area was remapped by J.T. Andrews (Maine Geological Survey Open-File Reports), and subsequently by G.W. Smith (Maine Geological Survey Open-File Reports) at scales of 1:24,000 and 1:62,500 in the early phases of the Maine Geological Survey's inventory mapping program. More detailed mapping of portions of the Kennebunk quadrangle was undertaken in the several stages of the Survey's aquifer mapping project (T. Brewer, M.G.S. Open-File Reports; A. Tolman, et. al., 1983). The surficial geology of the quadrangle is currently being revised and updated by G.W. Smith in conjunction with the Survey's detailed mapping program.

Information bearing on the stratigraphy and glacial geologic history of southern York County can be found in the following publications: Bloom, 1960, 1963; Smith, 1981, 1982, 1984, in press. Publications by Thompson (1978, 1982) and by Stuiver and Borns (1975) provide helpful general references to the surficial geology of the entire coastal zone.

GENERAL GEOLOGIC SETTING

Coastal Maine in late-glacial (Late Wisconsinan) time was covered by ice that extended to a terminal position on the continental shelf. Retreat of ice from its maximum position was underway between 17,000 B.P. and 15,000 B.P., and ice had withdrawn to the position of the present coastline in southern Maine by approximately 14,000 B.P. (Smith, in press). Stratigraphic evidence from the coastal region (Borns and Hughes, 1977; Thompson, 1982; Smith, 1981, 1982, in press) indicates that marine submergence of the isostatically depressed landscape was contemporaneous with ice withdrawal, and that ice retreat was accomplished, at least in part, by calving into the open sea.

Withdrawal of the marine-based ice in southern Maine appears to have taken place in shallow water (less than 10 m). The ice margin remained in the position of the present coastline until approximately 13,200 B.P., possibly due to the change in ice regimen that accompanied withdrawal from the deeper water of the Gulf of Maine. Subsequent retreat took place rapidly, so that ice had reached a position above the marine limit along its entire length between 12,600 B.P. and 12,400 B.P. The late-glacial marine transgression reached its maximum extent at this time. Emergence of the southern coastal zone, resulting from isostatic recovery, was complete by 11,500 B.P.

The general stratigraphy of the Maine coastal zone (Figure 1) has been described in earlier publications (Smith, 1982, in press). Subsequent work has led to the refinement of that stratigraphy, primarily in terms of the recognition of a wider variety of lithofacies (Table 1).

GLACIAL GEOLOGY OF THE KENNEBUNK QUADRANGLE

The glacial geology of the Kennebunk area is illustrated on the draft copy of the Kennebunk 15-minute quadrangle (south half) to be distributed at the start of the field trip.

Glacial till (Qt, Qwts), predominantly subglacial lodgement till, occurs throughout the area mapped. It is exposed as a blanket deposit of variable thickness over bedrock highs, and is inferred to underlie younger deposits in the valleys.

Ice-contact stratified drift (Qic, Qicd, Qe, Qem) has been mapped in a broad zone over the central portion of the quadrangle. These deposits occur primarily as a variety of end moraines and deltas (partial to fully-developed forms), that can be traced northwestward into the foothills of the White Mountains where they give way to esker sediments and valley trains. Of particular importance among this group of deposits are Merriland Ridge, Bragdon Road delta, and Perkins Town (L Pond) delta. These features are a succession of deltas constructed to a constant sealevel (approx. 220 ft. above present sealevel). Their occurrence relative to shallow bedrock provides constraints on efforts to reconstruct water depth during deglaciation of this part of the coastal zone. In addition, Merriland Ridge has been considered (Katz and Keith, 1917) to be a portion of the Newington Moraine of New Hampshire and southern Maine.

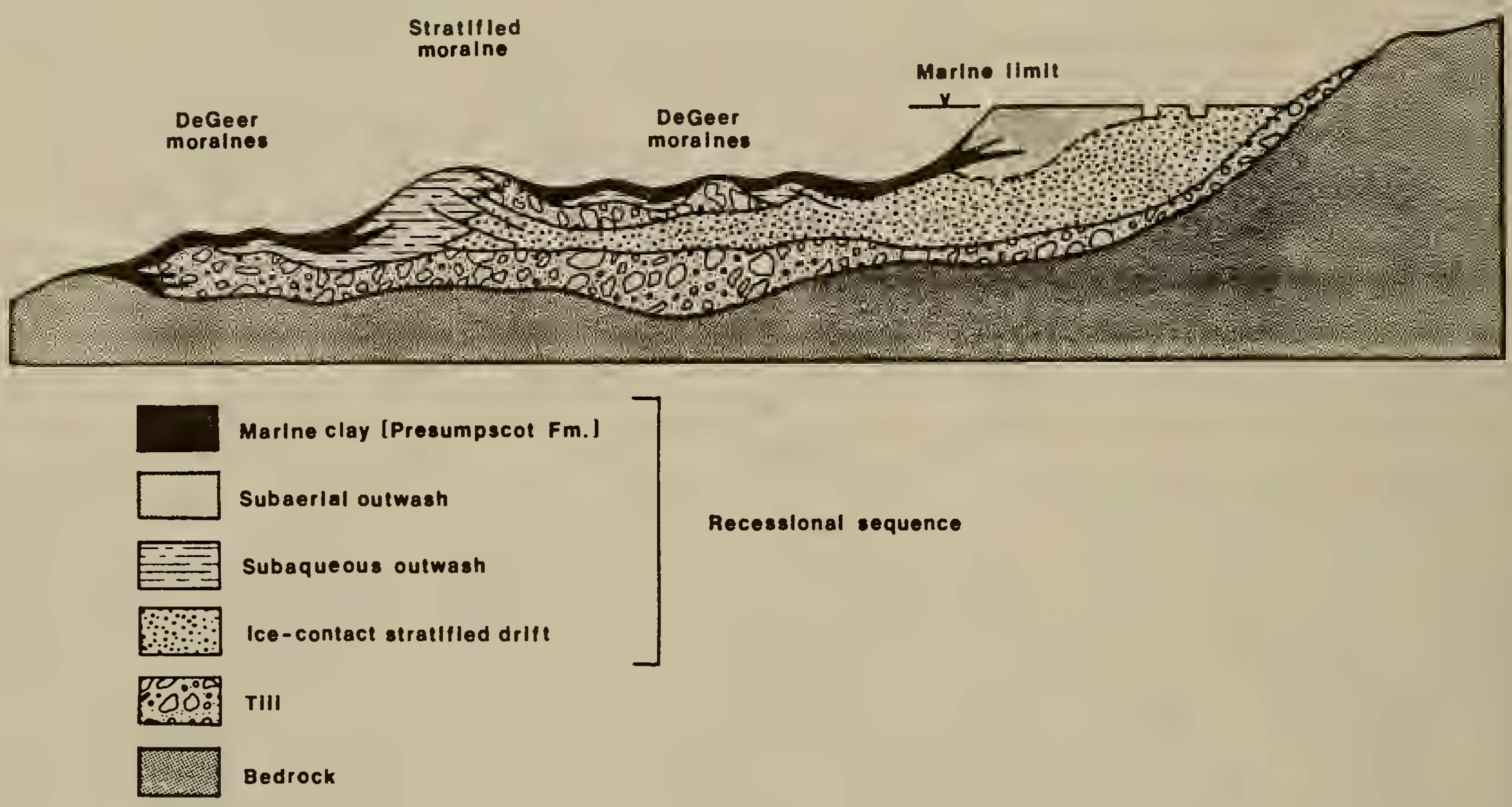

Figure 1. Generalized glacial stratigraphy of the Maine coastal zone.

Glaciomarine sediments (Qmc, Qms, Qmu), the Presumpscot Formation of Bloom (1960, 1963), occur as the youngest glacial deposits over large portions of the central and southwestern areas of the map. The marine clay (and silt) is the type Presumpscot Formation defined by Bloom. It underlies in gradational contact the sandy facies of the marine sediments, which are considered to be simply a regressive phase of the Presumpscot Formation.

Glacial outwash (Qow) comprises a thick accumulation of late-glacial sediment over the northern portion of the map. This sediment is part of a large delta that extends nearly to the northern margin of the quadrangle, and heads at a series of gaps in the uplands in the vicinity of Sanford and Alfred.

The episode of marine regression is recorded in several prominent wave-cut escarpments that range in elevation from approximately 40 feet (13m) to 220 feet (73m) above present sealevel. A well-developed beach ridge, upon which U.S. Route 1 is constructed between Ogunquit and Cozy Corners, records the position of falling sealevel at 60 feet (20m).

Holocene sediments are most common in the vicinity of the present coastline where they occur as tidal marsh (Qtm), beach (Qb), and tidal flat (Qtf) deposits. Inland, Holocene sediments include swamps (Qs) and scattered occurrences of floodplain alluvium (Qa).

Important glaciomarine deposits not exposed as surface materials include subaqueous outwash, ice-frontal debris flows, and bedded tills. These materials are, however, exposed locally and will be examined and discussed during the course of the trip.

Table I. GLACIOMARINE LITHOFACIES of the Maine coastal zone.

ORGANIC MUD (and SAND) WELL-SORTED STRATIFIED SAND	REGRESSIVE MARINE (SHOALING/TIDAL/BEACH)
MASSIVE/LAMINATED SILT and CLAY	DISTAL (QUIET WATER) MARINE
LAMINATED SILT and SAND	DISTAL SUBAQUEOUS OUTWASH
SORTED and STRATIFIED SAND and GRAVEL	PROXIMAL SUBAQUEOUS OUTWASH
POORLY SORTED and STRATIFIED GRAVEL	ICE-FRONTAL DEBRIS FLOW
DIAMICTON (TILL): BEDDED MASSIVE	ICE-FRONTAL/SUBGLACIAL (FLOW/LODGEMENT/MELTOUT)

DEPOSITIONAL PROCESSES ASSOCIATED WITH DEGLACIATION

Within the glaciomarine setting that persisted as ice withdrew across the coastal zone, sediments were deposited by a variety of complexly interrelated processes. The general nature of these processes can be inferred from the sediments themselves and by analogy to processes described by other workers in other areas.

During ice retreat and coastal submergence, sediments accumulated under two general depositional regimes: (1) ICE-DOMINATED, and (2) WATER-DOMINATED. Interestingly, the overwhelming volume of sediments appears to have been deposited by (melt)water-dominated processes. Under both regimes, deposition was influenced by the juxtaposition of glacial ice and meltwater (ACTIVE) and standing marine water (PASSIVE).

In the ICE-DOMINATED environment, general depositional processes included (Figure 2):

a. Subglacial lodgement,
b. Subglacial (and englacial) meltout,
c. Brash deposition at the ice front,
d. Debris flow, and
e. Iceberg rafting and brash deposition.

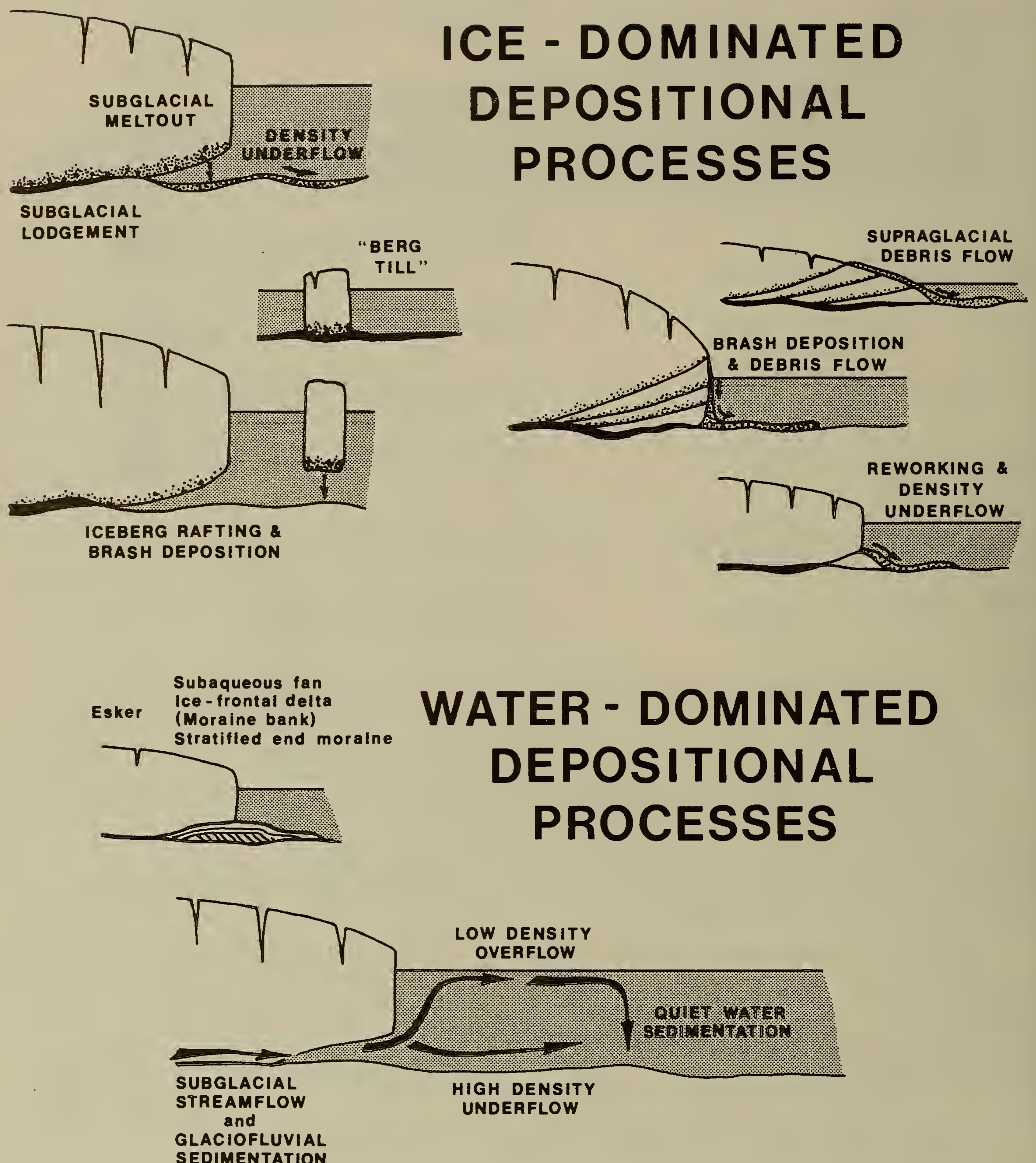

Figure 2. Depositional processes associated with retreat of a marine-based ice sheet.

Several factors influenced the relative contributions of each of these processes. Among these factors were: (a) bathymetry, (b) configuration of the ice front, (c) rate and volume of sediment influx, (d) rate and nature of ice retreat, and (e) occurrence and distribution of ice-frontal features.

In the WATER-DOMINATED environment, depositional processes included:

a. Subglacial streamflow (and glaciofluvial sedimentation),
b. Ice-frontal overflow and interflow and quiet water deposition, and
c. Ice-frontal density underflow.

The factors influencing the relative importance of each process were much the same as those for the ice-dominated environment. In addition, the processes operating in each environment overlapped, and relative roles of each varied in accord with these same factors.

In late stages of marine submergence and marine regression, previously deposited sediments were modified by beach and coastal processes.

GLACIOMARINE FACIES ASSOCIATIONS

On the basis of stratigraphic relationships and the areal distribution of glaciomarine lithofacies (Table 1), the following general facies associations are proposed for the Maine coastal zone (Figure 3).

1. Within the SUBGLACIAL environment, BASAL LODGEMENT TILL (and SUBGLACIAL MELTOUT TILL) record deposition in the ice-dominated regime. ESKERS are deposited in the (melt)water-dominated regime.

2. The ICE-FRONTAL/MARINE PROXIMAL association records the influence of the greatest variety of depositional processes and the most complex interrelationship of lithofacies.

   Bedded tills (MELTOUT TILL and FLOW TILL) are deposited predominantly in a near-ice setting. MELTOUT TILLS may give rise to DENSITY UNDERFLOWS, and FLOW TILLS may travel as SUBAQUEOUS DEBRIS LOBES for some considerable distance away from the glacier margin.

   POORLY SORTED and STRATIFIED GRAVEL may originate in the ice-dominated regime, but it more likely records deposition by stream-generated HIGH DENSITY UNDERFLOW (or DEBRIS FLOW).

   SORTED and STRATIFIED SAND and GRAVEL is the proximal facies of meltwater-transported SUBAQUEOUS OUTWASH. The distal facies of this outwash is LAMINATED SILT and SAND.

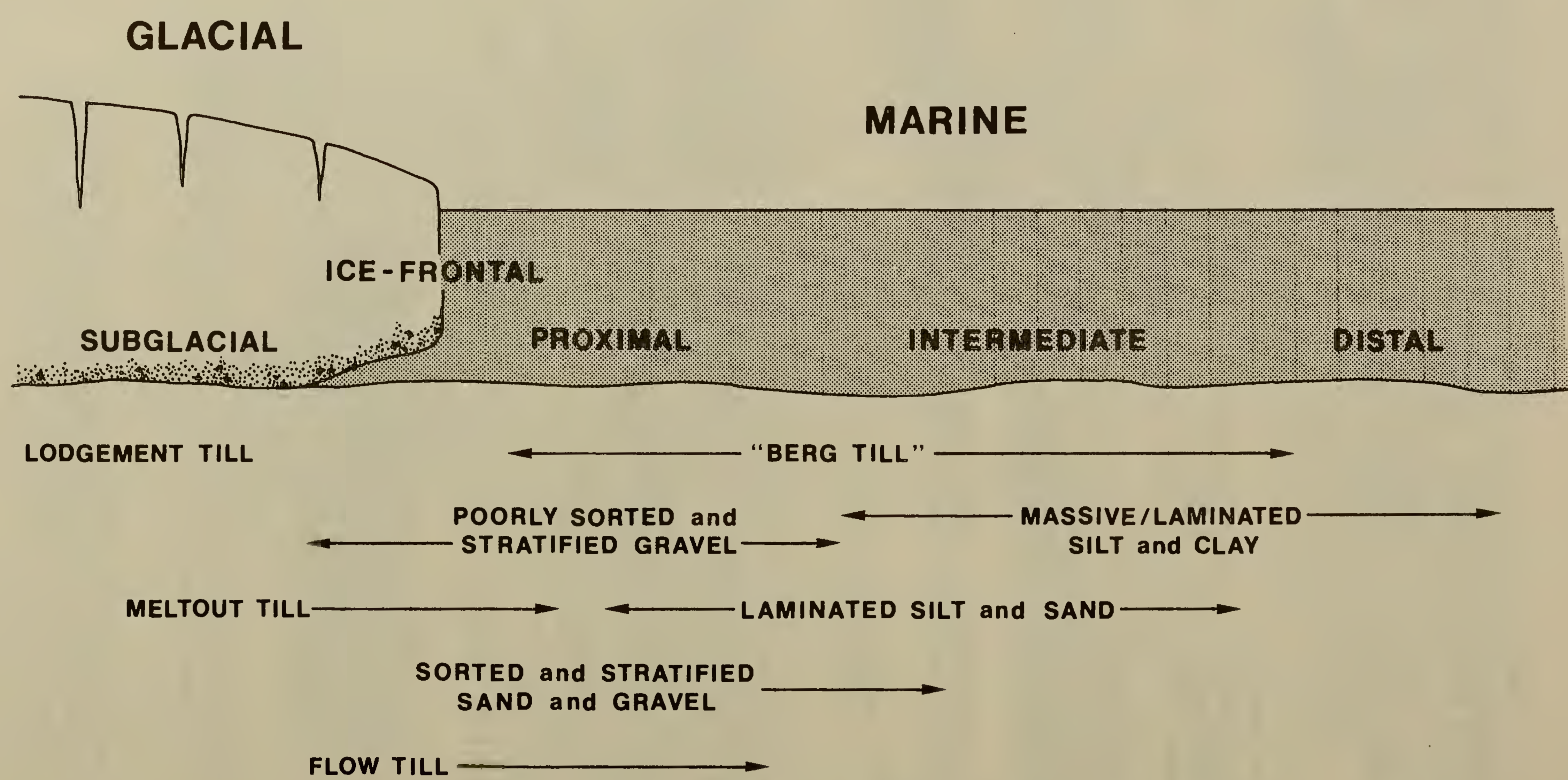

Figure 3. Glaciomarine facies associations.

A variety of important morphologic features are formed in the ICE-FRONTAL/MARINE PROXIMAL environment. These include: (a) DEGEER MORAINES, (b) "STRATIFIED" MORAINES (MORAINE BANKS), and (c) PARTIAL to FULLY-DEVELOPED DELTAS. These features incorporate a variety of the lithofacies described above.

3. Quiet water sedimentation in the DISTAL MARINE setting is characterized by MASSIVE/LAMINATED SILT and CLAY derived from LOW DENSITY OVERFLOW.

4. The REGRESSIVE MARINE association (not figured) includes WELL-SORTED and STRATIFIED SAND (shoaling marine) and ORGANIC MUD (tidal flat).

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ROADLOG

Mileage	
0.0	Field trip begins at Maine Turnpike (Interstate 95) toll gate 3 (NORTHBOUND - Kennebunk exit) at 8:30 am. Field trip materials will be distributed to participants at this time.
	Leave toll gate, heading south on State Route 35 toward Kennebunk. Route crosses distal portion of large outwash delta that heads to the northwest in the vicinity of Sanford and Alfred and extends to the coast at Kennebunk Beach. Local bedrock and till hills, covered with a thin veneer of outwash sediments, provide relief to the generally flat delta surface.
1.6	Junction of Routes 35 and 1. Turn left on Route 1 and proceed 0.1 mile to stoplight. Bear right on Routes 35 and 9A.
2.2	Overpass of Boston and Maine Railroad.
2.4	Kennebunk Beach Road. Turn right (south) Route continues on surface of Sanford Delta. In 0.7 mile, pass Kennebunk landfill site on left (Stop 2).
4.9	Junction with State Route 9 at Four Corners. Continue straight ahead (south) on Kennebunk Beach Road.
5.5	Great Hill Road. Turn right.

5.8 STOP I LIBBY'S POINT TILL EXPOSURE

Parking is a problem at this stop. Find places to park along the berm or in vacant driveways. We will walk from this stop to Stop Ia (Great Hill).

Examination of this outcrop will be dependent upon tidal conditions. Access to the exposure is a bit tricky at high tide. The till exposed at this locality is quite different from that seen at Great Hill (visible to your right as you face the ocean). It is, however, apart from color, similar to the surface till mapped throughout the coastal zone. The till is a distinct olive brown color. It is compact, with a well-developed fissility, and a strong fabric (N47W) that is generally the same as the trend of striations and grooves on subjacent bedrock outcrops (N48-52W). The till matrix is sandier than that of the till exposed at Great Hill, and the clast assemblage comprises a broad range of lithologies, unlike the till at Great Hill.

The till at Libby's Point is considered to be a SUBGLACIAL LODGEMENT TILL deposited by Late Wisconsinan ice during advance over this part of the coastal zone. This interpretation is, however, certainly open to discussion.

Following sufficient examination of the exposure and adequate discussion of its significance, proceed west for 0.3 mile to Great Hill (Stop Ia).

STOP Ia GREAT HILL

Great Hill is a wave-eroded hill of till and stratified drift at the west end of Kennebunk Beach. The exposure was described by Bloom (1960) in the context of evidence for the Kennebunk glacial advance. Bloom noted the occurrence of natural molds of Yoldia arctica in lenses and tongues of gray silty clay incorporated into till (the basal unit exposed here). He considered the till to be derived in part from reworking of the marine clay (Presumpscot Formation), and therefore to record advance of ice into the sea.

Andrews (unpublished data) described the section at Great Hill in terms of a hill of Presumpscot Formation overlain by gravel and sand. He collected shell material (broken valves and whole shells in growth position) from an horizon at the top of the "marine clay" just below the contact with the "gravel". A date of 13,830 ± 100 B.P. (QL-192) on this shell material indicates that ice had withdrawn to the position of the present coast in southern Maine by approximately 14,000 B.P.

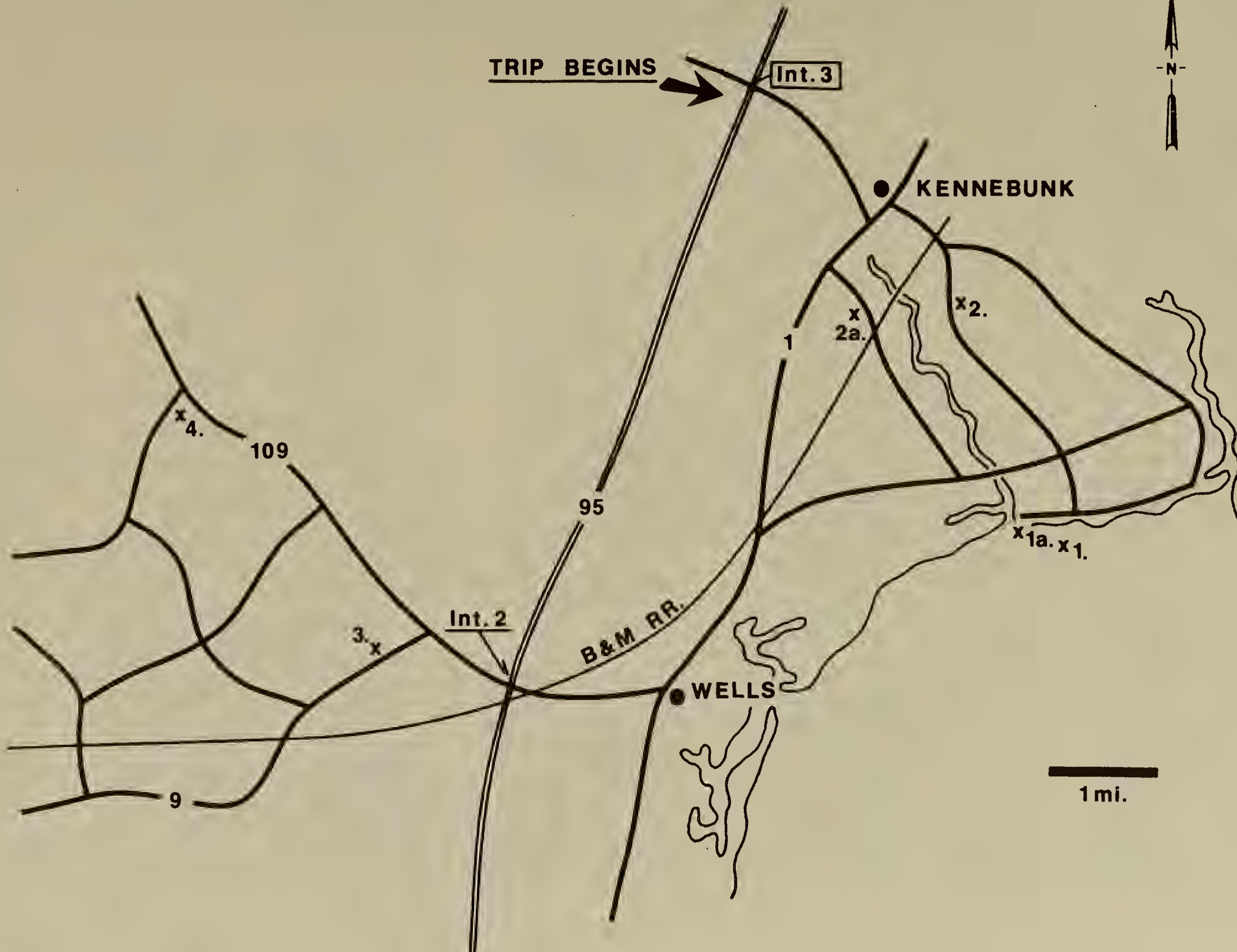

Figure 4. Road map of Kennebunk/Wells area, showing location of field trip stops (numbers).

The basal unit at Great Hill is a dark gray, silt-rich non-fissile sediment, consisting predominantly of local (Kittery Formation) clasts, and displaying a moderate to weak fabric of N40W. Throughout the unit, till (water-lain?) is interbedded with thin layers and lenses of silt and fine sand. The lower portion of the unit shows a conspicuous large-scale stratification.

Overlying the basal unit, in sharp but conformable contact, is a variable thickness (0-2m) of deformed fossiliferous marine silt and fine sand (Presumpscot Formation?).

An erosion surface with as much as 1m of relief separates the marine sediments from the overlying unit, a gray brown gravelly sediment with a high clast/matrix ratio and a clast composition that becomes measurably more cosmopolitan (igneous, metamorphic) from bottom to top of the unit. The clast fabric is moderate to strong (N30-35W), normal to the axes of DeGeer moraines that occur to the north of Great Hill.

The section is capped by a thin (less than 1m) layer of moderately well-sorted fine to coarse sand and pea gravel.

The sediments exposed at Great Hill are considered to have been deposited during retreat of the marine-based Late Wisconsinan ice sheet from southern coastal Maine. The basal unit, a BEDDED DIAMICTON (TILL), was deposited in an ICE-FRONTAL/MARINE PROXIMAL setting (see Figure 3). The overlying LAMINATED SILT and SAND records withdrawal of the ice and accumulation of PROXIMAL SUBAQUEOUS OUTWASH in an INTERMEDIATE or DISTAL MARINE setting. The upper POORLY-SORTED and STRATIFIED GRAVEL was probably deposited in the ICE-FRONTAL/MARINE PROXIMAL environment by HIGH DENSITY UNDERFLOW (DEBRIS FLOW). The uppermost unit here is a thin REGRESSIVE MARINE sand.

Return to cars. Turn around and retrace route to Kennebunk landfill site.

8.5 STOP 2 KENNEBUNK LANDFILL SITE

Park cars along right-hand berm or in entrance to the landfill site.

The geology of the Kennebunk landfill site (originally the Kennebunk town gravel pit) was first discussed by Bloom (1960). At the time of Bloom's work (late 1950's), exposures in the northeast part of the pit provided what Bloom considered to be the strongest evidence for what has come to be called the Kennebunk glacial advance. The section described by Bloom at this locality consisted of deformed sediments of the Presumpscot Formation in contact with till and overlain by poorly-sorted gravel and well-sorted fine sand. Deformation of the marine sediments was thought by Bloom to have been caused by "glacier ice pushing east and depositing till against the ice-pushed marine sediments." The sediments overlying the Presumpscot Formation were thought to have been deposited as outwash and proglacial deltaic sediments into the late-glacial sea. The exposure described by Bloom is no longer available for study. However, the units described by Bloom, and their stratigraphic relationships, are still visible in various parts of the pit.

Shells collected from the Presumpscot Formation at the landfill site (Stuiver and Borns, 1975) have been dated at 13,200 ± 120 B.P. (Y-2208). As at Great Hill, these shells include both broken shells and articulated valves in growth position. The date of 13,200 B.P. provides a maximum date for local readvance of ice at this locality during the general period of ice retreat from the coastal zone.

During the time since Bloom worked on the geology of this site, excavation of the southern part of the pit has exposed a variety of sediments and depositional features that bear on the interpretation of the glacial history of this part of the coastal region. Important among these are the sediments that comprise a series of small (3 to 4 m high, 10 to 17 m wide) ridges that trend ENE-WSW beneath a cover of well-sorted, and often deformed, sand (Smith, 1981, copies of which will be available on the trip). The composition of the ridges varies from poorly-sorted gravel to combinations of till, gravel, and sand. In general, however, most ridges consist of a core of poorly-sorted gravel overlain by massive till and interbedded coarse gravel and sand. The gravel and sand thicken between adjacent ridges, and are commonly deformed by small-scale shearing (sense of movement from the northwest). In some instances, the overlying sand contains thin lenses of gray silt that is also deformed.

The sedimentary succession exposed at the landfill site has been reinterpreted as follows (with, again, plenty of room for discussion):

1. The Presumpscot Formation, described by Bloom and underlying (but not exposed) in the southern part of the pit, is the MASSIVE SILT and CLAY lithofacies deposited in a DISTAL MARINE environment.
2. The overlying (and in part interbedded) POORLY-SORTED and STRATIFIED GRAVEL is an ICE-FRONTAL DEBRIS FLOW deposit, recording local readvance of the ice margin at this locality.
3. The sand exposed in this pit is in some places the distal portion of the Sanford outwash delta (DISTAL

SUBAQUEOUS OUTWASH?). Elsewhere, and more commonly, the sand is interbedded with gravel (SORTED and STRATIFIED SAND and GRAVEL) or silt (LAMINATED SILT and SAND). These sediments accumulated as INTERMEDIATE or DISTAL SUBAQUEOUS OUTWASH during withdrawal of ice to the north of the landfill site.

4. The small ridges are DEGEER MORAINES that were formed at the ICE FRONT, mostly by ice shove. They are composed of a variety of ICE-FRONTAL lithofacies, including BEDDED and MASSIVE DIAMICTON, POORLY-SORTED and STRATIFIED GRAVEL, and SORTED and STRATIFIED SAND and GRAVEL.

Return to cars. Turn around, and follow Kennebunk Beach Road south to Four Corners.

10.3 Junction with State Route 9. Turn right (west)

10.9 Cross Mousam River.

11.1 Brown Road. Turn right (north)

13.2 STOP 2a BROWN ROAD BORROW PIT (Optional)

Turn left onto dirt access road to borrow pit. Park at top of road that leads down into pit.

Shallow exposures along the road leading into the pit display the following succession of sediments (bottom to top of section):

1. GRAVEL oxidized; moderately well-sorted; generally fining upward from boulder/cobble to pebble/cobble with sand matrix; no apparent bedding or structure; 3 to 5 meters exposed; abrupt contact with...
2. SILTY CLAY gray brown; interbedded with fine mica sand; upper portion oxidized; 0.5 meter; gradational contact with overlying...
3. SILTY CLAY strongly oxidized (red brown to brown); cobbles and pebbles common; 0.5 meter; sharp contact with...
4. CLAY/SILTY CLAY blue gray; plastic; contains molds and casts of shells (no shell material); upper portion contains cobbles; small pebbles throughout; 1 meter; abrupt contact with...
5. TILL gray brown (oxidized); contains predominantly round to subround clasts in silty/fine sand matrix; 0.5 meter; sharp contact with...
6. SAND and SILTY CLAY alternating gray, blue gray, and yellow-red brown; becomes progressively sandier toward top; 1 to 2 meters; sharp contact with...
7. SAND medium to coarse at base, fining upward to bedded and cross-bedded medium to fine mica sand; locally reworked to form dunes; well-developed

podzolic soil (Spodosol).

The sediments exposed here record much the same sequence of events as discussed at Stop 2. The basal gravel (unit 1) is not well-exposed, but is considered to be an ICE-FRONTAL/MARINE PROXIMAL deposit of either PROXIMAL SUBAQUEOUS OUTWASH or DEBRIS FLOW. Units 2 through 4 include both MASSIVE/LAMINATED SILT and CLAY and LAMINATED SILT and SAND, both deposits of the DISTAL MARINE environment. The presence of abundant cobbles and pebbles records the proximity of ice and the influence of BRASH DEPOSITION from ICEBERGS. Unit 5 (Till) probably records local readvance of ice and deposition of SUBGLACIAL LODGEMENT TILL. The overlying sand and silty clay (unit 6) are considered to be DISTAL SUBAQUEOUS OUTWASH deposited as ice withdrew again from the area. Unit 7 is a REGRESSIVE MARINE deposit recording shoaling of marine water during coastal emergence.

NOTE: The stratigraphy of the first four stops (Stops 1, 1a, 2, and 2a) is summarized in Figure 5, and can be reviewed and discussed at lunch.

Return to Brown Road. Turn left (north), and continue to Kennebunk.

13.8 Junction with U.S. Route 1 at Kennebunk. Turn left (south), heading toward Wells. This will probably be the best place to have LUNCH (specific arrangements will be made as the trip progresses).

16.6 Junction with State Route 9 at Cozy Corners. At this point, the road rises onto the Wells beach ridge, a subdued linear ridge of sand that can be traced morphologically from this point to the town of Ogunquit. The crest of the ridge stands at 60+ feet (approx. 20m) above present sealevel. To the west (right) till and bedrock highs are covered with a thin veneer of sand, produced in large part by wave-reworking during coastal emergence. To the east, a prominent wave-cut escarpment drops to the level of modern tidal marsh and beach deposits.

18.4 Junction with State Routes 9&109 at Wells. Turn right (west). Route leaves beach ridge and crosses area of regressive marine sand before rising onto till and bedrock.

20.6 Junction of State Route 9 and State Route 109. Turn left (south) on Route 9. Just ahead and to your left as you make this turn is Merriland Ridge. Route 9 is constructed here on an apron of sand that spreads southeastward from the base of a wave-cut escarpment, visible behind the houses to your right.

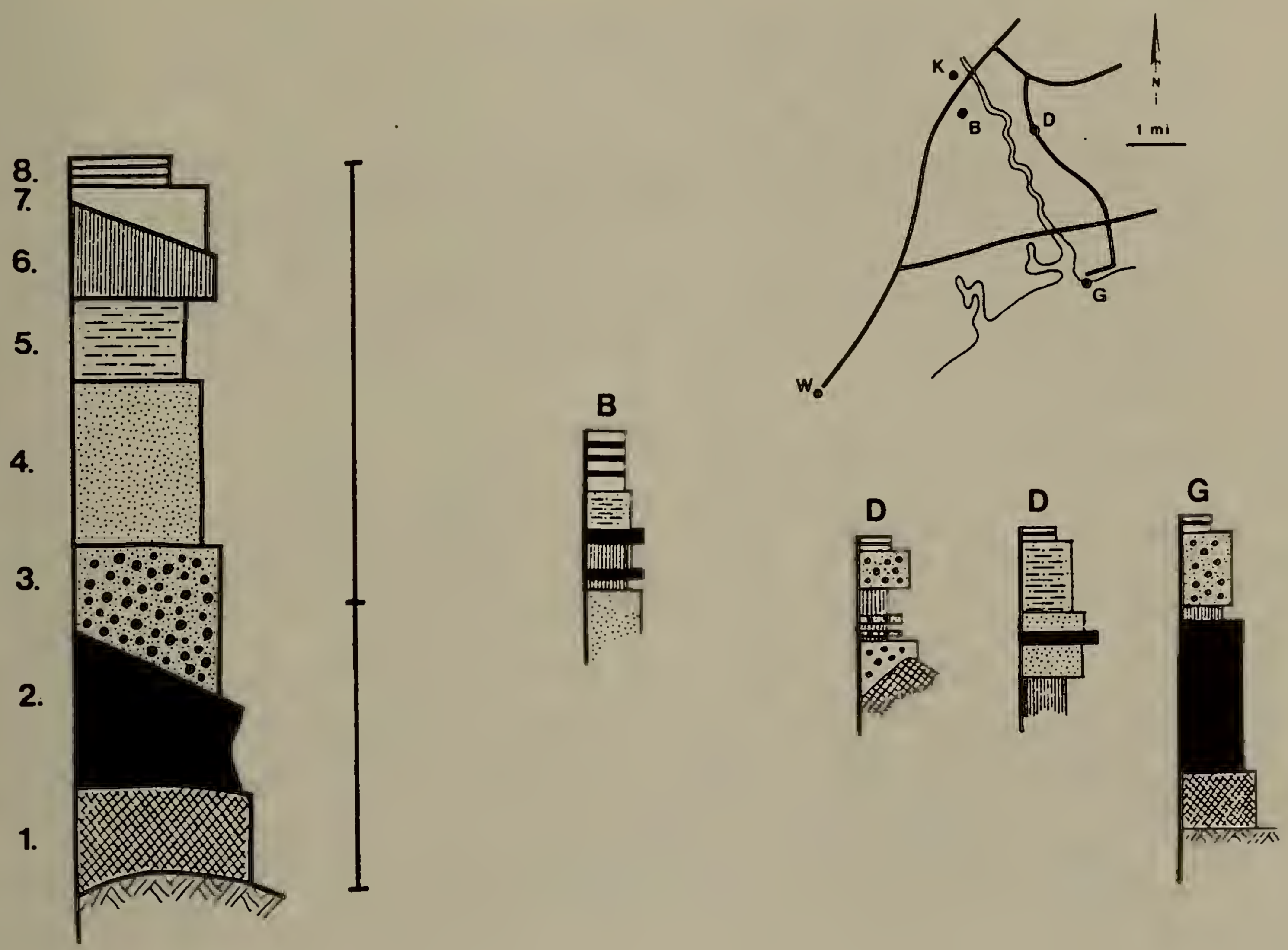

Figure 5. Summary of stratigraphy seen at Stops 1, 1a, 2, and 2a. Explanation: K - Kennebunk, B - Brown's Road, D - Kennebunk landfill, G - Great Hill, W - Wells. 1 - Massive diamicton, 2 - Bedded diamicton, 3 - Poorly-sorted and stratified gravel, 4 - Sorted and stratified sand and gravel, 5 - Laminated silt and sand, 6 - Massive/laminated silt and clay, 7 - Organic mud, 8 - Well-sorted stratified sand.

21.2 STOP 3 MERRILAND RIDGE

We will make a brief stop here to view the morphology of the distal portion of the ridge, and to discuss the significance of Merriland Ridge in the history of deglaciation of southern coastal Maine. Other stops here will depend upon the availability of good exposures.

Merriland Ridge is a narrow (200-450m wide) ridge of sand and gravel, 7 to 8m high, and approximately 7 km long, in the town of Wells. The general character of the ridge was described by Bloom (1960) as follows: "The basic parts of Merriland Ridge are, therefore, bedrock at each end and near southeastern salients of the surficial cover, three elongate hills of ice-contact stratified drift, two of which are connected by a saddle also of ice-contact stratified drift, and two ridges of sand, one across the railroad cut and the other extending west to bedrock at the western end of the ridge."

Exposures in the several gravel pits that line the crest of the ridge reveal steeply-dipping sand and gravel foresets (and backsets), in many instances capped by 1-2m of coarse gravel topsets. The ridge crest is generally flat and stands at an elevation of approximately 220 feet (73m) above present sealevel. Wave-cut escarpments occur along the length of the ridge at 210 feet (70m) and 190 feet (63m) above present sealevel.

The origin of Merriland Ridge has been variously described in the following terms:

1. Katz and Keith (1917) considerd the ridge to be one of the several segments of their Newington Moraine. They suggested that an ice front stood along the western face of the ridge and deposited a submarine apron of sediment toward the east.

2. Bloom (1960) argued that the ridge was shaped by littoral processes that reworked elongate hills of ice-contact stratified drift deposited by retreating ice. Significantly, Bloom's model requires that marine submergence was subsequent to deglaciation at Merriland Ridge.

3. Smith (1982) has suggested that Merriland Ridge is a partial delta (second form of delta of Glückert, 1975) constructed to prevailing sealevel. In the sense that it was built at the ice front, it is an end moraine (and is so mapped), though not a segment of any larger moraine system. As Katz and Keith suggested, the ridge was constructed by ice in a submarine setting. The late-glacial marine limit in this area is approximately 220 feet (73m) above sealevel, the elevation of the ridge crest. Bedrock occurs at shallow depth throughout the area surrounding Merriland Ridge and is exposed at either end of the ridge. Water was therefore probably no more than 10m deep at the time that the ridge was formed.

Continue west on Route 9, following distal slope of Merriland Ridge.

22.0 Johns Swamp Road. Turn right (north). Road rises to crest of Merriland Ridge, then drops down proximal face of the ridge to marine sediments and DeGeer moraines.

23.3 Junction with Bragdon Road. Just before reaching this junction, the road rises onto the Bragdon Road delta, a small kettled delta fronted by a till moraine that is partially covered by distal delta sediments. This delta is approximately 3 km long, 7-8m high, and roughly 1 km from front to back. The top of the delta is at an elevation of approximately 220 feet (73m). The proximal (NW) portion of the delta is marked by numerous kettles and several linear moraine ridges paralleling the ice-contact slope. Several DeGeer moraines occur both north and south of the delta.

Turn right (east) on Bragdon Road. Route follows surface of delta for about 1.5 km then drops to surface of marine sediments.

24.9 Junction with State Route 109. Turn left (north). Road rises from surface of marine sediments to till and bedrock.

26.2 Bear left at Y junction to Saywards Corner.

26.6 Junction with Quarry Road. Turn left (west).

26.7 Turn right (north) into gravel pit.

STOP 4 SAYWARD'S CORNER ICE-CONTACT DELTA

This pit has lately provided the best opportunity to observe the internal structure of the ice-frontal deltas found in this part of the coastal zone. The character of the sediments and the general morphology displayed on this delta are similar to those seen in both Merriland Ridge and the Bragdon Road delta in the past when exposures there were fresh.

Return to cars. Retrace route to State Route 109. Continue south to Maine Turnpike Exit 2 (turnpike entrance is on the left). Follow Maine Turnpike (Interstate 95) south to Danvers...

BRANCH BROOK IN YORK COUNTY MAINE: FORMATION AND MAINTENANCE OF A DRAINAGE NETWORK BY GROUNDWATER SAPPING

By Denis D'Amore
Department of Environmental Quality Engineering
Commonwealth of Massachusetts

INTRODUCTION

The Sanford outwash plain is 163 km ($62.8\ mi^2$) in area and is located in east central York County, Maine (Fig. 1). The late Wisconsinian sequence of very permeable stratified drift ranges in thickness from 8 m (26 ft) near South Sanford to a maximum thickness of 21 m (70 ft) in an area known locally as "The Plains". The outwash plain supplies groundwater to all or part of six streams in the region; the Merriland, Great Works, Mousam and Kennebunk Rivers, and Branch and Day Brooks. The latter two streams are the only ones whose watersheds are completely contained within the outwash plain.

Because of the low relief in the Branch Brook basin (55 m, 180 ft) and the high permeability of the surficial deposits, channel initiation and development is controlled principally by groundwater sapping and seepage rather than by surface runoff. Thus groundwater, not overland flow, is the dominant morphological agent responsible for the evolution of the drainage network of Branch Brook and also for that portion of the Mousam River that incises into the outwash plain.

Bedrock Geology

Four major rock units have been identified in the Sanford-Wells region (Hussey, 1962). Proceeding from east to west, the first unit is a 8 km (5 mi) wide band of calcareous sandstone, siltstone, and graywacke. This is followed by a thin strip of metamorphosed shale, siltstone and minor sandstone. These northeast trending units are referred to respectively as the Kittery and Elliot Formations and are of lower-middle Silurian age. Immediately west of the Kittery and Elliot Formations is a 4.8 km (3 mi) wide belt of lower Devonian calcareous metasedimentary rocks. This unit also trends northeast and is referred to as the Berwick Formation.

The remaining bedrock unit in the area consists of three granitic plutons of upper Devonian age. The largest of these trends northeast and underlies the northwest portion of the outwash plain. This portion also forms the northwest proximal perimeter of the stratified drift. One of the other two plutons forms the distal southern boundary of the outwash plain. It appears that this more southerly pluton diverted both the outwash and Branch Brook around its northern extremity (Fig. 2).

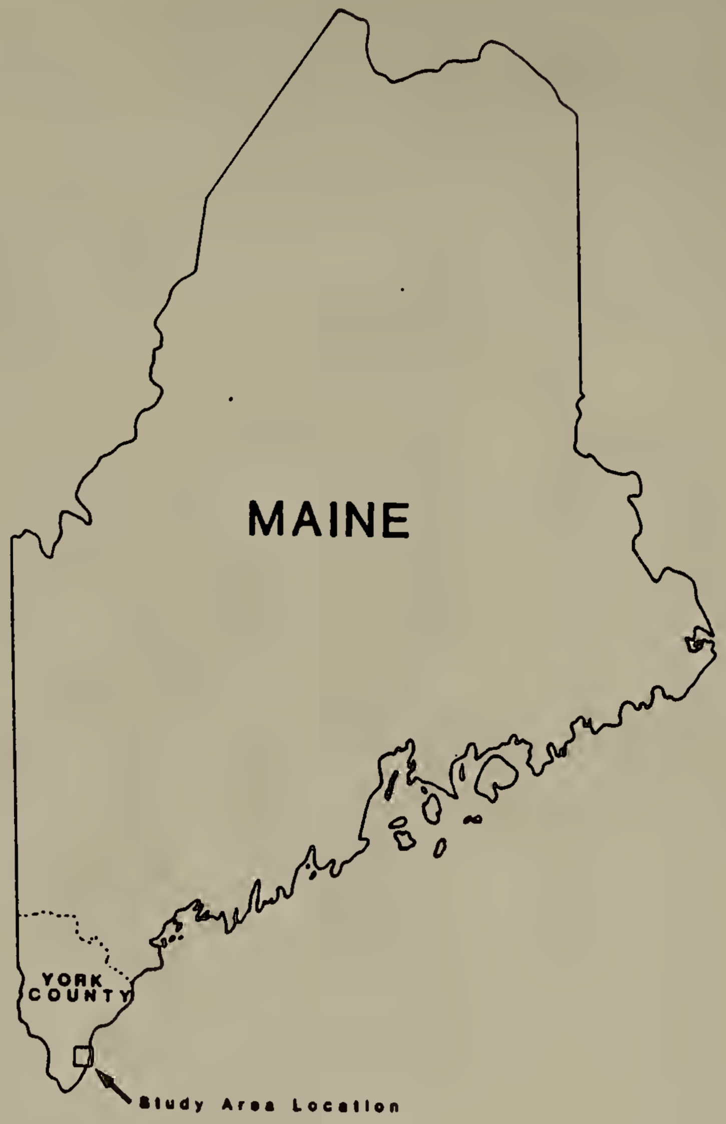

Figure 1. Location map of the study area.

Figure 2. Location of the 3 plutons in the Sanford-Wells region with an approximation of the areal extent of the outwash plain (Hussey, 1982 and Smith, 1977).

When sapping occurs, the water table intersects the land surface and groundwater discharges along a seepage face (Fig. 3). Thus, it should be appropriately termed a seepage face phenomenon.Distinctions between piping and sapping and seepage have been summarized in Table 1. The equipotential lines (lines representing the location of equal values of hydraulic head) are, in a homogeneous isotropic medium, at right angles to those flow lines (Fig. 4a). If there is an embayment such as tidal inlet along the land margin or if the aquifer is heterogeneous and anisotropic, the flow of groundwater is directed toward the embayment or more permeable area and a spring head develops (Fig. 4b). With increased groundwater flow, pore pressure increases at the spring head. The sediments dilate and at some critical threshold gradient the grain to grain contact is lost. When this gradient is reached, the finer particles are washed out of the matrix causing a collapse of the coarser fraction. This leads to an enlargement of the embayment. In the case of more permeable sediments flanked by less permeable material this causes the formation of a small gully. This process results in more flow lines being concentrated at the gully head as the equipotential lines become more distorted around the incipient valley (Fig. 4c). With groundwater now flowing at the head and along the valley walls, the process can repeat itself and tributaries can begin to form as the trunk stream begins to meander (Fig. 4c).

Dunne (1980) calls the concentration of flow lines a positive feedback mechanism whereby the more the gullyhead retreats, the probability of repeated failure due to sapping increases. Because the groundwater flow paths are perpendicular to the discharge face, tributaries generally form at right angles to the first order channel, resulting in what Higgins (1982) refers to as a pectinate drainage newtwork (Fig. 5). Although the initial orientation of the tributaries is perpendicular to the higher order stream, if there are variations in horizontal hydraulic conductivity within the aquifer (horizontal anisotropy), the tributaries will become re-aligned and channel development will follow the path of the more permeable sediments (Fig. 5). As channels and their tributaries continue to migrate headward, at some point springs begin competing for the same groundwater and the positive feedback mechanism becomes less effective and eventually the sapping process stops.

Threshold Hydraulic Gradient For Sapping in Unconsolidated Sediments

A curve identifying the threashold hydraulic gradients necessay to induce sapping in unconsolidated sediments of various permeability is shown in Figure 6. It was developed by measuring headwall hydraulic gradients and conductivities of beach foreshore drainage networks (D'Amore, 1983). This curve represents the minimum conditions necessary for the initiation of the sapping process in unconsolidated sediments of similar densities. The negative slope indicates that coarser sediments (with greater values of K) require smaller gradients (J) to initiate sapping. There is an upper limit, possibly in the granule range above which the sediments become too coarse, thereby decreasing the effectiveness of the seepage face. Above this limit the grains would be too heavy to be dislodged by the outflow of groundwater at the seepage face and other erosional processes would dominate. There is also a lower limit as well, perhaps in the silt or silt-clay range, below which piping rather than seepage becomes the dominant erosional process. It has been suggested that this lower limit might be a function of the degree of coheseivenes of the sediments (La Fleur, pers. comm.).

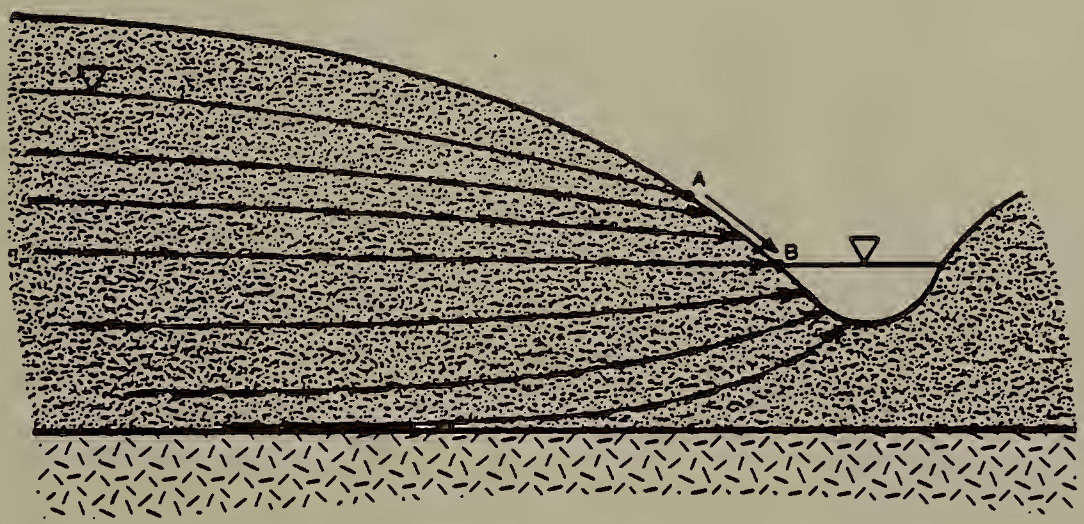

Figure 3. Groundwater discharge along a seepage face. Whenever a phreatic surface approaches a downstream external boundary of a flow domain, it will always terminate on it at a point (A) that is above the water table. The surface A B represents the seepage face (after Bear, 1979).

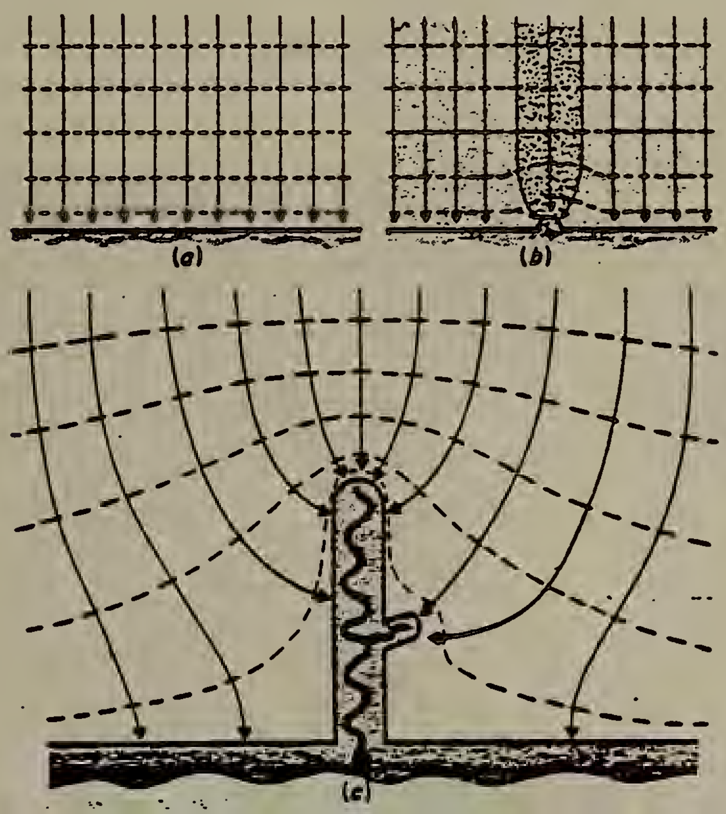

Figure 4. Plan view of a groundwater flow net during extension of spring heads to form a drainage network. Solid arrows are flow lines, dashes indicate equi-potential lines. a) Groundwater flows towards the land margin; b) convergence of groundwater flow at the head of an embayment produced by a small piping failure or by an initial irregularity in the land margin; c) increased convergence of flow lines around a spring head that has retreated headward from the land margin extending a valley. A second piping failure has occurred on one side of the valley and is distorting the flow field in that region. Modified after Dunne (1969).

TABLE 1 Differences Between Piping, Sapping, and Seepage in Unconsolidated Sediments

		Piping	Sapping and Seepage
1.	Type of material	Silty clays and clay loams	Sands and fine gravels
2.	Flow velocity	Turbulent	Laminar
3.	Hydraulic gradient	Horizontal or vertical	Horizontal
4.	Occurrence of erosion	Along the entire interior of the pipe	At the spring headwall
5.	Evolution of surface drainage features	Collapsed roof topography	Headwall retreat
6.	Flow conditions	Saturated or unsaturated zone	Saturated zone

Figure 5. Pectinate drainage patterns formed on a beach foreshore. Many of the tributaries that form perpendicular to the higher order channel tend to become realigned up the beach face due to the anisotrophy of the sediments.

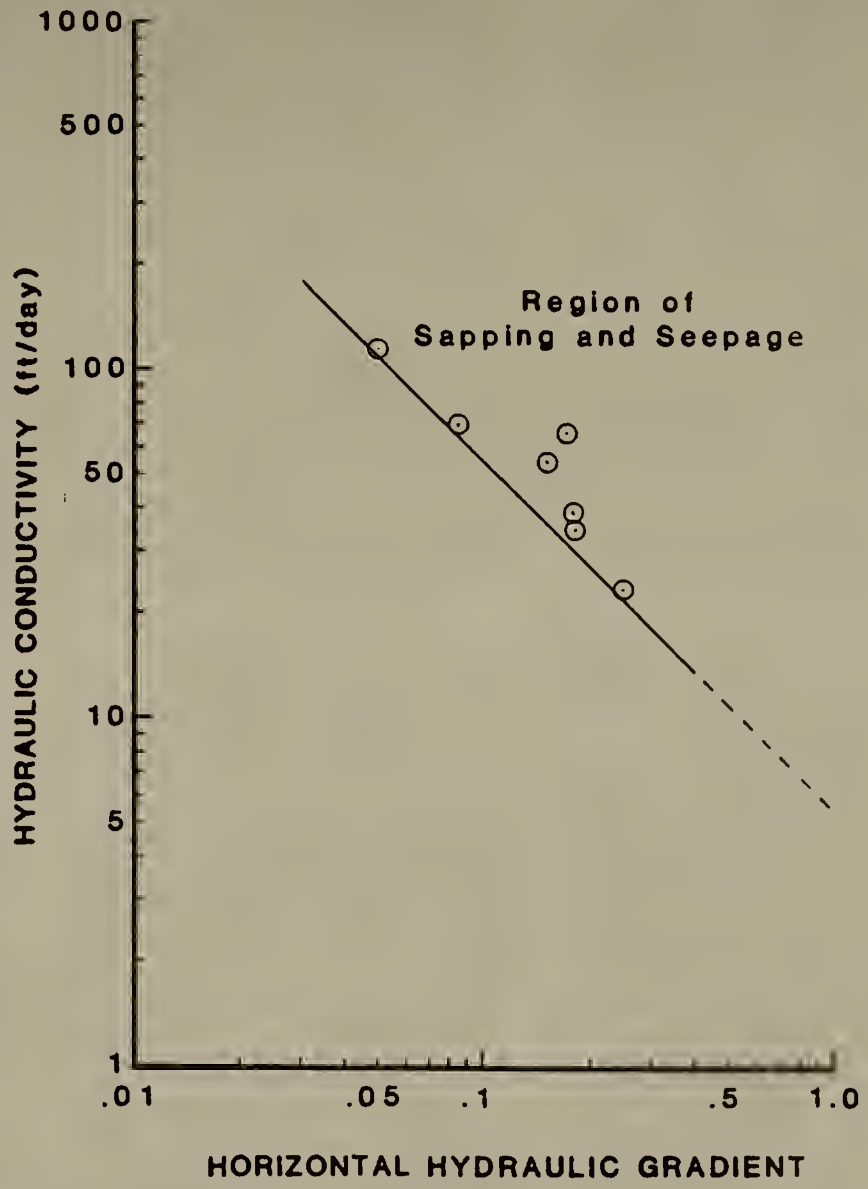

Figure 6. Relationship between hydraulic conductivity and horizontal hydraulic gradient. The solid line represents a theoretical approximation of the minimum slope necessary for the initiation of sapping and seepage in unconsolidated sediments of similar densities. The circles are measured values.

Occurrence of Sapping in the Sanford Outwash Plain

Of the six coastal streams in the outwash plain, the sapping process is the dominant morphologic agent in only those stream that dissect the sand and gravel of the stratified drift. The process is not observed in streams that are incising into the Presumpscot Formation or into till. Branch Brook and that portion of the Mousam River that is within the outwash plain show the consequences of extensive sapping (evidenced by the presence of numerous steep sided gullies trending at right angles to the main stream) (Fig. 7). The Merriland River, Day Brook, and the Kennebunk River, all of which dissect either till or sandy, silty clay, show no effects of sapping.

In the regions of the outwash plain where sapping is prevalent, many of the springs are ephemeral, flowing only during the late spring and early summer when the water table and thus the hydraulic gradient is high. Even in those springs that flow year round, headwall sapping is intermittent, occurring when the threshold gradient for those sediments has been reached. The driving mechanism for this process is natural recharge. During the months when evapotranspiration is low, the local hydraulic gradient to a flowing spring increases until the threshold is reached at which time the sapping process begins. In like manner, flow is initiated in ephemeral springs and then at some higher gradient headwall sapping starts.

Model for the Evolution of Branch Brook

On beach foreshores, as the tide recedes, base level is lowered more rapidly than groundwater discharge occurs. If the difference between the rate of groundwater outflow (i.e. the seepage velocity) and the receding tide is great enough a threshold gradient will be reached and sapping can begin. In contrast to the beach foreshore model, tectonic uplift or glacial rebound might produce a similar effect in fluvial drainage systems. Along the southwestern coast of Maine isostatic rebound following the retreat of the Late Wisconsian Laurentide Ice Sheet might have occurred quickly enough to establish such a gradient (Fig. 8).

This type of rebound particularly if it occurred spasmodically rather than continually could have allowed the saturated subaqueously deposited outwash plain to begin discharging its saline groundwater into the ocean with sufficient force to initiate sapping. The path of least resistance for groundwater flow (i.e. the most permeable sediments in the outwash plain) would have been the distributary meltwater channels buried in the stratified drift which were preferentially oriented down the long axis of the outwash plain. Groundwater flow would have been directed to the ocean via these underground highly conductive zones. The presence of an embayment at the mouth of the present-day Branch Brook would have further accentuated groundwater flow to the inlet causing rapid retreat of the headwall thus permanently establishing the stream channel. As salt water continued to be discharged, it would be replaced with meteoric waters. In fact, saline waters still remain in some of the deeper sediments in the region where groundwater flow is stagnant (Caswell, 1979).

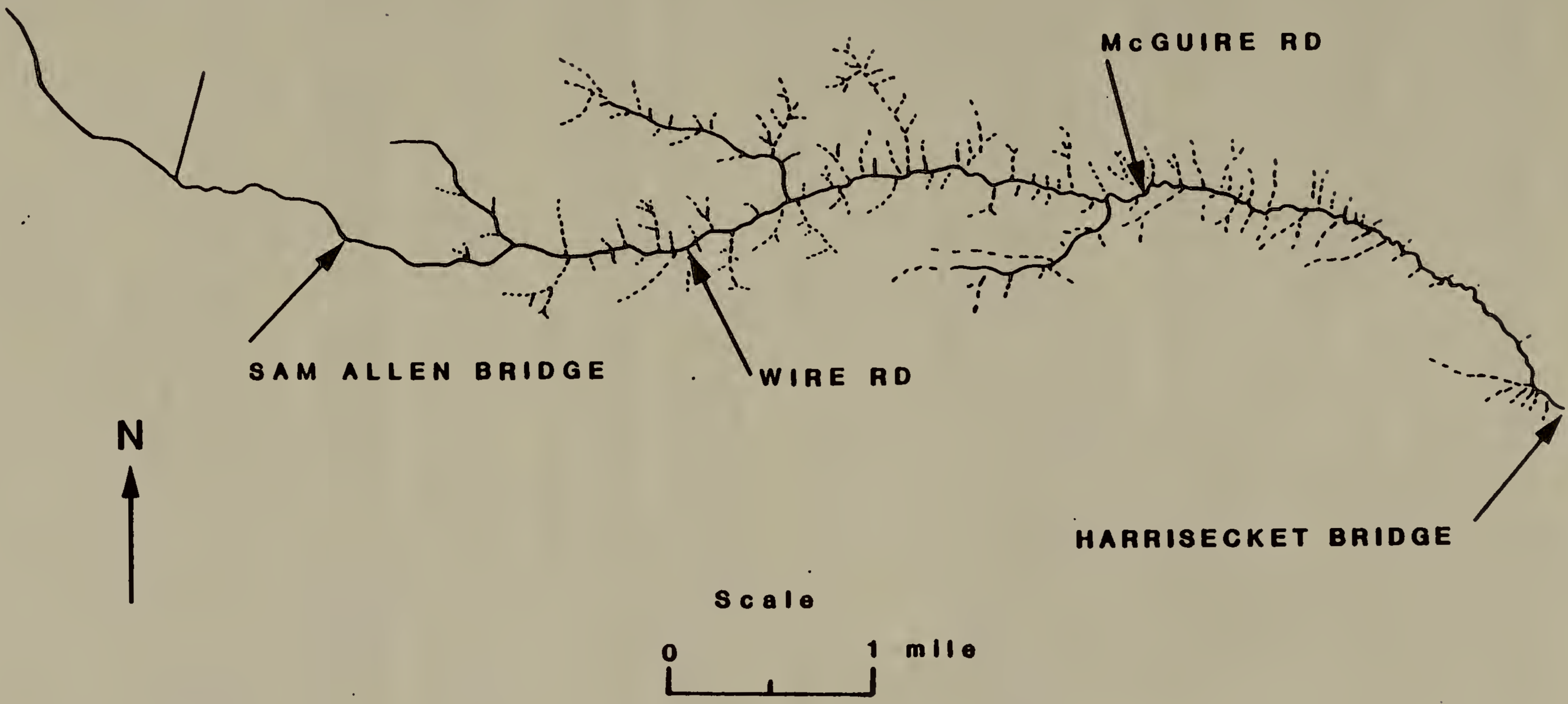

Figure 7. Pectinate drainage pattern of Branch Brook (dashed lines represent ephemeral streams). All of the larger tributaries have become reoriented in the direction of maximum or preferred hydraulic conductivity which is in the flow direction of the glacio-fluvial meltwater streams of the retreating Wisconsinan Ice Sheet.

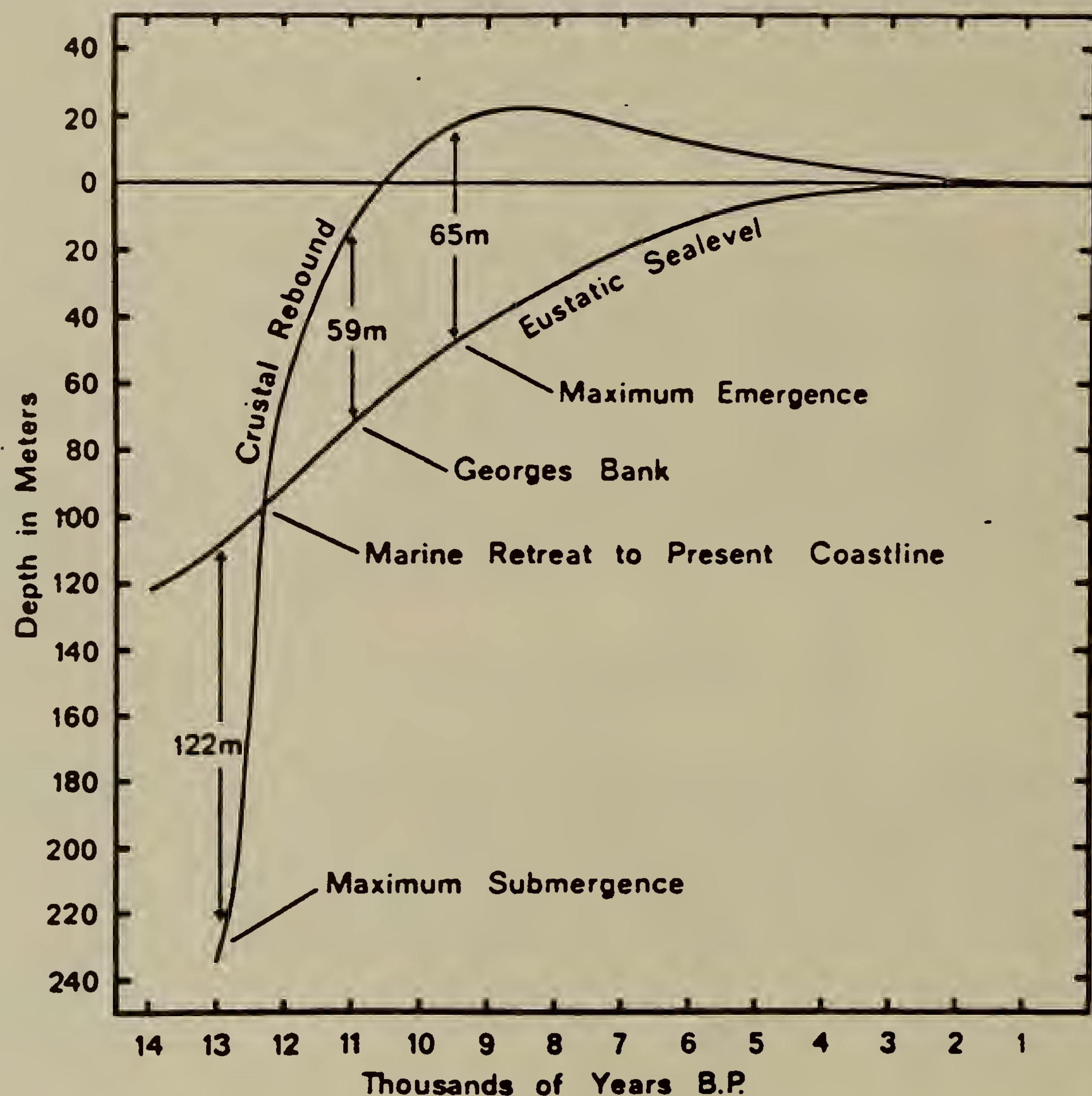

Figure 8. Crustal rebound curve of the Gulf of Maine area (Schnitker, 1974). Sea level curve from Milliman and Emery, 1968.

Road Log

The field trip will leave the conference headquarters in Danvers promptly at 6:30 A.M. Low tide is at 6:57 A.M. so we would like to be on the beach and engaged in discussion by then.

Go east on Rte, 62 to Rte, 128. Proceed north on Rte. 128 to the Manchester, School St. exit (Exit 15). Take a right onto School St. for approximately 0.55 miles and go left on Lincoln St. (It is the second left after the blinking yellow light). Go to the end of Lincoln St. (0.5 mi.) and take a left onto Rte, 127. Take the first right onto Ocean St. after going under the railroad bridge (Ocean St. is 0.85 miles after Lincoln St.). White Beach will be on the right (o.35 miles).

(45 min.) STOP 1. Here we will look in detail at the seepage face phenomenon; how the process works and examine what the controls are operating on the drainage networks that form in this kind of an environment.

After leaving White Beach we will return to Rte. 128 by going north to Rte. 133, west on Rte. 133 to Rte. 95 and north on Rte. 95 to the Wells exit (Interchange No. 2 on the Maine Turnpike). Go east on Rte 9 & 109 to Wells Center and take a right onto Rte. 1 north. Mileage will start here at Wells Corner.

Mileage	
0	From Wells Corner go north on Rte. 1. (We'll stop for coffee and donuts here.)
2.8	Turn left onto Harrisecket Road which is directly across from Johnson Hall Auctioneers (look for the white columns).
3.2	Turn right into a gravel pit which is located at the distal end of the outwash plain.
(30 min)	STOP 2. This is an excellent location to observe the heterogeneity of the outwash deposits and to hypothesize as to what the preferred groundwater flow paths are under saturated conditions.
4.6	Proceed down Harrisecket Road and stop at the bridge. (Park well off the road being careful of traffic and soft shoulders.)
(45 min)	STOP 3. On the flood plain of Branch Brook we will be standing on the Presumpscot Formation (an outcrop of which we will see). We will be at the base of the stratified drift suggesting that the drift is younger than the Presumpscot Formation. We will also look at some interesting morphological features of Branch Brook.
5.4	Take left at the stop sign onto Rte 9A cross over Rte 95 and bear to the right onto Maguire Rd.
5.9	Go right onto Whitten Street and park just left of the guard rail.

Road Log
Continued

Mileage

(30 min) STOP 4. At this location, Day Brook, we will again be in the Presumpscot Formation only now we will be above the stratified drift. From this we can conclude that as the Laurentide Ice Sheet retreated, the outwash plain was saturated with saline ground water.

Make a U turn and return to Maguire Rd.

8.4 Proceed west on Maguire Rd. After 2 miles it will become a dirt road. Pass beneath the transmission lines and look for a white PVC observation well on the left.

(1 hr) STOP 5. Here we will look at some actively eroding gullies and some that are no longer in an erosional profile and offer suggestions as to why this happens. We will also discuss a model for the evolution of Branch Brook.

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DEGLACIATION OF THE MERRIMACK RIVER VALLEY, SOUTHERN NEW HAMPSHIRE

Carl Koteff[1], Byron D. Stone[1], and Dabney W. Caldwell[2]

Due to time and distance problems of the authors, a road log for this field trip is being issued separately. For similar reasons and because exposures of glacial deposits in this area are so ephemeral, specific field trip stops are not identified in this article. Detailed descriptions and interpretations are included, however, for the most likely stops. We intend to visit at least one glacial lake spillway, an exposure of collapsed lake sediments overlain by noncollapsed stream-terrace deposits, and exposures of deltaic deposits that pertain to the problems of lake damming, lacustrine and bog sedimentation, and postglacial uplift. Other appropriate stops can be made as time permits.

Introduction

This trip is concerned chiefly with deglacial meltwater features resulting from the last glaciation, but a few comments on earlier glacial history are in order. Although there were several Pleistocene glaciations that probably covered the region, solid evidence for just two major ice advances in New Hampshire has been demonstrated by detailed field work in recent years (Koteff, 1970; Newton, 1978; Koteff and Pessl, in press). Newton (1978) discusses the possibility of a third glaciation, although the evidence is very fragmentary. At that, the older of the two recognized glaciations is known only from scattered exposures of a compact, relatively silty till that is oxidized to a depth of as much as 25 feet (7.6 m). In contrast to the older (lower) till, the younger (upper) till is very widespread in distribution, is more sandy, and has been only slightly affected by oxidation. Therefore, the upper till is considered to be late Wisconsinan in age. Dating of the lower till is more problematical, and at present it is considered to be either early Wisconsinan or older, possibly Illinoian in age.

Deglaciation of the Merrimack Valley by the late Wisconsinan Laurentide ice sheet was characterized almost entirely by systematic stagnation-zone retreat (Koteff and Pessl, 1981), as was the case for all of southern and most of central New England. In this concept, the retreating ice sheet is viewed as having had a fringe of stagnant ice, perhaps at most a few kilometers wide, that acted as a buffer between the live ice and the meltwater deposits, or morphosequences, which were deposited partly within but mostly beyond the stagnant zone of ice. The numerous morphosequences in the Merrimack Valley are chiefly ice-marginal deltas and related fluvial and lake-bottom meltwater deposits that indicate ice retreat took place, for the most part, in a glaciolacustrine environment. The ice-marginal deltas and related features mark many former and successively younger positions of the stagnant ice edge from Tyngsboro, Massachusetts, to Concord, New Hampshire, and beyond. Only

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one interruption in the more or less steady ice retreat occurred near Manchester, New Hampshire, where evidence of a minor readvance of no more than a few kilometers has been found (Stone and Koteff, 1979).

Glacial Lakes

Formation of glacial lakes

In previous years (Goldthwait, 1925), it was thought that the rising marine waters that followed the retreating ice margin during deglaciation inland from the New England coast had extended up the Merrimack Valley as far north as Manchester, New Hampshire. However, Goldthwait (1938) makes no mention of this premise in his later article on uncovering of New Hampshire by the last ice sheet, and White (1938), in a brief discussion of glacial lakes in the Merrimack Valley, indicates that the former water body located northward from Nashua, New Hampshire, was lacustrine. So apparently by this time, the idea that the late-glacial sea had occupied this part of the Merrimack Valley had been discarded. Detailed mapping in recent years (Koteff, 1970, 1976) shows that three separate and successively formed glacial lakes existed in this region of the Merrimack Valley (Figure 1): Lake Tyngsboro, which occupied a small area across the Massachusetts-New Hampshire border; Lake Merrimack, which extended from Nashua to Manchester; and Lake Hooksett, which extended from about 3 mi (4.8 km)south of the village of Hooksett to north of Concord. The history of these lakes is somewhat unusual in that each of the last two lakes formed at a slightly higher level than that of the preceeding one (Figure 2). In north-draining valleys in New England, successive glacial lakes formed at successively lower levels as the retreating and damming ice uncovered successively lower spillways or outlets. In the south-draining Merrimack Valley, however, a different situation existed. Lake Tyngsboro formed behind meltwater deposits, probably deltaic, that dammed the valley near the village of Tyngsboro, Massachusetts, when the ice front stood there (Figure 1). This lake emptied via a bedrock-floored spillway at a present day altitude of about 140 ft (43 m). When the ice margin had retreated about 2.5 mi (4 km) north of the Massachusetts-New Hampshire border, deposition of the last of a series of Lake Tyngsboro deltas completely filled a narrow part of the valley there. These deposits, which have an altitude of at least 185 ft (56 m), blocked meltwater flow and served as a dam for the newly formed glacial Lake Merrimack. At the same time, an adjacent north-south gap or spillway floored in bedrock was uncovered (Figure 1). Although this spillway, at an altitude of 170 ft (52 m) was at or possibly 3 ft (1 m) lower than the inferred level of Lake Tyngsboro at this location, its narrowness caused hydraulic raising of the Lake Merrimack level to an altitude of about 178 ft (54 m). This allowed meltwater of the newly formed and expanding Lake Merrimack an escape route around the damming delta deposits of the final stage of Lake Tyngsboro, rather than overtopping and eroding them. Thus, if it were not for the propitiously positioned spillway for Lake Merrimack, meltwater flow would have prevented Lake Tyngsboro deposits from completely blocking the valley, and Lake Merrimack would have been confluent with Lake Tyngsboro, rather than being at a higher level.

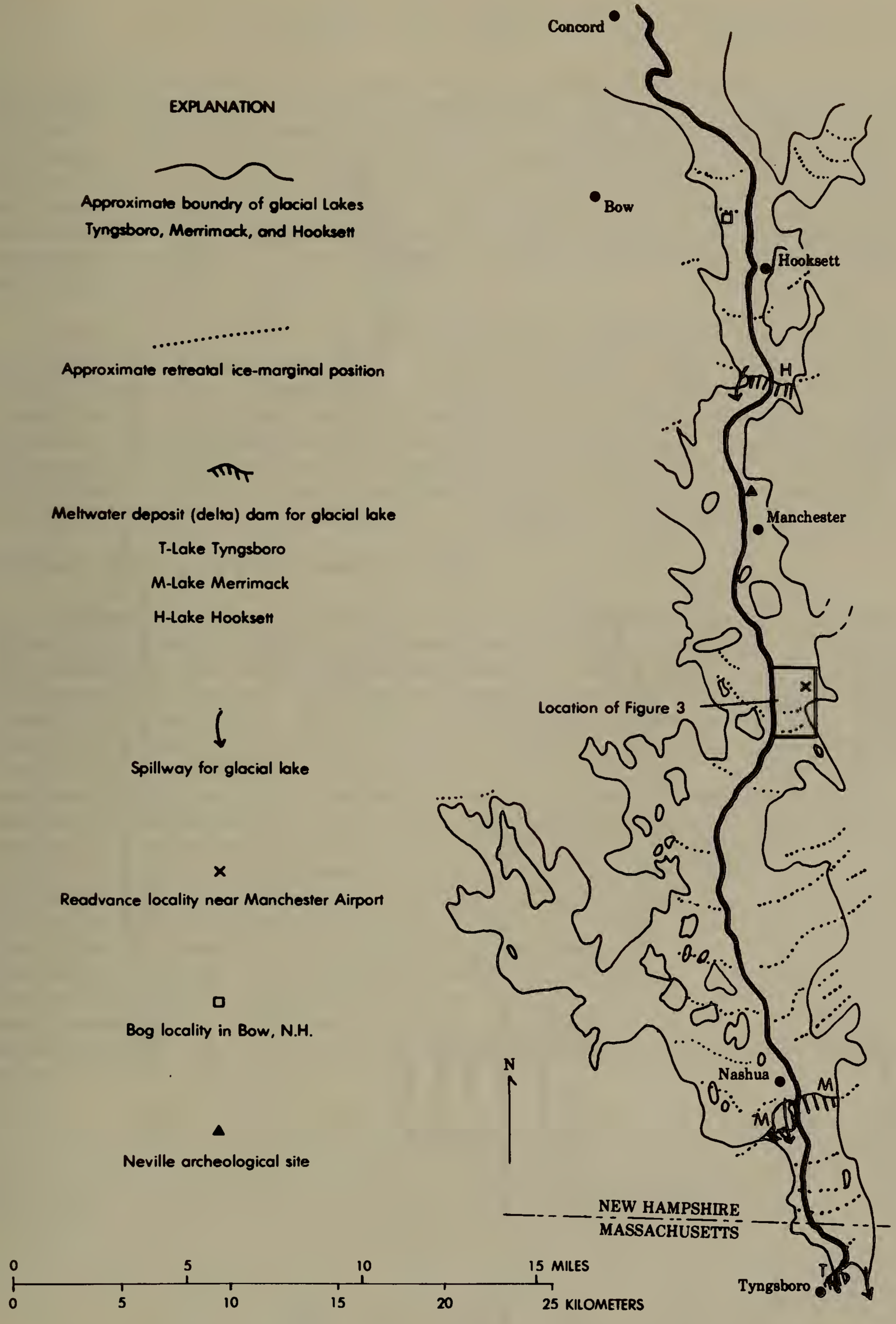


Figure 1. Outline of glacial lakes in part of the Merrimack Valley, Massachusetts—New Hampshire.

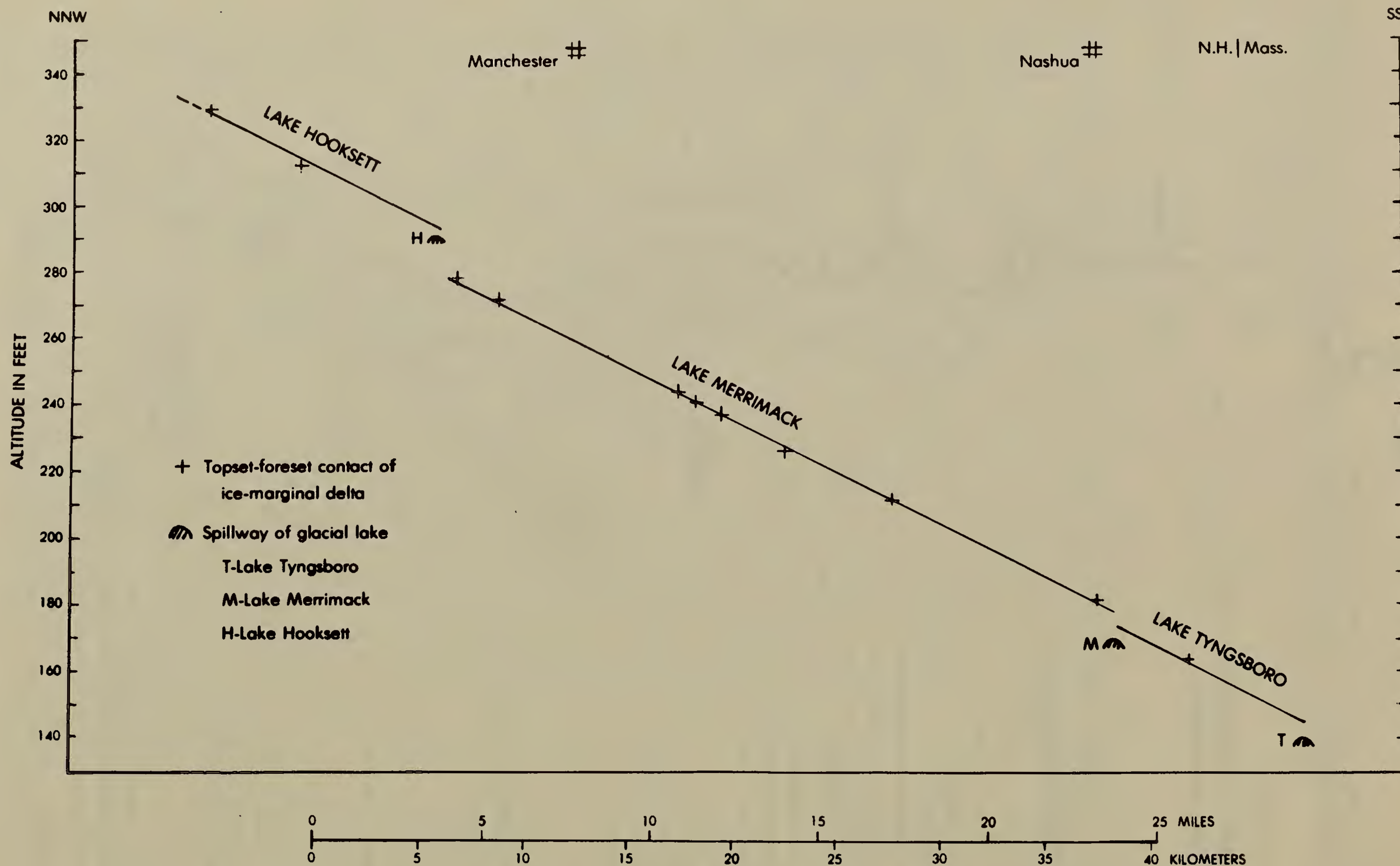


Figure 2. Profiles of glacial-lake levels in the Merrimack Valley. Projected gradient is 4.7-4.9 ft/mi (0.89-0.93 m/km) up to NNW. Direction based on best fit by inspection of line to the data points.

A combination of local topography and delta deposition north of Manchester (Figure 1) and about 3 mi (4.8 km) south of the village of Hooksett similar to that which caused the stepping up from Lake Tyngsboro to Lake Merrimack, resulted in the formation of Lake Hooksett at a higher level than that of Lake Merrimack. At this location, the projected level of Lake Merrimack is about 280 ft (85 m). The surface of the last Lake Merrimack delta deposits that blocked the valley here is about 300 ft (91 m), and the bedrock floor of the nearby spillway for Lake Hooksett is just over 290 ft (88 m) in altitude.

The delta dams that separated the three lakes were not very large, and their heights of less than 10 ft (3 m) above water seem somewhat low to have allowed them to be preserved for any length of time. However, as long as Lakes Tyngsboro and Merrimack existed, their relatively high water levels probably retarded sapping and headward erosion of their respective dams. Subsequent erosion has destroyed much of these dams, and only remnants are left today. It is not known how long the three lakes lasted, but they all appear to have been coincident for at least part of late glacial time.

Sedimentation in Glacial Lake Merrimack

Sedimentary facies exposed in the vicinity of Manchester Municipal Airport (Figure 3) include good examples of both ice-marginal deposition in Lake Merrimack and later fluvial erosion of a delta plain to a lower level. Altitudes of projected lake levels (Figure 2) in the area covered in Figure 3 range from about 230 ft (70 m) in the southern part to about 245 ft (75 m) in the northern part. The hill at more than 230 ft (70 m) altitude (Figure 3, locality a) and the terrace at more than 240 ft (73 m) altitude to the southeast are interpreted as remnant deltaic deposits representing two successive retreatal positions of the stagnant-ice margin. The nearly flat surfaces at Manchester Airport at over 220 ft (67 m) and east of locality a at over 210 ft (64 m) are stream terraces cut into older higher delta plains. To a large extent, the slope and distribution of the deltaic structures and sediments beneath Manchester Airport have been concealed by the stream-terrace plain and obliterated by erosion by the Merrimack River on the west side of the plain. However, textures, sedimentary structures, attitudes, degree of ice-contact collapse deformation, and the position of various facies seen in past exposures at localities a through e in Figure 3 permit a detailed delineation of at least two ice-marginal deltas and associated lake-bottom sediments.

The generalized vertical stratigraphy of the glaciolacustrine sediments is bipartite. Lake-bottom silt, clay and sand in irregularly spaced rhythmic laminations crop out at the base of exposures along the brook that bisects the terraces. At one place, across the brook from locality e, Figure 3, these rhythmites directly overlie bedrock. The lake-bottom sediments are overlain by deltaic foreset strata which coarsen upward from sand at the base to pebble-cobble gravel near the level of Lake Merrimack at the top of deltaic sections. Because of extensive stream-terrace erosion in this area, probable delta-topset beds of coarse gravel are preserved only locally (Figure 3, locality a). Shallow excavations in the surface of the plain around Manchester Airport exposed coarse pebbly sand and pebble gravel of the stream-terrace deposits that disconformably overlie the eroded deltaic beds. The textures of the terrace beds are similar to textures of well exposed, younger, lower, and

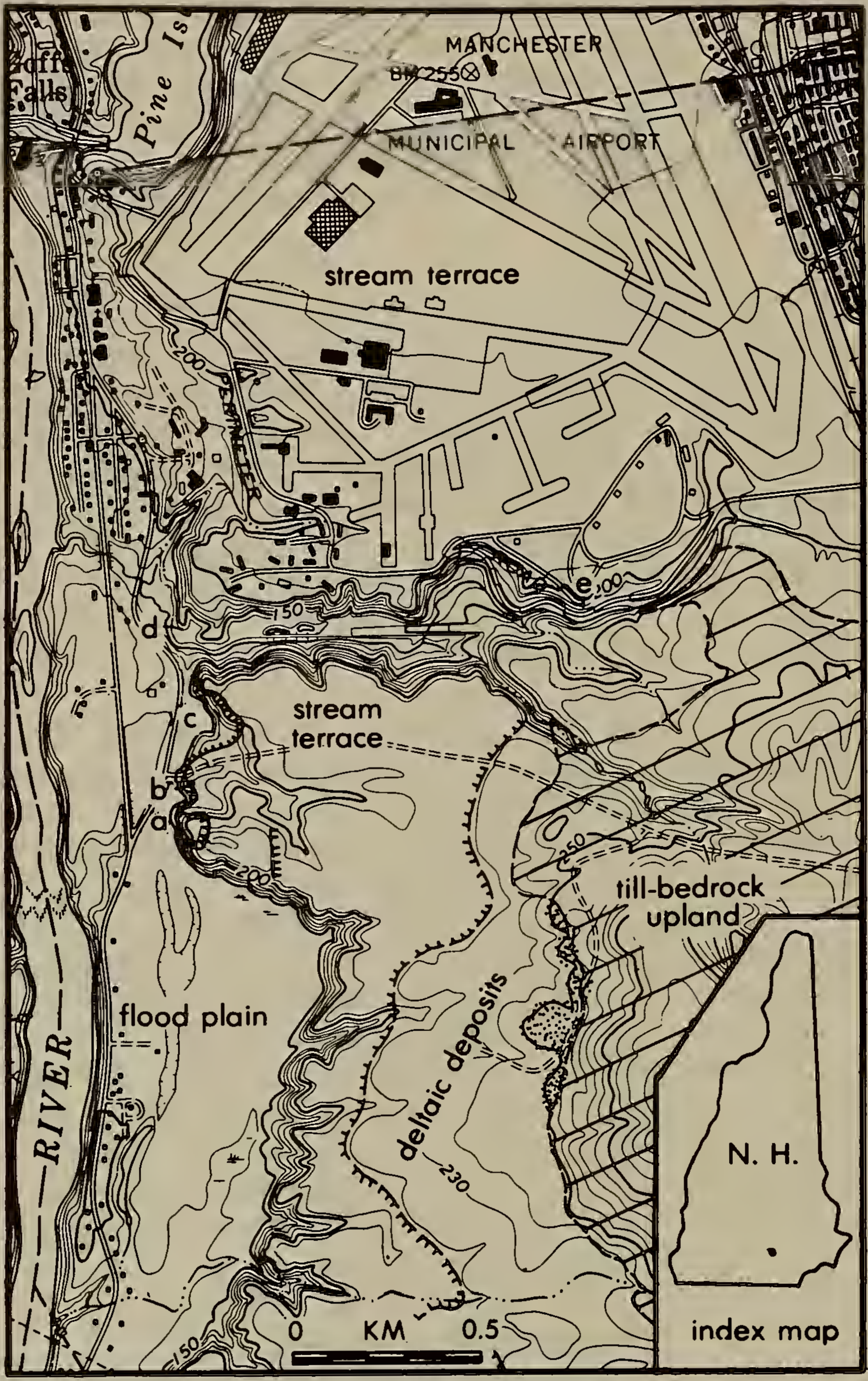


Figure 3. Morphology and stratigraphic localities of stream-terrace and glaciolacustrine deposits in the Manchester Municipal Airport area. Stippled pattern shows area of eolian sand deposits. Lettered localities referred to in text. Dashed ticked line indicates erosional scarp, ticks on downslope side. Base map from U.S. Geological Survey Manchester South 7¼′ quadrangle, 1968 edition. Continuous hachured line shows extent of pit excavations, 1970.

better defined stream-terrace deposits elsewhere in the valley. The terrace deposits are notably finer grained than gravel topset beds of ice-marginal deltas in the Merrimack Valley and of ice-marginal deltas in general (Stone, 1984), but are coarser than most delta foreset and lake-bottom sediments.

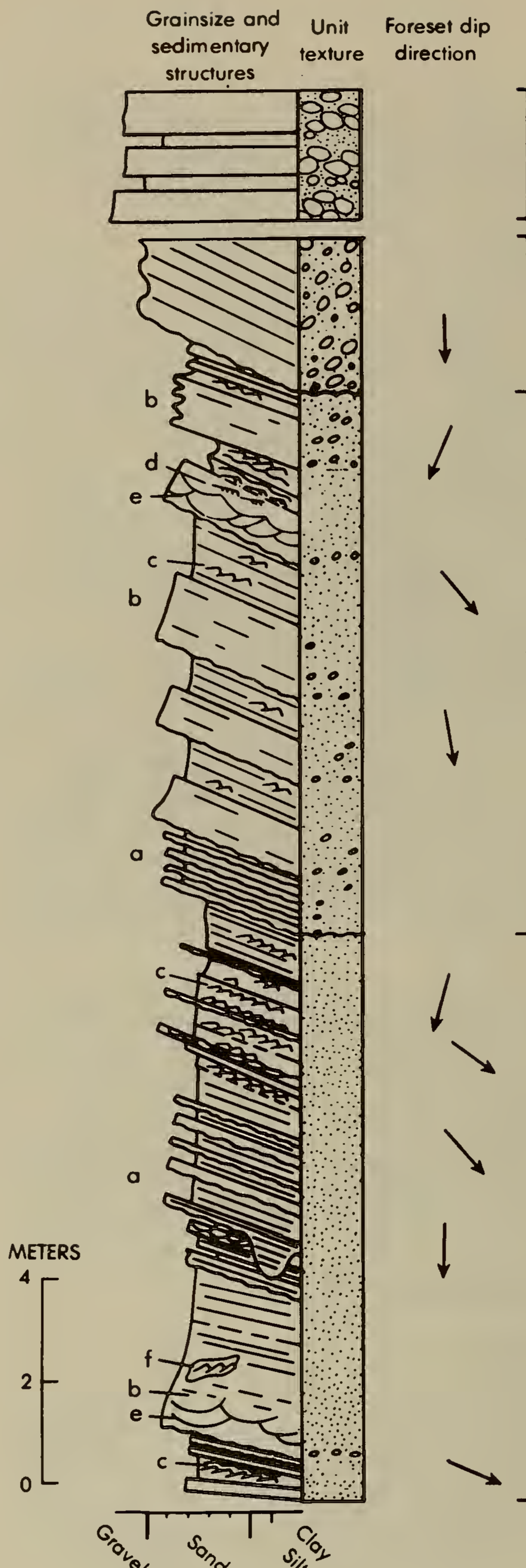
Vertical measured section in ice-marginal deltaic beds

Figure 4 is a composite stratigraphic section at locality a in Figure 3 showing coarsening-upward deltaic foreset beds and probable topset beds at the top. The foreset beds are divided into three units on the basis of textures and sedimentary structures. The lowest unit in Figure 4 (unit 1) consists of medium to coarse sand in laterally continuous planar beds with sharp, erosional lower boundaries, and in beds containing sets of climbing-ripple cross laminations. These structures, and the sharp erosional discontinuities between sets of foreset strata, suggest traction-load deposition by density underflows down slopes of 12-24°. Beds containing sets of climbing ripples and trough cross beds intercalated with locally tabular sets of coarse sand beds in unit 2 indicate that these foresets were deposited by density underflows, alternating with periods of avalanche, grain-flow deposition in foreset beds dipping 18-27°. The coarse foreset beds of unit 3 are contained in a tabular set of beds that terminates at a sharp disconformity at the base. These beds are interpreted as products of grain-flow deposition just beneath the surface of the lake. A laterally extensive exposure across the strike of these coarse foreset beds showed that they prograded into the lake along an azimuth of 178°. This upper foreset progradation direction and the range of other foreset dip azimuths are shown in Figure 5A, which illustrates the initial stages of progradation of the upper part of the delta at locality a, Figure 3.

A previously excavated zone of ice-contact deformation of the delta at locality a, Figure 3 showed normal faults that displaced the deltaic foreset strata down to the northwest. These structures, as well as gravity-flow deposits that flowed from the ice-contact slopes out onto the adjacent lake bottom indicate that the supporting ice wall of the delta extended through the pit at locality a. The direction of gravity-flow transport and related ice-contact faulting and gravity-flow deposits (Figure 3, locality b; Stone, 1976) are shown in Figure 5B.

Younger ice-marginal deltas and associated lake-bottom facies

Sandy foreset beds of an ice-marginal delta younger than the delta at locality a have been exposed in a pit locality c, Figure 3. Discontinuous exposures from the younger delta to the area of locality a showed that delta bottomset sand beds and distally equivalent lake-bottom silt and sand extended from the younger delta to the foot of the ice-contact slope of locality a. At locality a these younger bottomset beds are composed of more than 15 ft(5 m) of rhythmically laminated silt, fine sand, and gritty clay. More than 200 rhythmite couplets,averaging one inch (2.5 cm) in thickness, were exposed. Coarse layers of each couplet contain multiple laminations and microlaminations, local scour lenses, and ripple cross-laminations. The ripples indicate southeast paleocurrents, and the sediments presumably were derived from an active delta or lacustrine fan source to the northwest. The fine layers of the couplets are composed of gritty clay, are 1-7 mm thick, and occur as a drape in some places above ripples. At the base of the section,



Description of Units

Unit 4. Fluvial beds
Interbedded pebble-cobble gravel, pebbly coarse sand, and medium to coarse sand in horizontal beds.

Unit 3. Upper delta-foreset beds.
Pebble-cobble gravel and pebbly coarse sand in dipping tabular beds, shown schematically.

Unit 2. Middle delta-foreset beds.
Chiefly coarse sand and pebbly sand in dipping strata containing: (a) sets of parallel planar beds, (b) sets of planar laminae and very thin beds with indistinct bedding, (c) sets of B-type and (d) A-type climbing-ripple cross laminations, and (e) trough cross-beds.

Unit 1. Lower delta-foreset beds.
Medium to coarse sand in dipping strata containing: (a) sets of parallel planar beds, (c) sets of B-type climbing-ripple cross laminations, (e) trough cross-beds, (b) sets of planar laminae with indistinct bedding, and (f) intraclast of rippled sand.

Figure 4. Measured section of ice-marginal deltaic beds exposed at locality a, Figure 3. Figure modified from method of Selley (1970) to show delta foreset dips, bedding within unit sets of foreset strata, and unit textures.

Figure 5. Interpretive maps showing successive ice-marginal deltas and inferred positions of the ice margin in the Manchester Airport area (central part of Figure 3). (A) Initial extent of delta at locality a, Figure 3, including range and average of foreset dip directions of Figure 4. (B) Final extent of delta at locality c, Figure 3; (1) northwest direction of gravity-flow deposits and (2) slump slip-line direction (Stone, 1976) at ice-contact slope of delta, locality a, Figure 3, and (3) range of paleocurrent directions in lake-bottom sediments, locality a, Figure 3.

gravity-flow deposits, composed of poorly sorted sand and containing silt-lamina intraclasts are intercalated within the rhythmite sequence. These rhythmites are interpreted as proximal lake-bottom deposits in a prodelta basin of probably limited local extent. The couplets may be annual deposits; however, because the couplets are dominantly coarse-layered (similar to rhythmite facies III of Ashley, 1975) and because the thin, fine-grained layers are silty clay and appear to be genetically related to silt draped laminae, we suggest that the couplets are not varves but may reflect a regular, perhaps diurnal cycle of turbid underflow deposition. Spectacular intraformational isoclinal folds deform sets of several couplets. These folds do not show a systematic sense of overturning or translation and are interpreted as resulting from differential compaction and liquefaction. Following this load deformation, collapse deformation caused by melting ice faulted and rotated the rhythmites valleyward, down to the northwest.

Even younger ice-marginal deltaic strata underlie the terrace at Manchester Airport. These foreset beds probably accumulated in a composite delta built from successive ice-marginal positions now concealed beneath the stream-terrace deposits. These younger deltas prograded over strata earlier deformed by an ice readvance (Figure 3, locality e), and over a different sequence of rhythmites (Figure 3,locality d). Because these rhythmites are not overlain by till, and because they do not contain ripple cross-laminations or relatively thick coarse-layered part of couplets, they are interpreted as younger lake-bottom sediments associated with basin filling by the deltas at the airport.

Ice readvance near Manchester, New Hampshire

Evidence for only one minor readvance (Stone and Koteff, 1979) during systematic and relatively steady ice retreat of the glacier margin has been found in the area near Manchester Municipal Airport (Figure 3, locality e). An exposure at this locality in previous years showed till overlying deformed rhythmically bedded lake-bottom clay, silt and fine sand. Rhythmites in the lower 3 ft (1 m) of section are broadly warped and contain small normal and thrust faults having traces shorter than one inch (less than 2 cm). A thin shear zone separates the warped rhythmites from overlying discrete blocks of rotated rhythmites that are as much as 3 ft (1 m) across. Bedding in the rotated blocks dips generally 80^{o} to the southeast. Another shear zone bounds the upper part of the deformed lake-bottom sediments and is overlain by as much as 5 ft (1.5 m) of till. The till, which is clayey at the base, grades upward into sandy till similar to the widespread surface till of the area. Undisturbed gravel and lake-bottom sand associated with the large Manchester Airport delta overlie the till. Our interpretation of the exposure is that it represents a local ice readvance, probably of less than a few kilometers, over lake-bottom sediments of Lake Merrimack. The advancing ice deformed and detached blocks of these sediments and overrode them, depositing till. The source of the clayey till was the local lake-bottom sediments, and the sandy portion represents ice transport from a greater distance. During subsequent ice retreat, deposits of the Manchester Airport delta covered the readvance locality, and systematic and steady ice retreat once again characterized deglaciation of the valley.

Age of Deglaciation

Although there are slightly conflicting data, the retreat of the last Wisconsinan ice sheet from the Merrimack Valley area is thought to have taken place before 14,000 years BP. Peat obtained from a kettle that formed in Lake Hooksett deltaic deposits at Bow, New Hampshire, (Figure 1; Caldwell et al., 1978) has been dated at 13,050 ± 325 years BP (GX-4554), and Kaye and Barghoorn (1964) have reported a date of 14,250 ± 250 years BP (W-735) for barnacles in marine clay at West Lynn, Massachusetts. These dates suggest that deglaciation of southern New Hampshire occurred near or slightly later than 14,000 years ago. However, Davis and Ford (1982) have reported a date of 13,870 ± 560 years BP (GX-5429) from organic silt at Mirror Lake, New Hampshire, about 45 mi (72 km) north of Concord, N. H. They feel that sedimentation in Mirror Lake began by at least 14,000 years ago Also, a limnic sediment at Hawley Bog Pond (Bender, et al., 1981) near Plainfield, Massachusetts, has been dated at 14,000 ± 130 years BP (WIS-1122). Detailed mapping of Massachusetts and southern New Hampshire clearly indicates that ice retreat in the region was systematic and uncomplicated by major readvances. This mapping demonstrates the physical continuity of meltwater deposits and the successively younger ages of ice-marginal positions of the retreating glacier from Massachusetts into the Merrimack Valley of New Hampshire. Therefore, we feel that the data from Mirror Lake and Hawley Bog Pond provide a more realistic date for at least the minimum age for deglaciation in the area, which is earlier than 14,000 years BP.

Postglacial Erosion and Deposition

All three glacial lakes drained at some point during late glacial and early postglacial time when their respective drift dams failed, probably after 13,000 years ago, the date obtained from the peat at Bow, New Hampshire. Whether or not the drainage of all three lakes was a domino effect set in motion by initial draining of Lake Tyngsboro when the deltaic dam at Tyngsboro was breached is not known. But, as previously mentioned, the small size and low heights of the drift dams suggest that it would have required very little subsequent erosion to cause their failure once each lake to the south drained. Upon lake drainage, the ancestral Merrimack River began eroding deeply into the glaciolacustrine sediments and constructing a series of successively lower stream terraces. The stream-terrace deposits, as much as 15 ft (4.6 m) thick, are composed mostly of sand and gravel, and are coarser than most of the underlying glaciolacustrine sediments.

Most of the terrace building was completed and the Merrimack River had cut down to about its present course probably before 8,000 years ago. Charcoal obtained from the Neville archeological site located on a stream-terrace (Dincauze, 1976) near the Amoskeag Bridge in Manchester, New Hampshire (Figure 1), has yielded a radiocarbon date of 7740 ± 280 years BP (GX-1746). This terrace is the lowest stream-terrace above the modern floodplain, and the Merrimack River there flows on bedrock over a rapids, referred to locally as the Amoskeag Falls. Dincauze (1976) concludes that the Neville site was occupied by people attracted by migrating fish that could be taken at the falls. Thus, the Merrimack River appears to have been established by that time in its present course, at least in this part of the valley and probably for most of its length.

During downcutting by the Merrimack River and the formation of stream terraces, there was intensive eolian activity in parts of the valley, particularly on the east side between Nashua and Manchester. Large sand dune deposits as much as 40 ft (12 m) thick were formed predominantly by winds from the northwest. The dunes, which are composed of fine to medium well-sorted sand, occur as transverse and longitudinal dunes as well as amorphous masses. The sand was derived from glaciolacustrine sediments and probably from the higher stream-terrace deposits also. No dunes occur on the very lowest stream terraces and the modern floodplain.

Postglacial Uplift

Features related to glacial Lakes Tyngsboro, Merrimack, and Hooksett have provided evidence for the isostatic response of the crust that resulted from unloading of the late Wisconsinan ice sheet. Altitudes of the topset-foreset contact of ice-marginal deltas deposited in the lakes and of their spillways show a north-northwest trending profile (Figure 2) of uplift for this area that has a gradient of 4.7 to 4.9 ft/mi (0.89 to 0.93 m/km) up to the north. The north-northwest trend of the profile is an approximate best fit done by inspection of the data points; none of the projected profiles of the three lakes is long enough, and there too few altitudes of topset-foreset contacts of deltas of Lakes Tyngsboro and Hooksett, to allow a more precise determination of uplift direction at this time. The altitude of topset-foreset contacts shown in Figure 2 are believed to represent the lake level to within 3 ft (1 m), but the spillway floors are only the minimum lake altitude. A depth of water that varies in height from 4 to 8 ft (1.2 to 2.4 m), depending on the particular spillway, can be projected from the topset-foreset altitudes to the thresholds. The profiles shown in Figure 2 are based only on altitudes of noncollapsed parts of ice-marginal deltas. Many points that fall below these projections have been obtained from apparent topset-foreset contacts but these have been found to represent either collapsed topset-foreset contacts or tops of delta foreset beds that have been modified by later erosion. In the eroded deltas, later stream-terrace fluvial deposits overlie eroded foreset beds, mimicking topset-foreset contacts. As previously mentioned, these stream-terrace beds generally are finer grained than delta topset beds in the Merrimack Valley. Because the timing of the modification by erosion is not well known, data points from the eroded deltas can not be connected with any confidence and are not shown.

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A TRIP DOWN THE ALTON BAY FLOW LINE

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INTRODUCTION

The purposes of this trip are to obtain an overview of the glacial deposits of southeastern New Hampshire and southwestern Maine and to examine deposits critical to several classical controversies.

The glacial deposits have been described by many authors (Goldthwait et al., 1951; Tuttle, 1952; Bloom, 1960; Bradley, 1964; Smith, 1982; Thompson, 1982). A discontinuous mantle of till a few meters thick overlies bedrock. Characteristically the till is sandy and composed of clasts transported only a few kilometers from their source. Drumlins, often with bedrock cores (Birch and Trask, 1978), may include till over 50 m thick. The internal structure of these drumlins has not been studied nor has the "two till" problem. The till is overlain by stratified sand and gravel including ice-contact deposits and outwash. These form a large variety of landforms including kame terraces, eskers and outwash plains. Younger deposits include dune sands, swamp deposits and alluvium. Below the marine limit of about 70 m (Bradley, 1964) "marine clay" occurs above the till and below or interbedded with the stratified sands and gravels (Bradley, 1964). Subaqueous outwash, as mapped in coastal Maine (Smith, 1982), has not been described from New Hampshire.

Major controversies have concerned the mode of deglaciation (Koteff and Pessl, 1981). Was it normal retreat of active ice? Was it regional stagnation? Was it stagnation-zone retreat? In the coastal zone the controversy has been whether the retreat occurred on land (Goldthwait et al., 1951; Tuttle, 1952) or whether the glaciers calved into the sea (Lougee, 1940; Smith, 1982; Thompson, 1982). At the following sites evidence for these hypotheses can be viewed and discussed.

FARMINGTON, NEW HAMPSHIRE: STOPS 1 and 2.

Detailed study of stratified deposits indicates that here deglaciation was by stagnation-zone retreat (Coupland and Mayewski, 1980). The principal evidence is a branching set of flat-topped eskers and its relationship with kame terraces and the head of a collapsed outwash deposit (figure 1). The eskers were apparently open-ceilinged and recorded pulsating melt-water discharge. Southward narrowing of the Cocheco River valley may have contributed to glacier stagnation.

BERWICK, MAINE: STOP 3.

The Wentworth gravel pit lies along an extension of the "Newington Moraine". This controversial feature was initially interpreted as a marine end moraine (Katz and Kieth, 1917) but later authors disagreed (Tuttle, 1952; Bloom, 1960). Recently the general concept of marine end moraines has been well-accepted by many authors (Bradley, 1964; Smith, 1982; Thompson, 1982)

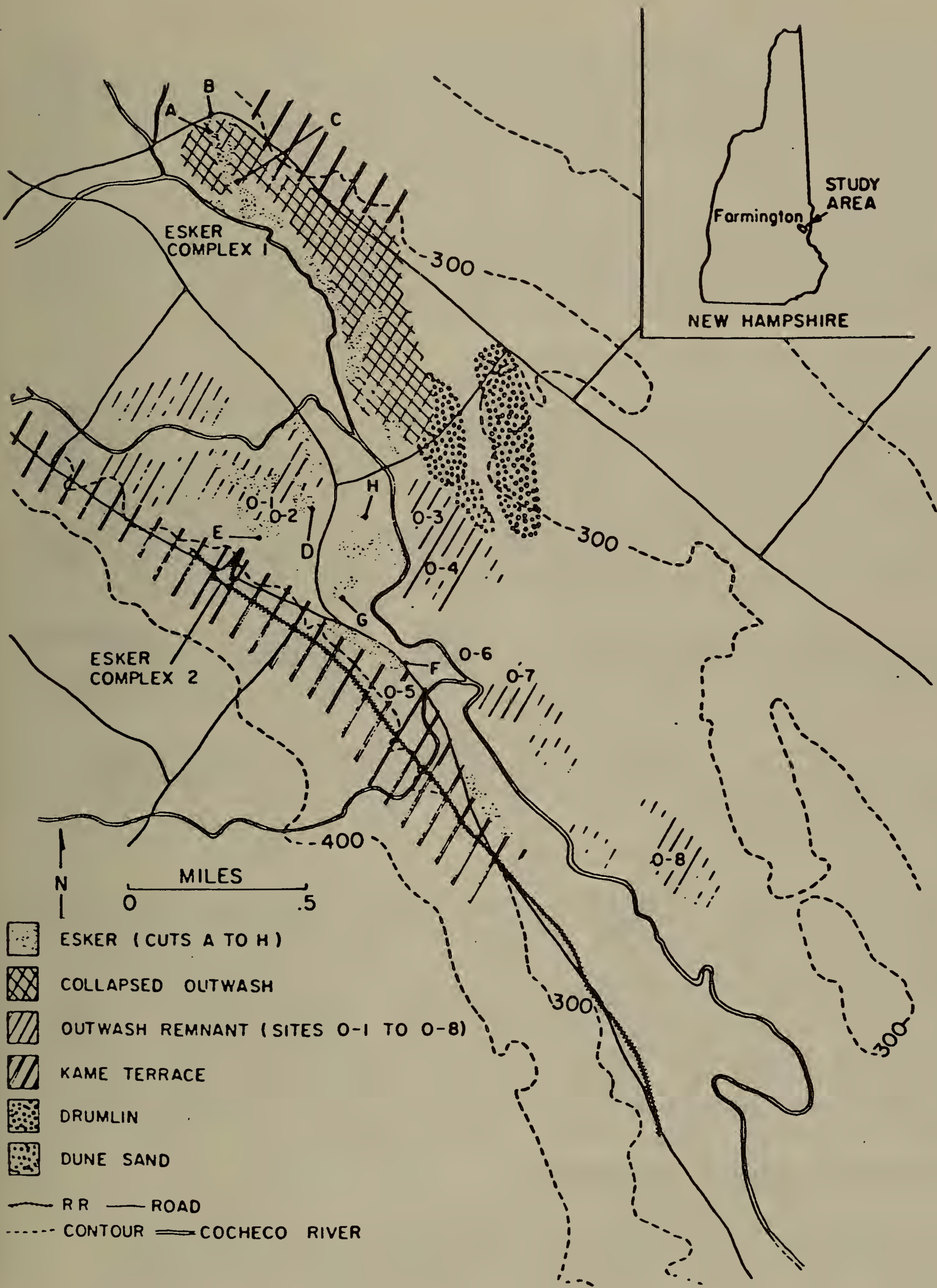

Figure 1. Esker complex in Farmington. Figure is from Coupland and Mayewski (1980) courtesy of Northeastern Geology.

although controversy surrounds individual features. In this pit marine clays are interbedded with stratified sand and gravel. Marine fossils are reputed to have been found here. Folding and thrust faulting of the deposits have been interpreted as signs of active ice during deglaciation.

ELIOT, MAINE: STOP 4.

Great Hill drumlin, as revealed by extensive excavations, is composed of massive till at the northwest end and laminated silt, sand and gravel at the southeast end. Here are an apparent wave-cut terrace and sea-cliff presumably marking the local marine limit at an elevation of about 61 m.

DOVER-MADBURY, NEW HAMPSHIRE: STOP 5.

The Barbadoes Pond deposit is now interpreted as an ice-marginal marine delta (Bradley, 1964; Hensley, 1978; Birch, 1980, Moore, 1982) in contrast to earlier interpretations as a stream or lake deposit (Goldthwait et al., 1951; Tuttle, 1952). Rows of similar deposits lie along bedrock strike ridges roughly parallel to the coast (Figure 2) (Birch, 1980) and mark retreatal positions of the ice sheet (Moore, 1982). The basic deltaic geometry has been confirmed by detailed seismic refraction studies (Birch, 1976; Hensley, 1978; Birch, 1980) and seaward paleocurrents have been mapped (Moore, 1982).

This deposit has been studied in exceptional detail because of hydrologic conflicts between landowners (shores of a small kettle lake), the City of Dover (municipal wells) and the City of Portsmouth (nearby reservoir) (Hall, 1976).

A major unresolved question is the location of the meltwater streams. Were they sub-, en- or supra-glacial? Published evidence is inconclusive (Moore, 1982; Birch, 1980).

EPPING, NEW HAMPSHIRE: STOP 6.

The "marine clays" of southeastern New Hampshire once formed the basis of an extensive till and brick industry (Chapman, 1950; Goldthwait, 1953). Despite the name, these deposits are primarily rock flour rather than clay minerals and contain large amounts of silt and sand as well as larger dropstones (Chapman, 1950; Bloom, 1960). They are interbedded with the ice-contact marine deltas (Bradley, 1964), evenly drape bedrock (Birch, in press) and correspond to the Presumpscot Formation of coastal Maine (Bloom, 1960). This formation has not been studied in any detail in terms of vertical or horizontal variability, age, fossils or paleomagnetism. A few foraminifera have been collected here (D.W. Collins, personal communication, 1984).

INNER CONTINENTAL SHELF, NEW HAMPSHIRE

Although not on the itinerary for this trip, the deposits offshore merit some description for comparison and contrast. Based on detailed geophysical surveys (Birch, in press) with modest sample control (Flight, 1972; Mills, 1977; Folger et al., 1975; Collins, in prep.; Brooks, in prep.) there are four major sedimentary units. The oldest is presumably till forming scattered patches, drumlins and a large flat-topped bank. This latter feature is unsampled and enigmatic. The next unit is a draped fine-

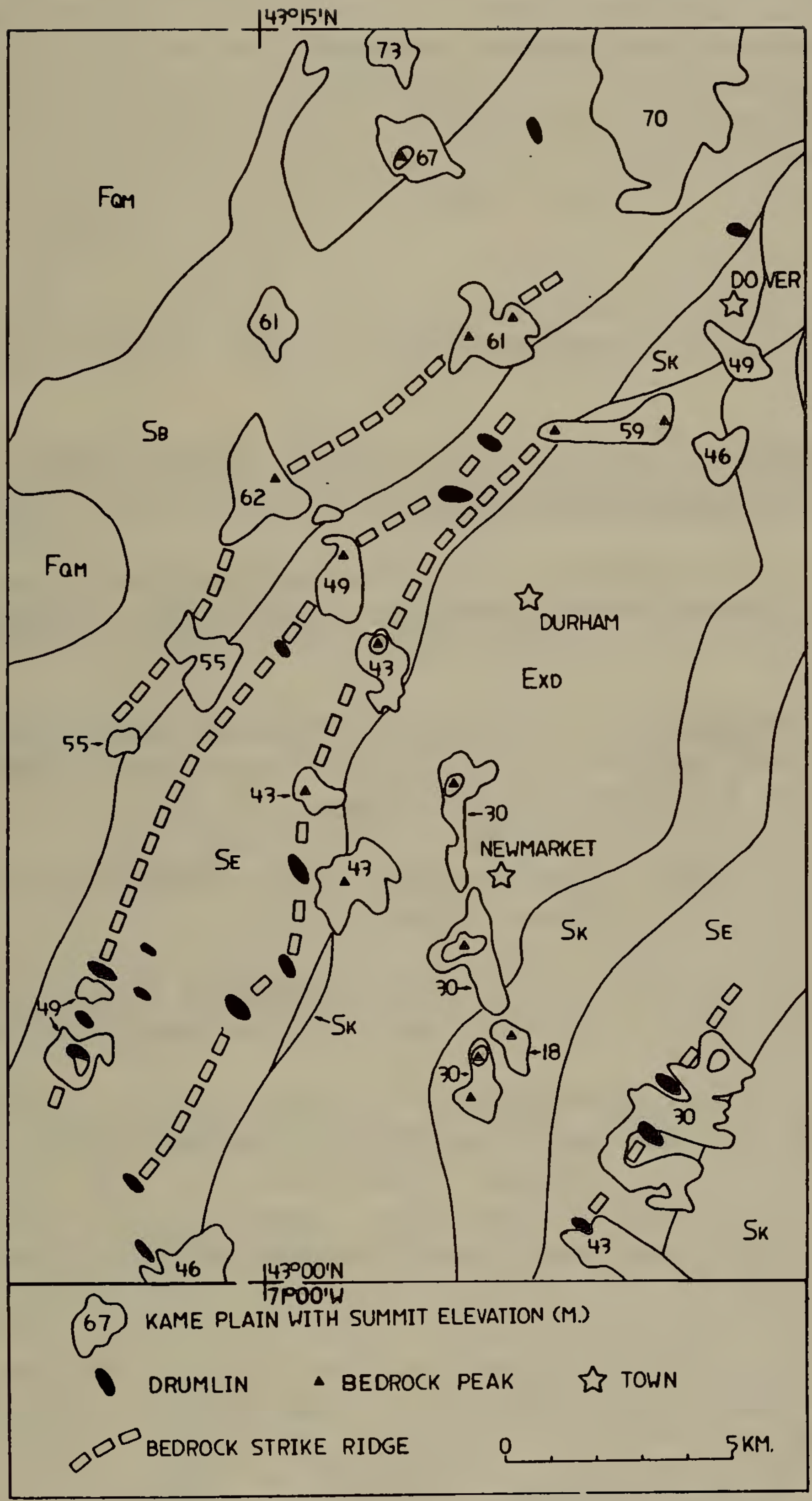

Figure 2. Relationship of kame plains (ice-contact marine deltas) and drumlins to bedrock. Bedrock units include: Fqm, Fitchburg quartz-monzonite; Se, Eliot Formation; Sb, Berwick Formation; Sk, Kittery Formation and Exd, Exeter diorite. Figure is from Birch (1980) courtesy of Northeastern Geology.

grained deposit correlative with the Presumpscot Formation. The central part of this unit contains slumps and debris flows possibly generated by an earthquake. Above the draped unit in deep water is a thin Holocene silt and clay and, at the mouth of the Merrimack River, a drowned sandy delta (Oldale et al., 1983). Finally, Holocene sands lie along the present beaches and also form mobile (?) sheets and mounds at depths of 20 to 30 m.

Conspicuously absent are clear signs of end moraines, ice-contact deposits or outwash.

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ROAD LOG

This road log includes only highway mileage, not any excursions onto side roads at the various stops. At some stops permission of owners may be required. Hard hats may be required also.

0 miles:	Start at parking lot for "Mt. Washington" cruises on Route 11 on the west side of the lake at Alton Bay. Proceed south on route 11 through Alton Bay towards Farmington.
14.4 miles:	Stop where railroad crosses Route 11. STOP 1. Turn around and proceed north on Route 11.
16.2 miles:	Stop at intersection of Routes 11 and 153. STOP 2. Turn around and proceed south on Route 11 into Rochester.
21.8 miles:	Bear left on Route 11 (Walnut Street) just after shopping mall. Continue into the center of Rochester: here Route 11 becomes Route 16. Continue on Route 16 to south.
23.1 miles:	Pass Frisbie Memorial Hospital.
25.3 miles:	Pass Skyhaven airport.
26.9 miles:	Turn left onto Route 16A. Continue into Somersworth.
29.0 miles:	Turn left onto High Street, Somersworth (Routes 9 and 16A south turn off to right here).
29.5 miles:	Cross Salmon Falls River, bear right and then left onto Routes 9 and 236. Continue east on Route 9 (School Street).
33.9 miles:	Turn right onto Wentworth Road.
34.8 miles:	Stop by white house (Mr. Wentworth) on left with barn across on right. STOP 3. Continue south on Wentworth Road.
35.4 miles:	Turn right on country road at intersection.
35.9 miles:	Cross railroad tracks and turn right onto Route 4. Proceed towards South Berwick.
38.2 miles:	In center of South Berwick turn left at intersection to continue on Route 4.
38.2 miles:	Turn left onto Route 236 towards Eliot and Kittery.
42.9 miles:	Turn left onto Route 101 and proceed south.
44.7 miles:	Stop with Great Hill on left. STOP 4 Turn around and proceed north on Route 101.

46.5 miles:	Cross route 236. Continue on Route 101 into center of Dover.
50.1 miles:	At intersection turn left into Dover.
50.4 miles:	Bear right on South Portland Street.
50.6 miles:	Turn right on Saint John's Street.
50.65 miles:	Turn left onto Route 9 (Winter Street).
50.7 miles:	Turn left on Route 9 (Central Avenue).
51.3 miles:	Turn right on Route 9 (Silver Street).
52.1 miles:	Bear left following Route 9. Crossover Spaulding Turnpike.
52.6 miles:	Turn right following Route 9.
55.0 miles:	Stop at Iafolla gravel pit. STOP 5. Continue west on Route 9.
55.8 miles:	Cross Bellamy River reservoir.
58.6 miles:	Turn left onto Route 125. Proceed south towards Epping.
63.1 miles:	Go half-way around traffic circle and continue south on Route 125 towards Epping.
72.4 miles:	Stop at abandoned Sunoco gas station in Epping. Clay pits are behind W.S. Goodrich store on east side of Route 125. STOP 6.

FIELD TRIP C8

SURFICIAL GEOLOGY AND ARCHAEOLOGY ON THOMPSON ISLAND, BOSTON HARBOR, MASSACHUSETTS

PART 1 SURFICIAL GEOLOGY OF THOMPSON ISLAND

TRIP LEADER: D.W. CALDWELL

PART 2. ARCHAEOLOGY ON THOMPSON ISLAND

TRIP LEADER: RUSSELL BARBER

SURFICIAL GEOLOGY OF THOMPSON ISLAND

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INTRODUCTION

Thompson Island is one of the larger of the Harbor Islands and is in the Dorchester Bay, in the southeastern part of Boston Harbor. A long spit exposed at low tide on the southeastern end of the island comes within a few meters of Squaw Rocks in the Squantum section of Quincy. The next nearest land is Spectacle Island, about 1 km east of Thompson. The island is about 1.7km long and less than 0.5km wide. Thompson Island is reached by motor launch from Kelley's Landing in South Boston, about 2 km to the northwest.

Barber (this volume) has described the previous habitation and use of the island. The Thompson Island Educational Authority currently operates the island as a resourse for the City of Boston.

GLACIAL DEPOSITS

In the basis of its glacial deposits Thompson Island may be divided into two parts: 1) the north end or upper island consists of the remains of a drumlin and 2) the south end or lower island is underlain by highy collapsed outwash. A map of the surficial deposits of the island is shown in Figure 1.

Drumlin Orientation

The northern half of the island consists mostly of the remains of a drumlin. The very northeastern end is composed of silty gravel overlying glacial-marine clay. The orientation of the upper island appears to be more the result of wave erosion than of ice flow direction and original drumlin shape. Kaye (1976) describes drumlin orientation (and related ice-flow direction) in the Boston area as ranging from east, 135 ^{o} around to the southwest. It is the writer's opinion the the northeast-southwest orientation of Thompson Island is mostly the result of post-glacial marine processes. Indeed, the nearest uneroded drumlins on land (Savin Hill, Telegraph Hill, Boston South Quadrangle) are oriented between east and southeast.

Drumlin Till

Drumlin till is exposed along both the north and east sides of the upper island. The till here is uniformly oxidized to a reddish brown color throughout the 10 m-high exposures. This confirms the observation of Schafer and Hartshorn (1965) that drumlin till in eastern Massachusetts is commonly weathered to a depth of 4-9m and may reach as much as 20m. They also conclude that the drumlin till is the so-called older till of southern New England (Drift III of Kaye (1961) in the Boston Basin). It the opinion of Schafer and Hartshorn (1965) that the drumlin till is early to middle Wisconsinan in age and it may be Illinoian in age. This suggests that the drumlins in the Boston area were either deposited prior to the late Wisconsinan or were eroded out of older till during the late Wisconsinan glaciation.

Figure 1. Surficial geology of Thompson Island. Qt refers to till, Qg indicates sand and gravel deposits, Qb denotes beach deposits, Qs are swamp deposits.

Figure 2. Coastal features near Thompson Island and locations of field trip stops and meeting place. From Coastal Chart 13270 (NOAA).

Outwash Deposits of the Lower Island

The southern portion of Thompson Island consists of well-washed sand and gravel. One ephemeral exposure on the southeast shore displays deltaic cross bedding on occasion. Collapsed bedding was observed in a pit on the northwest shore. In the interior of the lower island are numerous kettles and knobs and the estuary on the southwest end of the island may be a breached large kettle.

There is marine clay on the northern end of the island and it is found elsewhere in the Boston area up to 15m above present sea level (Kaye, 1976). The topset-foreset elevation of deltaic sediments in the lower island is about 4m. These data suggest that the deltaic outwash on Thompson Island are glacio-marine in origin. The higher sea level indicators in the northern part of the Boston area are the result of the greater isostatic rebound of a single water plane in the north than in the Thompson Island area (Stone and Peper, 1982). This would mean that Thompson Island has never been submerged to a level higher than about 4m above the present sea level since the late Wisconsinan.

Lithology of the Drift

Clasts in the till on Thompson Island were derived in large part from Boston Basin rocks. In decreasing abundance, the rock types on the island are the Cambridge Argillite, the Brookline (Roxbury) Conglomerate, and the Squantum "tillite". These rocks have recently been shown to be Precambrian in age (Late Proterozoic Z) by Lenk, et al. (1982). Crystalline rocks from outside the Basin make up most of the other clasts in the drift.

COASTAL GEOLOGY OF THOMPSON ISLAND

Following the deposition of the outwash on the Thompson Island, sea level fell, relative to the island, to a low of 20m below present sea level around 10,000 years B.P. (Kaye and Barghorn, 1964). The sea then gradually rose, reaching its present level about 2000 years ago (Kaye, 1976). The bathymetry east, south, and west of the island suggest that at some lower than present sea level, the sand and gravel deposits in this area formed a continuous plain, perhaps with the ancestral Neponset River cutting into it (see figure 2). The rising sea level gradually submerged this plain, perhaps while some higher kames were being eroded. Seismic profiling in other parts of Boston Harbor and to the east, in Massachusetts Bay, indicates that many drumlins were submerged by the rising sea level (Kaye, 1976).

The initial configuration of Thompson Island, 2000 years ago, may have consisted of a drumlin island and a separate island of outwash, both much larger than at present. With the stongest wave attack coming from the north-east, it appears that the drumlin offered some protection for the outwash, perhaps preventing it from being completely eroded. Eventually the two island were connected by tombolos, with marsh and overwash deposits filling the area between the two.

Shoreline Character and Glacial Deposits

The shoreline of Thompson Island bears an obvious relationship to the glacial deposits. On the upper island, where till is undergoing active erosion, the shore consists of striated boulders and cobbles. On the lower island, continuous sandy beaches and spits occur. Near the northeast end of the island, a cuspate spit has been constructed from the silty sand and gravel which occurs in that area.

For a complete discussion of the coastal processes of Thompson Island see Rosen (this volume).

EROSION DURING THE GREAT BLIZZARD OF FEBRUARY 6 AND 7, 1978

Fetch

The greatest fetch effecting Thompson Island is 2.5 nautical miles northeastward from the island toward Deer Island and Shirley Gut. During the blizzard of February, 1978, the reported 48 knot winds would produce waves more than 1 meter in height (FitzGerald, personal communication). These were apparently the largest waves to effect the island in nearly 100 years.

Prevailing and Domiant Winds

Northwest winds prevail on the island, but are somewhat limited in their effect by the fetch in that direction of about 1 nautical mile. The predominant winds are from the northeast.

Tide and Storm Surge

The height of the storm conditions of the blizzard of 1978 coincided with the new moon spring tide. The northeast winds outside of Boston Harbor reached nearly 90 knots , which produced a storm surge of 1.3m within the harbor which was added to the spring high tide. Eyewitness accounts indicate that the wharf on the island was completely submerged.

Erosion Estimates on the Island

From a comparison of pre-storm photographs with those taken after the storm, it appears that little erosion occurred on the south, southwest and west shores. Increasing amounts of erosion occurred from west to east along the steep north-facing cliff of the drumlin. The pelting snow and rain saturated the till, causing slump failures to occur, the slumps turning into turf-covered mudflows at the base of the cliff. The pounding surf removed the toes of the flows and winnowed fines from the till. As much as a meter of till was eroded from this part of the island (Caldwell, 1978). On the far eastern end of the island, in the very teeth of the storm, up to 2m of sand, gravel, and clay were lost, much of the coarser debris ending in an overwash fan 10m south of the eastern tip. The erosion here exposed an ancient shell midden (see Barber, this volume).

The greatest erosion during this storm occured on the banks of sand and gravel on the southeast side of the island. About 100m south of the marsh in the center of the island, these banks rise steadily from less than 1m to over 7m in height near their south end. Erosion esimates here are based on the coherent blankets of sod that rested at the base of the sand banks after the storm and

Figure 3. Blankets of sod resting on beach after blizzard of 1978. These masses of sod, up to 3 meters wide, are a measure of the amount of erosion during the storm.

ranged from 1 to over 3m in width. The maximum erosion occured where the banks are between 3 and 4m in height. The erosion is thought to have occurred in the following manner: during the height of the storm, the combined astonomical high tide and storm surge nearly submerged the 3 to 4m bank; the movement of the water in swells caused the sand in the banks to liquify, flowing out on the beach; as the sod was undermined, the sod floated; as the water level lowered after the peak of the storm, the sod was lowered onto the liquified sand (Caldwell, 1978). Because only one layer of sod was on the beach after the storm, it is possible that the 1 -3m of erosion occurred in only a few hours during the high water within the harbor. It is also possible that wave attack was not the principle energy source involved in the erosion, but liquifaction and gravity. Other accounts of erosion and coastal processes during this storm are found in FitzGerald (this volume).

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Field trip log appears at end of accompanying article by Barber, trip C8, part2

ARCHAEOLOGY ON THOMPSON ISLAND

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Introduction

Thompson Island is an inner island of the Boston Harbor group, and its position near the mainland shielded it from early European exploration. While the Boston Harbor islands in general were first described somewhat earlier, our first record of European exploration of Thompson Island occurs only in 1621 with Miles Standish. This date has twofold significance, for it marks both the end of the era of prehistoric occupation of Thompson Island by Native American peoples and the beginning of the era of historic occupation of the island and surrounding areas by European and other Old World peoples.

The prehistoric era, by definition, is marked by the absence of written records to aid in reconstructing the ways of life and significant events in the lives of the island's early occupants, so archaeology, by default, must shoulder the entire burden. Both surveying to locate sites and excavating to recover artifacts, food remains, and other information can provide insights into these past ways of life.

The history of prehistoric research on Thompson Island is a short one. While Euro-American settlers no doubt collected Indian relics from their earliest excursions to the island onward, most of these haphazardly collected items have been scattered since, and their scientific value is lost. The owners of the island made an effort to collect many of these items along with locational information, but artifacts and notes were lost in a tragic fire in 1971.

Around 1970, scholarly interest in the prehistory of Thompson Island began coalescing into field research. This island, along with the others in Boston Harbor, was surveyed briefly by Dincauze (1974), and a more in-depth program of survey and excavation was undertaken by Barber (1983) and Shaw (1984). The latter research is ongoing.

But archaeology does not disappear with the historic era. Historic documents tell much about the people they describe, but they are subject to the biases and assumptions of their authors. Many segments of society are rarely mentioned, and reality may be distorted severely. Historical archaeology, utilizing the techniques of both archaeology and history, tries to produce a more comprehensive picture of the historic past.

Historical archaeology is a recent development, and its application on Thompson Island is very recent. The first such study was begun in 1984 (Beaudry 1984) and consists of a basic inventory of known historic sites; its results were not available at the time of writing.

Synopsis of Thompson Island Archaeology

Prehistoric Era

The earliest human settlement known for New England dates to 10,500 B.C., the period immediately following deglaciation; this is known as the Paleo-Indian period. Although their contemporaries in western North America consumed Pleistocene megafauna as a part of their diet, Paleo-Indians in the East ate a broad range of foods, but megafauna rarely were part of their diet. No traces of these early peoples have been found on the Boston Harbor islands, probably because marine transgression has drowned the sites of their camps.

Following the Paleo-Indian period came a long period known as the Archaic period (8000-1000 B.C.). It is during this period that the first evidence of Indian occupation is found on several islands in Boston Harbor, but the earliest traces so far known on Thompson Island date back only to about 4000-3000 B.C. These peoples collected their food solely by hunting and gathering, using the islands as regular stops on a seasonal cycle. They used tools of stone and native copper but did not yet use ceramic pottery. In the last two or three thousand years of the Archaic period, when marine transgression flooded the lowlands with salt water and transformed hilltops into islands, they made extensive use of marine resources, including finfish, shellfish, and sea birds.

The Woodland period (1000 B.C.-A.D. 1600) was marked by the development of ceramic pottery, in contrast with the steatite pottery that had been in use for the preceding 1300 years. This development may not have had tremendous effects on the lives of these Indians -- they continued hunting and gathering and seasonal movements -- but the presence of ceramic sherds makes a convenient marker for prehistorians trying to date a site. Partly because of probably growing populations, partly because of drowning of earlier, riverside settlements, Woodland sites are most numerous on Thompson Island and its neighbors.

Around A.D. 1000 a quiet revolution took place. At about this time, corn, beans, and squash (all introduced to New England from points south and west) were first cultivated in southern New England. There were no overnight transformations of society with the growing of crops, but there were subtle shifts toward larger, probably more permanent settlements, especially along fertile floodplains. An important question current in New England prehistory is whether the presence of abundant coastal resources discouraged the adoption of cultivation in the coastal zone. At present there is no evidence that crops were grown on any of the Boston Harbor islands (Luedtke 1980).

The period between the first European visits to New England in the mid-sixteenth century and the end of the aboriginal way of life in the middle to late seventeenth century belongs to both the prehistoric and historic eras. For convenience, it is frequently called "the Contact period." From around 1550 onward Europeans carried on a thriving trade, especially for the furs that had become so scarce and expensive in Europe; this period is represented on Thompson Island by a trading post. Eventually disease and warfare would kill most New England Indians and force the rest to leave for areas to the west and north or to pursue lifestyles more like the dominant Euro-Americans.

Historic Era

Although the presence of Europeans in New England before Columbus is widely accepted, there is very little archaeological support for it. Stone chambers variously ascribed to Ibero-Celts or Phoenecians have been shown to be nineteenth-century root cellars (Neudorfer 1979); supposed Ogham writing of early Irish monks appears to be glacial striae and plow scars; "sacrificial altars" appear to be the bases to colonial cider presses. Unfortunately, the considerable time investment necessary to refute these wild claims has discouraged most qualified archaeologists from discussing the issue in print; as a result, the mass of shoddy claims and data probably would obscure a legitimate claim, if it did exist.

The Norse case stands in contrast. The L'Anse aux Meadows site in Newfoundland (Instad 1977 and others) is almost universally accepted as Vinland now, and a Norse coin from good context is known from Maine. While there are no known Norse remains from Boston Harbor and the various purported runes, towers, and "mooring holes" from southern New England all appear to be either recent productions or erosional phenomena, there is a possibility that evidence of the Norse may someday be found in Boston Harbor.

The history of Thompson Island after Columbus may be divided into three periods: the frontier period (1626-1634), the farming period (1634-1833), and the institutional period (1833-present). (These periods have been devised without the results of Beaudry's ongoing research and doubtless will be refined and reformulated by her.)

The frontier period begins with the establishment of a trading post on Thompson Island by David Thompson in 1626. Thompson died in 1628, and is unclear how long the post may have been managed by his son, John, but it appears to have closed by 1630 (Thompson and Thompson 1979). During its operation, the post attracted both Indians and colonists for trade.

The farming period began with the acquisition of Thompson Island by Dorchester in 1634. Documentary and archaeological evidence both suggest continuous occupation or tillage from that date forward until the establishment of the Boston Asylum and Farm School on the island in 1833. Since that time forward, the island always has been devoted to education.

Conclusion

Archaeology is not geology, of course, but the presence of this discussion in a geological guidebook attests to the overlap of interests. Geologists, especially glacial and surficial specialists, share many common concerns with archaeologists in terms of sediment deposition. Each can materially aid the other: geologists can provide a picture of regional or local patterns and archaeologists frequently can provide precise dating for deposits in restricted areas. This research partnership has already proven its value in various studies.

It also behooves the archaeologist to alert geologists and others of the fragility of archaeological sites, especially on eroding islands. The Thompson Island Education Center, managers of Thompson Island, have acted on their concerns for archaeological sites on the island and solicit everyone's cooperation in preserving and reporting newly eroding sites to them.

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FIELD TRIP LOG

Assemble at The Inn at Danvers at 9:00 AM. Turn right onto U.S. 1. Drive south on Route 1; about 1.5 miles I-95 joins US 1; after about 2 miles, I-95 turns right, continue south on US 1. After about 10 miles, toll booths for Mistic River Bridge, toll $ .50; after decending from bridge, road bears left up incline, get into left lane for left turn at top of incline and merge with I-93. Continue south on I-93 through downdown Boston. After about 1 mile, enter short (South Station) tunnel; continue south on I-93. Note turn onto Massachussetts Turnpike; continue south on I-93 (Southeast Expressway) in right hand lanes. Bear right at Columbia Point, Univerity Massachusetts-Boston, etc. Proceed to stop sign and turn left under Expressway. Enter traffic circle at 6 o'clock, leave at 11 o'clock onto John Day Boulevard, water now on your right. Continue past bath houses (L-Street Bath House is home of L-Street Brownies, who swim in Bay all winter) and yacht clubs to end of residential area which is on your left. Kelley's Landing is green structure on right, with dock behind building. Park along street beyond landing. Kelley's Landing and Thompson Island are shown on the Boston South Quadrangle.

11:00 AM: Leave Kelley's Landing on Thompson Island launch

Stop 1. Warf on Thompson Island. Note updrift acumulation on north side of warf abutment,which acts as a groin. Principle longshore transport is southwestward, material being derived from eroded drumlin. Field trip stops are shown on Figure 2. Proceed about island in a clockwise direction. Total walk about 2 miles. Refer to Barber (this trip) for archaeological descriptions of a number of sites along this route.
Notice increase in size of clasts as we proceed northeastward.

Stop 2. North end of island. Exposure of drumiln till. Note turf-banked mudflows. Till is oxidized throughout the exposure is may be either early Wisconsinan or Illinoisian in age. See Barber, location 1.

Stop 3. Northeast end of island. Silty shingle gravel overlying glacial marine clay. Shell midden here was first exposed during storm of February, 1978. Erosion during that storm was from northeast wind-driven waves. Some eroded material was deposited in a now overgrown overwash fan 10m south of point.
See Barber, location 2.

Stop 4. Cuspate spit. Material was transported from the north, from the deposits seen in the last stop, and from the east shore of the island, carried by currents formed by winds from the southwest. Note the small chips of Cambridge material oriented vertically in small nests on the berm. Any theories as to origin?

Note eroding salt marsh peat on beach near center of island. See Barber, location 3.

Stop 5. Outwash sands and gravel. One exposure has displayed topset-foreset bedding upon occasion. Coherent blankets of sod resting on beach here after storm indicate as much as 3 meters of erosion occurred during the blizzard of 1978. Longshore transport here, toward the southwest, driven by refracted northeast winds has resulted in long spit built to the southeast. This spit is also contructed by currents coming from the west, driven by southwesterly and northwesterly winds. See Barber, locations 4 and 5.

Stop 6. Kettled topography near center of lower island. The numeber of kettles in this area suggest that this may represent the stagnation zone of the retreating glacier, with the deltaic outwash being the proglacial portion of the morphosequence. The estuary to the south may be a breached kettle (note that no estuary existed when Boston South quadrangle was surveyed (figure 1) but that it is shown in the Boston Harbor coastal chart (figure 2).

Optional Stop 7. Tidal estuary on southwestern end of island.

Return to wharf and departure for Kelley's Landing. See Barber, location 6 on way to wharf. Thank you very much.

Locations of Archaeological Interest

The locations noted below do not constitute an exhaustive catalogue of archaeological sites or other points of interest on Thompson Island. Rather, they are representative of various areas on the island. The locations described here are numbered according to Fig. 1.

1 -- Erosion area: Such eroded areas frequently reveal broken shell or other indications of an archaeological site. As erosion cuts away the bank, it aids in the discovery of sites; at the same time, it progressively destroys the site. Because human settlements almost always have different areas reserved for different activities, the destroyed portions may be unique and critical to understanding the function of a site.

2 -- Site HL-5: This site demonstrates how rapidly erosion can destroy a site. In the 1920s, a thick deposit of shell was present, and a human skeleton eroded from the bank; by 1970, vestiges of a severely eroded shell deposit remained; by 1982, the deposit was totally gone. Looking at the eastward-facing bank, one can see an eroding deposit of black soil with some admixed shell near the top of the bank. This deposit is the remains of a trash pit that received the remains from cleaning fish (sturgeon and cod) and preparing soft shell clams; despite subsurface testing further inland, it is the last known trace of HL-5. On the basis of material recovered from the trash pit, it appears that HL-5 was occupied at least in late summer, probably after about A.D. 200. We will never know the full range of activities that took place there.

3 -- Clay deposits: Fragments of misfired and discarded ceramic pottery (that never would have been carried there from the mainland) have been found on Thompson Island and other islands in Boston Harbor, so pottery manufacture must have taken place there. All of the necessary ingredients were available: clay from deposits such as this one eroding in the bank and rigid material to use as temper to minimize breakage during firing. Most early prehistoric ceramics in the Northeast were tempered with sand or crushed rock ("grit"), while later peoples tended to use crushed shell as temper. On Thompson Island, as in other coastal areas, however, shell temper is moderately common in earlier ceramics.

Fig. 1. Archaeological locations discussed, Thompson Island.

4 -- Thompson's trading post: This location, according to oral history and secondary accounts, was where David Thompson's trading post was built in 1626. The post is supposed to have been a substantial log house with a full, brick-paved cellar. In 1889, a foundation began eroding out of the bank at this point, but our only description of the ruins (recorded in Thompson and Thompson 1979:129) gives precious little detail; we do know, however, that white clay pipe stems and bricks were found. Again in 1937, as reported in The Boston Globe, a foundation eroded out of the bank and was attributed to Thompson. This time, bricks were measured, and their dimensions are consistent with seventeenth-century bricks. Were these foundations part of the same structure? Was either part of Thompson's trading post? Certainly not all of the accounts of the post are believable, since Charles Evans, recalling his boyhood as a student in the early years of the Farm School, credited a barn still standing as Thompson's; this claim would make the barn over two centuries old.

5 -- Site HL-15: The shell exposed in the bank is the landward portion of a shell midden that once was much larger. Testing in the field above shows that the site extends only a meter or two from the bank, and in a few years this site will be destroyed completely. Pottery and stone tools from this site are few, as typical at shell middens such as this one, but indicate an Early Woodland occupation, probably around A.D. 200. Bones and shell from this site indicate that soft shell clam, white tail deer, wild turkey, cormorant, rabbit, turtle, and unidentified fish were being eaten by the site's inhabitants. As is usually the case, shellfish remains were the most prominent, but the high ratio of shell to meat in a clam makes this prominence misleading: shellfish were only a minor contributor to the diet. Prehistoric archaeological findings on Thompson Island confirm early accounts stating that the Boston Harbor islands were once wooded and abounded with game. The stone tools and waste from their manufacture are made mostly of felsites from the southern shore of Boston Bay, although some use was made of inferior local beach cobbles and some stone was imported from as far away as Pennsylvania. In earlier periods (before about 2000 B.C.), the islands' inhabitants seem to have relied more heavily on stone from the northern shore of Boston Bay.

6 -- Barge wreck: This twentieth-century barge wreck is a reminder that archaeological sites are still being made. Whether or note this wreck survives to become part of the archaeological record depends on the complex interplay of geology, materials preservation, and human intervention.

ROCK LITHOLOGY AND GLACIAL TRANSPORT SOUTHEAST OF BOSTON

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Introduction

The purpose of this field trip is to examine some of the rocks in the southern part of the Boston Basin and the "granites" south of the basin. South of the granites there are no coastal exposures of any bedrock until south of Cape Cod.

The boulders found in the cliffs and drumlins and on the beaches from Scituate to Cape Cod reflect the unique lithologies of the rocks to the North and West. By tracing these lithologies the minimum distances of glacial transport of these boulders can be established. In addition, by examining the lithology of the cobbles and pebbles on the local beaches, minimum transport of ice and waves can be ascrtained.

Since the tide is the controlling factor for this field trip participants should plan on leaving Danvers by 7 a.m.

Regional Geology

The Boston Basin is a low lying structurally complex synclinorium bounded on three sides by thrust faults. The Boston Bay Group includes most of the bedrock found in the Boston Basin. A good description of the rocks in the Boston area is given by Billings (1976, 1982) on whom the following is based. Usually, the Boston Bay Group is divided into two formations, the Cambridge Argillite which is coeval and partly overlies the Roxbury Conglomerate. The Roxbury Conglomerate has been divided into three units which are, in ascending order, the Brookline Member, the Dorchester Member and the Squantum Member (Emerson, 1917, La Forge, 1932). The Brookline member has a maximum thickness of 4300 ft but thins rapidly to less than 500 feet at the southern end of the basin. The clasts in the Brookline Member are predominantly quartzite, quartz monzonite, granite and felsite. The pebbles are well rounded and range in size from one to fifteen centimeters in size. The matrix is a gray feldspathic fine to medium grained sandstone. The color can range from white to pink to red. Interbedded in the Conglomerate are also argillites. These argillites are really siltstones or mudstones. They are laminated (0.05-0.4")(0.13-1.02cm) but do not possess the papery or platy splitting characteristics of shales. They do however, split into flags or slabs (0.4-24") (1-60cm) thick. The argillite is typically red, pink or gray-green. The Brookline member of the Roxbury Conglomerate averages 60 percent conglomerate, 20 percent sandstone and 20 percent argillite.

The Dorchester Member of the Roxbury is very similar to the Brookline Member except that the percentages are different. The Dorchester Member is 15 percent conglomerate, 25 percent sandstone and 60 percent argillite. There are also some purplish gray and

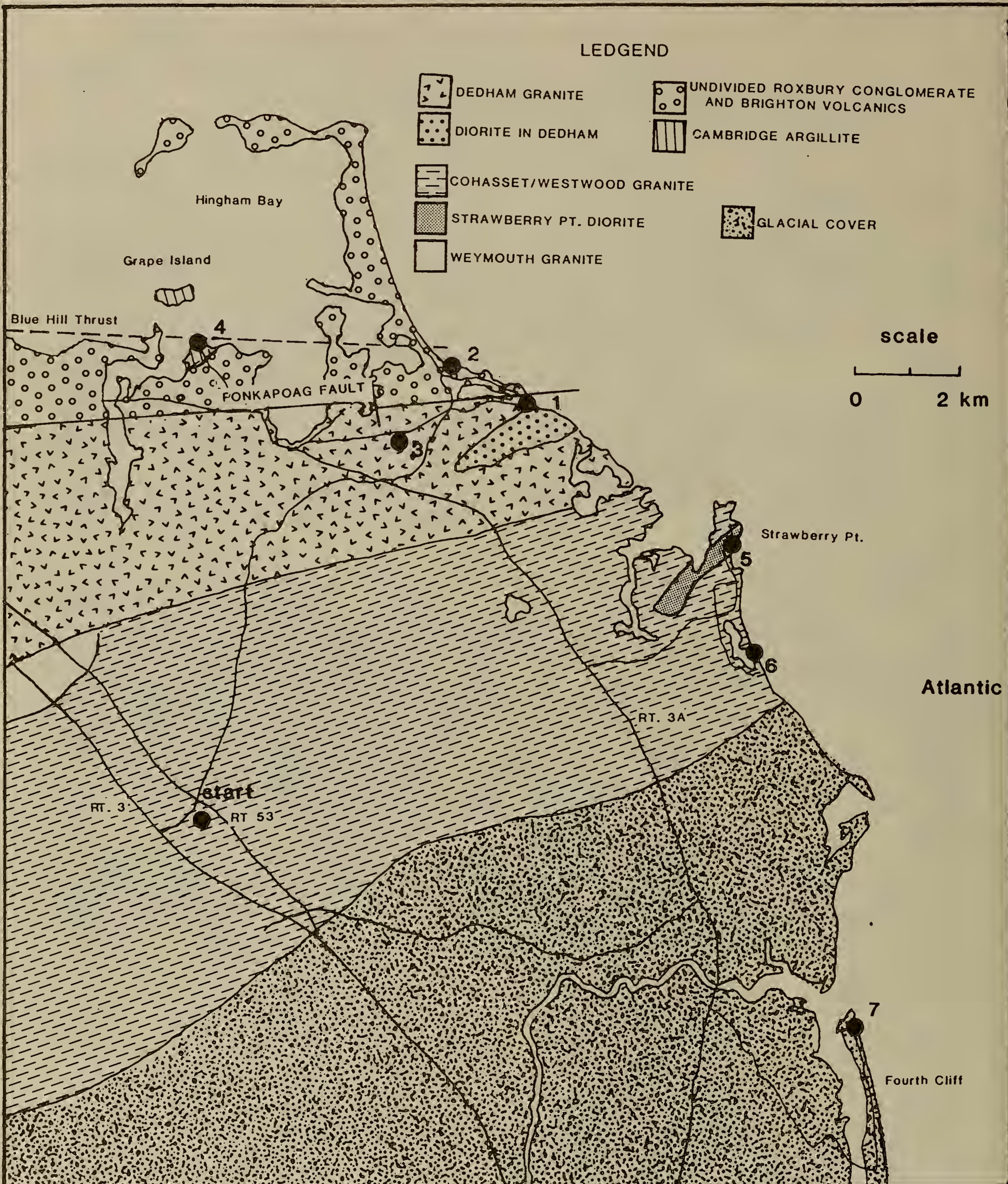
LEDGEND
DEDHAM GRANITE
DIORITE IN DEDHAM
COHASSET/WESTWOOD GRANITE
STRAWBERRY PT. DIORITE
WEYMOUTH GRANITE
UNDIVIDED ROXBURY CONGLOMERATE AND BRIGHTON VOLCANICS
CAMBRIDGE ARGILLITE
GLACIAL COVER
scale
0
2 km
Hingham Bay
Grape Island
Blue Hill Thrust
PONKAPOAG FAULT
Strawberry Pt.
Atlantic
RT. 3A
start
RT. 3
RT 53
Fourth Cliff
1
2
3
4
5
6
7

FIGURE 1

greenish gray colors present in the argillite. The Dorchester Member is about 1000 ft thick.

The Squantum Member is only between 70 to 400 ft thick. Many geologists consider the Squantum a tillite. It has subrounded to angular clasts from two inches to three feet in size of granite, quartz monzonite, quarzite, felsite and "melaphyre".

The Cambridge Argillite is never well exposed. It consists almost exclusively of gray argillite in which the beds range in thickness from 0.05 to 3 inches. The differences in their gray color is grain size dependent. The larger the size, coarse silt to fine sand, the lighter gray the color. The smaller the size, clay and fine silt, the darker is the gray color. Graded bedding is rare in the argillites. Many of the beds show a rhythmic layering due to the alteration of lighter and darker colors.

Within the Cambridge Argillite is a hard, white sericitic quartizte which is 400-500 ft thick. This has been named "Milton Quartzite" by Billings in 1929. The quartzite is visible for about two miles in Quincy.

The Boston Bay Group has recently been dated as Proterozoic Z, the late Precambrian, by microfossils from the Cambridge Argillite located north of Harvard Sqaure. (Lenk and others, 1982).

Scattered throughout the Boston Basin are the Mattapan/Brighton Volcanics. These are hard, dense white, pink, and red rhyolites. Also included are "melaphyres" (which are altered basalts and andesites) which are dark to light green and are composed chiefly of secondary minerals albite, hornblende chlorite and epidote. Recently Kaye and Zartman (1980) obtained a Pb-U date of 602 +/- 3 m.y. from analysis of zircon.

The Dedham Granite underlies most of the area south of the basin which is separated by the Ponkapoag Fault (see figure 1). This unit is more of a cartographic unit than a lithologic one (La Forge, 1932). It includes several "minor" units that Emerson thought were related to the Dedham. In the past thirty years, various attempts have been made to distinguish these other varieties from the Dedham. Chute (1966) describes the Westwood Granite as a younger, fine grained granite that intruded the Dedham; also a variety of ages has been reported from various localities in the unit, leading some to speculate as to whether or not there is more than one rock type present here (Zartman and Naylor, 1984).

To answer this question we have begun detailed chemical and petrological analysis in the Cohasset-Hingham area. This initial work indicates that there may be more than one magmatic series within the Dedham and that these units may be of different ages.

In the Cohasset area the "typical" Dedham Granite extends from the Ponkapoag Fault south appoximately four miles, where it comes in contact with a second type of granite. This contact is located somewhere within an extensive mylonite zone which parallels the Ponkapoag Fault Zone. The actual contact between these two granites has not yet been located. The Dedham Granite defined by Emerson (1917), is a coarse biotite granodiorite, light greenish white to dark grey, sometimes pink where microcline is plentiful. The quartz is vitreous and stained pale green. In

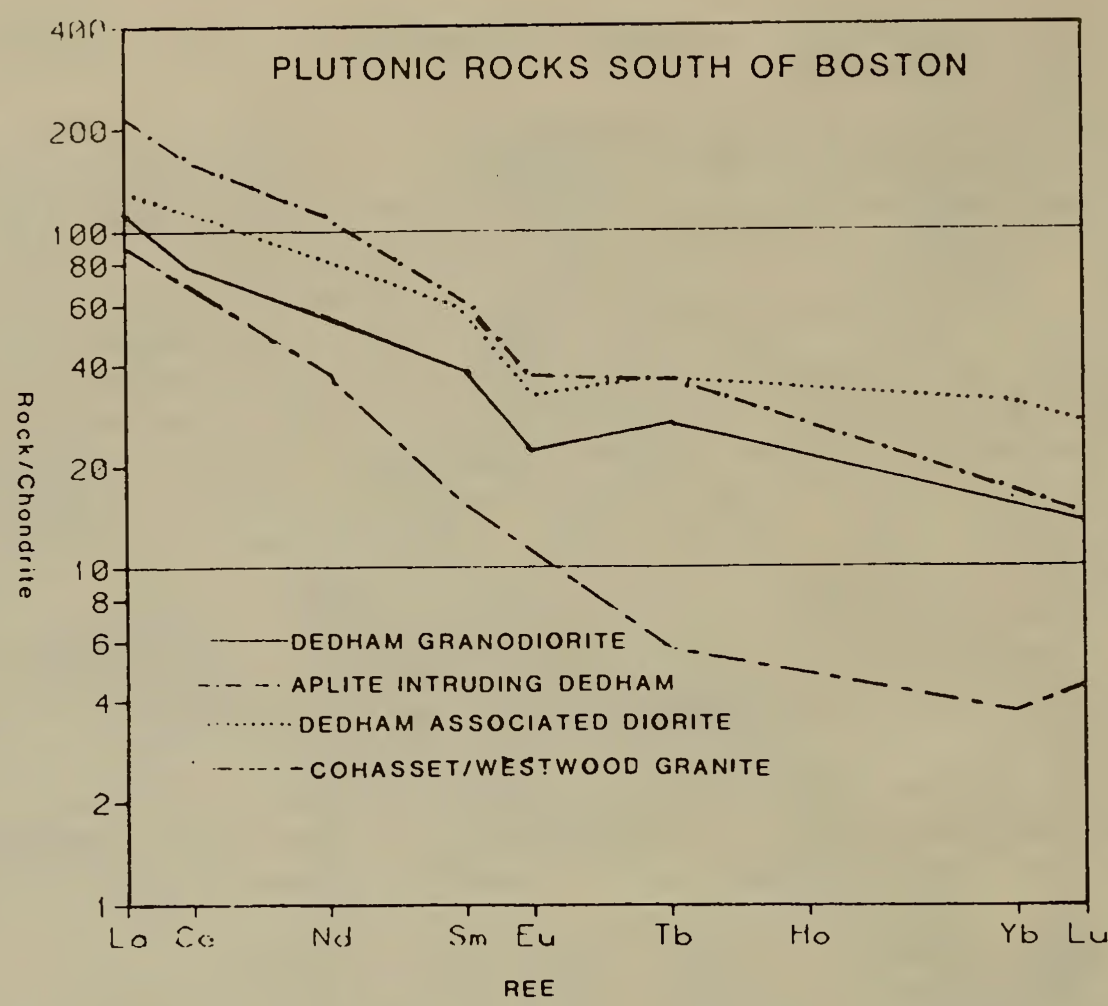

FIGURE 2

Major Elements of the Dedham and the Cohasset/Westwood Granites

	Dedham			Coh./Westwood	
	PD-1	PD-8	PD-7	PD-10	PD-4
SiO_2	66.48	58.57	60.21	71.80	55.63
TiO_2	0.58	1.21	0.34	0.34	1.56
Al_2O_2	15.74	17.36	18.84	14.56	15.13
Fe_2O_3*	4.55	7.51	6.82	2.17	9.48
MnO	0.14	0.19	0.16	0.07	0.18
MgO	1.08	2.54	0.36	0.93	3.76
CaO	3.16	5.90	1.65	2.29	6.65
Na_2O	4.57	4.72	7.22	4.43	4.01
K_2O	3.12	1.54	4.10	3.11	2.17
P_2O_5	0.19	0.44	0.22	0.11	0.87
Total	99.61	99.97	99.93	99.79	99.45

Table 1. PD-1 is from Courthouse Quarry, representing the porphyritic phase of the Dedham; PD-8 is from Black Rock Beach and may represent the parental phase of the Dedham; PD-7 is grey aplite injected into the Dedham; PD-10 is average Cohasset/Westwood; PD-4 is Strawberry Pt. Quartz Diorite, associated with the Cohasset/Westwood Granite. Analysis was done by XRF at the University of Massachusetts.

*Total Fe as Fe_2O_3.

some places it is foliated or banded, in others it is porphyritic with phenocrysts of microcline up to two inches long. The Dedham is composed essentially of microcline (perthite), andesine, quartz, chlorite, and epidote. The most recent date for the Dedham (Zartman and Naylor, 1984) is one of 630+/-15m.y., by U-Th-Pb analysis.

The Dedham Granite is well represented in the Cohasset area. It is mostly the porphyritic granodiorite with large phenocrysts of pink microcline. Locally there is a dark green dioritic variety, which appears to be the parent rock to the granodiorite. Mineralogically and geochemically it seems to be calc-alkalic (see Table 1). The second granite is located south of the mylonite zone and for convenience has been named the Cohasset Granite. Also, because of the chemical similarity to the finer grained Westwood granite to the west, it will be referred to as the Cohasset/Westwood Granite. This granite appears to have been formed under different conditions than the Dedham Granite as it lacks a Europium anomaly and is depleted in the heavy REE (see Figure 2). It is a pink, finer grained granite with subhedral quartz grains and is less altered than the Dedham. It is composed of microcline (perthite), oligoclase, quartz, hornblende and epidote. Chemically it is a true granite (see Table 1).

The diorite (Strawberry Point Diorite) associated with the Cohasset is found only at Strawberry Point in Situate (see Figure 1). It is composed of zoned plagioclase, hornblende, chlorite, quartz, microcline and biotite. It may be comagmatic with the Cohasset/Westwood, but the relationship is still uncertain.

A third granite is exposed in the Hingham area (see Figure 1), referred to locally as the Weymouth Granite. It is a foliated biotite granite composed of 30-60% plagioclase altered to sericite, epidote and albite; 10-35% K-Spar; 5% secondary biotite. Chute (1965) believed this granite intruded the Westwood Granite (Brenninkmeyer, 1976). Its magmatic relationship to the other two granites in the area is not yet known.

Glacial History

The Pleistocene deposits in the Boston area are surprisingly poor in till and rich in outwash deposits. The Boston area seems to have been located at the margin of two major ice lobes. The direction of ice flow varies from southwest to east through an arc of 135 degrees. However, the flow directions can be divided into four separate groups (Kaye, 1976). These range in approximate order of occurence: A. Lobate, spreading to south and east (Back Bay Readvance), B. South (Beacon Hill Advance), C. Southwest, D. South and southwesterly (mainly 32, 22 and 16 degrees), E. Easterly and southeasterly (variable 80 to 38 degrees). The oldest of the glacial indicators, the deep wide grooves, are easterly.

Many of the drumlin axes show the same range in orientation. In several places the long axis of the drumlin diverges from the striations exposed in the bedrock beneath them. Moreover, although there are many classically shaped drumlins, there are many that are round in plan, as well as compound, curved and sigmoidal shapes.

Road Log

From Danvers follow Rt. 128 south. At the intersection of Rt. 128 and Rt. 3 follow Rt. 3 south to exit 14, Rt. 228, which is the Rockland-Nantasket exit.

Cummulative Milage	Interval Milage	
0.0	0.0	At the bottom of the ramp turn left (228 N) towards Hingham-Nantasket.
0.8	0.8	Assemble in the Star Market parking lot at 8 A.M.
0.9	0.1	Get back on 228 and cross the intersection of Rt. 53.
4.6	3.9	228 takes a sharp right, follow the signs
5.1	0.5	Bear right at the fork.
5.4	0.3	Bear to the left leaving the Hingham Library to your right.
6.5	1.1	Cross the intersection of Rt. 228 and Rt. 3A.
6.8	0.3	At the intersection of Rt. 228 and East St follow the signs for 228, leaving Glastonbury Monestary to your left.
8.0	1.2	At the intersection of 228 and Jerusalem Rd., 500 ft. past "With Richards" restaurant, turn right onto Jerusalem Rd. Notice the sharp rise in elevation to the right caused by the upthrow of the Ponkapoag Fault
9.0	1.0	Turn left into Wadleigh Park Rd. and stop. Walk across Forest Ave Extension to Black Rock Beach.

Stop 1: Blackrock Beach/Green Hill

At low tide two phases of the Dedham Granite are exposed in a wave polished outcrop, fifty feet to the north of the main exposure. The first phase is a dark green, medium grained diorite, composed of plagioclase, microcline, hornblende, chlorite, quartz and minor sphene and magnetite. There are a number of xenoliths of fine grained diorite ranging in size from 10 to 30 cm. The second is porphyritic Dedham Granodiorite. This second phase is the same as at Stop 3, while the diorite is only present locally. Both phases have been foliated, giving them a banded appearance. This foliation generally strikes N20W, which is perpendicular to the local faulting. The major element and REE patterns (See Table 1, Figure 2) indicate that they may have been generated from the same magmatic series. It is also a possibility that the diorite is the parent of the granodiorite, if hornblende was a residual phase during fractionation, which the REE patterns seem to indicate.

Continuing north on the beach towards Green Hill, we cross the Ponkapoag Fault zone and enter the Boston Basin. The fault is orientated N 75 degrees East and dips 80 degrees north. Billings (1982) now believes that the Ponkapoag fault is an unrotated normal fault downthrown to the northwest. Notice the large boulders, up to 5 m in diameter, scattered in front of Green Hill, which is a drumlin. These boulders were eroded from the drumlin before the seawall was erected. Many of the boulders are Roxbury Conglomerate, and the porphyritic phase of the Dedham.

Pass the sea wall at Green Hill and continue North, approximately 500 m further. Along the beach is an outcrop of the Dorchester member of the Roxbury Conglomerate. There are no other exposures of the Roxbury from this point south. There are a number of clasts of weathered Dedham here. Many of them are added to the beach as the less resistant sandstone matrix is eroded away by wave action.

Cummulative Milage	Interval Milage	
10.3	1.3	At the ocean side of Wadleigh Park Rd. turn left onto Atlantic Ave. Follow it until it intersects Nantasket Ave. Turn right onto Nantasket Ave.
10.4	0.1	Turn right into the MDC parking lot as close to "On the Rocks", bar as possible. Walk onto beach and around the building.

Stop 2 Atlantic Hill

At the south end of Nantasket Beach is Atlantic Hill. Here a good cross section of the basal volcanic units of the Mattapan Volcanics and the Boston Bay Group Sediments are exposed. These units have been brought to the surface by the Blue Hills Thrust Fault (see figure 1). Atlantic Hill is on the very south east margin of the Boston Basin within the southern limb of a north east plunging anticline.

Just left of the bar "On the Rocks" is a hard dense greenish gray andesite lava about 18 m thick. These andesites are pillowed, as outlined by the lighter green epidote veins. They also contain bombs. Above the andesite is a 9 m bedded tuff (strike N65E, dip 25SE). This layer is a water lain ash deposit. In the upper part of this unit there are many coarse volcanic fragments. Above this are thin lenticular beds of andesite and tuff. The hill is capped by another greenish grey andesite lava. These andesites are deuterically altered. Plagioclase, chlorite and epidote are the major constituents plus accessory magnetite. There is little primary quartz, but quartz and calcite may be abundant as secondary minerals (Bell, 1964, Skehan, 1975).

Below the andesites and best exposed at low tide on Long Beach Rock is the sedimentary sequence of the Boston Bay Group. The lowest units, to be seen only at low tide on the north side of Long Beach Rock, are tuffaceous conglomerates and agglomerates which include fragments of the underlying Dedham Granodiorite and Lynn Volcanics and also arkose boulders. Above this are 45 cm thick beds of intercalated red sandstone and 15-30 cm layers of banded green porcelanous shale.These thin beds show brecciation and penecontemporaneous faulting and baking. The contact between the volcanics and the sediments can also be seen at this base of the cliff (around the corner from the bar).

The dikes, especially those on Long Beach rock are parallel to the local faulting and seem to be related to the volcanic activity that produced the andesite flows. The dikes predate the faulting (Crosby, 1893).

Return to Nantasket Ave. Turn left.

Cummulative Milage	Interval Milage	
10.7	0.3	Follow Nantasket Ave. (Rt. 228). Turn right.
10.9	0.2	Bear rght at the intersection of Rt. 228 and Rockland St., leaving the Mobil gas station to the left.
12.2	1.3	Follow Rockland St. until it intersects George Washington Blvd. Observe the large glacial erratic to the left enclosed in a small park. A plaque is embedded in one side, dedicating it to W.O Crosby and making the erroneous statement that it had been transported thousands of miles.
12.7	0.5	Turn right onto George Washington Blvd. and drive to the large parking lot located just past the Hingham District Court house. Pull into the far end of the lot and park near the overgrown dirt road. Walk down the road into the quarry.

Stop 3: Courthouse Quarry

The quarry shows the granodioritic phase of the Dedham which is the same green porphyritic rock, with large pink microcline phenocrysts described by Chute (1966), in North Abington. It is also the same as the second phase seen at Stop 1, except it is not foliated. It is composed of andesine, orthoclase, quartz, chlorite, sericite and minor amounts of hornblende, apatite, sphene, epidote, magnetite and flourite. The chlorite and epidote are probably alteration products of biotite and hornblende, as the rock has been extensively fractured. The fractures have also been filled with chlorite and epidote.

Nellis and Hellier(1976) note several different dike sets here, which are best exposed in the northwest wall of the quarry. One set is composed of dark green diabase and ranges in thickness from a few centimeters up to two meters. This set may be related to either the Brighton Volcanics, making them Precambrian, or to the Triassic dikes of the Conneticut Valley (Billings, 1976). It is also possible that some are Precambrian and some are Triassic; at this point there is no definite answer to this problem. The second set of dikes is aplitic. They range in color from pink to gray and from half a centimeter to 30 centimeters in thickness. This set is chemically similar to the wall rock (see Table 1, Figure 2) although the slight enrichment of the light REE may indicate a more evolved magma. In any case, it appears that this dike set originated from the same magma as the granodiorite and that the aplite was injected into the granodiorite while it was cooling.

Cummulative Milage	Interval Milage	
13.1	0.4	From the parking lot turn left onto Washington Blvd. Bear right onto Rockland St.
13.9	0.8	Follw Rockland St. to Hingham Rotary. Continue to follow Rt. 3A north.
15.6	1.7	Follow 3A leaving "Pages" restaurant to the left, through a stop light and around a sharp left turn, to the intersection of 3A and Downer Ave.
15.8	0.2	Turn right onto Downer Ave. and follow it to Planters Field Rd, across from the Foster School.
16.0	0.2	Turn left onto Planters Field Rd. and follow it to Wompatuck Rd.
16.1	0.1	Turn left onto Wompatuck Rd until you reach Kimball Beach Rd.
16.5	0.4	Turn left and follow Wompatuck Rd until it meets the ocean. Park and walk to the beach.

Stop 4 Cambridge Argillite

This stop can be reached only at low tide. The outcrop is exposed approximately 100 meters north of the end of te road. This is the only exposure of Cambridge Argillite in this area of the basin, although, it can also be found across the Bay on Grape Island and on some of the islands north of Nantasket. This exposure is typical of the Cambridge with alternating dark grey argillite and light gray siltstone, the beds strike N70E and dip 70S. Note the small beach above the outcrop is composed mainly of the argillite and the nearby volcanics, with only a small percentage of granite

Cummulative Milage	Interval Milage	
17.4	0.9	Return to 3A.
19.1	1.7	Turn left on 3A and drive south to the Hingham Rotary. Follow the signs for 3A South to Cohasset and Scituate.
24.7	5.6	Follow 3A south, through three stop lights, past a small shopping plaza on

		both sides of the road, followed by several car dealerships, to the exit for Minot and North Scituate.
25.2	0.5	Turn left onto Bailey Rd. and follow it until it intersects Country Way at a stop light.
26.7	1.5	Go straight through the lights, the road changes its name and now is Ganet Rd. Follow Ganet Rd bearing right at the fork until the intersection with Hatherley Rd.
27.0	0.3	Follow Ganet Rd. until it ends and then turn left onto Glades Rd.
27.8	0.8	Follow Glades Rd. until the gates of the Adams Estate. SPECIAL PERMISSION MUST BE OBTAINED BEFORE ENTERING.
28.0	0.2	Drive into the estate until you reach a small dirt path on the ocean side of the road.

Stop 5 Cohasset Granite and Strawberry Point Diorite

This stop requires about a half mile walk through somewhat rugged teraine. Excellent examples of both the Cohasset Granite and the Strawberry Point Diorite, can be seen here. Walk towards the ocean on the short dirt path into the abandoned quarry. The Cohasset here is a medium grained, pink to gray granite containing microcline (perthite), oligoclase, quartz, chlorite, hornblende, epidote, magnetite, apatite and sphene. The quartz is usually well formed in subhedral crystals and tend to be equigranular with the feldspars. In several spots, there are fine grained xenoliths of diorite, approximately 30 cm in diameter.

Walk north out of the quarry and along the shore. As you approach the contact with the diorite, the xenoliths become more common. About a thousand meters out of the quarry, the dioritic xenoliths become larger some up to two meters in diameter. The contact between the granite and the diorite is ten meters further north, just south of the gray house on the point. The diorite is composed of zoned plagioclase, hornblende, chlorite, quartz, microcline, biotite and sulfides. It has a dike like apperance in thin section and most of the unit has been hydrothermally altered. The contact here is sharp but neither margin is chilled.

Continiung north across the diorite, there are a number of large xenoliths of the Cohasset in the Diorite. At approximately 1500 m, there is a large xenolith of Cohasset (greater than one meter in length) that incorporates a xenolith of diorite (20 cm in diameter), this initial evidence seems to indicate that the two units were comagmatic, but wether they belong to the same magma series is still uncertain. Continue north to the contact with the Cohasset and then the short walk back to the cars along the road.

Cummulative Milage	Interval Milage	
28.5	0.5	Return through the gate and follow Glades Rd to the first road on the right (Baileys Causeway).
28.8	1.3	Follow Baileys Causeway until it intersects Hatherley Rd.
30.1	1.3	Turn left on Hatherley Rd. Go through the intersection with Garnet Rd. Follow Hatherley until you reach Mann Hill Rd.
30.1	0.2	Turn left onto Mann Hill Rd. and follow it to the intersection of Stanton lane. Pull off the road as far to the left as possible. We have to turn around at the bottom of the road.

Stop 6 Mann Hill Beach

Mann Hill Beach is a shingle bay mouth bar. At low tide a sand bar stretches across the entire length of the beach. The shingle reaches 10 m above low water and shows well developed imbrication, several berm lines and cusps. The pebbles and cobbles of the shingle are both size and shape sorted.

The sediments at Mann Hill ranges from fine sand to boulders over 30 cm in size. On the lower foreshore fine sand is dominant. On the middle foreshore, halfway between low and high tide, pebbles start to make their appearance. On the upper foreshore shingle is found. The average of all the shingle sizes, based on the 30,000 measurements of the largest diameter, is 4.3 cm. The average intermediate diameter is 3.14 cm, and the smallest diameter is 1.65 cm.

There is a gradual increase in size from the upper foreshore to the berm. The largest sizes are found at the berm. Generally, a decrease in size is found landward of the berm. If more than one berm is present, as is often the case, there is a small decrease in size just landward of the berm in the swale.

Along the beach there is a gradual increase of shingle size from north to south. In the north, the average size is 3.1 cm, while in the south it is 5.3 cm. The only places where there are local variations on this along beach increase in sediment size is within the cusps. Inside the cusp bays the coarsest sediment is found in the landward end and on the north side.

Shingle Lithology

There are eleven common lithologies found in the pebbles on Mann Hill Beach. As can be seen in table two, the most frequently found lithology are the granites. Three types of these are found. The most common is the foliated, light pinkish grey medium grained Dedham granodiorite, which has 25% quartz, 10% altered biotite and the remainder of the minerals sauseritized plagioclase and orthoclase. In the Dedham there are many xenoliths of two types of diorite and a fine grained amphibolite. Less commonly found is

Table 2 Shape differentiation between the thirteen common pebble and cobble lithologies at Mann Hill Beach.

	Blade	Ellip Plate	Ellipsoid	Needle	Plate	Short Rod	Sphere	Thick Blade	Thick Plate
Dedham	0	1	27	1	1	18	15	1	36
Cohasset-Westwood	0	1	25	1	1	17	16	2	36
Weymouth	0	0	31	2	3	15	10	1	38
Diorite in Dedham	0	2	35	1	6	10	7	2	37
Strawberry Pt Diorite	0	2	37	2	9	9	3	2	36
Mattapan Andesite	1	3	36	2	5	11	5	2	36
Amphibolite in Dedham	0	2	26	4	9	17	6	4	33
Basalt	1	6	34	2	8	10	3	3	33
Roxbury Conglomerate	1	5	17	1	9	11	6	6	42
Milton Quartzite	0	1	31	2	3	17	10	1	36
Mattapan Rhyolite	1	1	32	2	3	14	7	1	39
Cambridge Siltstone	1	10	36	1	13	5	2	7	25
Cambridge Sandstone	1	8	30	2	9	11	4	3	33

Table 3 Percentage Distribution of the Size in centimeters of the Different Lithologies of the thirteen common pebble and cobbles at Mann Hill Beach.

	1	1-2	2-3	3-4	4-5	5-6	6-7	7-8	8-9	9-10	10
Dedham	1	10	35	30	13	6	3	2	1	1	1
Coh./Westwood	1	17	39	23	11	5	2	1	1	1	0
Weymouth	1	13	28	28	14	7	3	4	2	1	0
Dio in Dedham	1	11	36	28	15	5	2	2	1	1	1
Straw. Pt Dio.	0	1	46	28	14	8	3	1	1	0	0
Mattapan And.	1	18	36	25	12	5	2	1	1	1	1
Amphi. in Dedham	2	19	35	20	13	6	4	2	0	0	0
Basalt	1	17	39	26	11	5	1	1	1	0	1
Roxbury Congl.	1	11	30	34	14	5	1	3	1	0	0
Milton Quartzite	1	18	40	23	10	5	3	1	0	1	1
Mattapan Rhy.	1	20	38	24	11	4	1	1	1	1	0
Cambridge Silts.	1	19	39	23	11	5	1	1	1	1	1
Cambridge Sands.	0	13	33	28	15	6	3	1	1	1	1

the Cohasset-Westwood granite. This granite is pinkish grey, fine to medium grained with 25-35% quartz, 10% microperthite and accessory apatite, sphene and magnetite (Chute, 1965). Even rarer is the Weymouth Granite, which is medium grey, foliated biotite granite composed of 30-60% plagioclase which is almost completely altered to sericite, epidote and albite, 10-35% orthoclase and microcline, and 5% olivine green biotite and large sodic plagioclase phenocrysts (Chute, 1965).

The siltstone is a greenish to brownish grey, fine grained, thin bedded argillite which contains quartz, sericite and opaque minerals (probably graphite) in a finer matrix too small to be identifiable. The siltstone probably belongs to the Boston Bay Group, specifically the Cambridge Argillite or possibly the Braintree Argillite.

The sandstone is obviously coarser, but has the same appearance as the argillite. It is not half as common as the argillite. The sandstone owes its origin to the Boston Bay Group in the Roxbury Conglomerate and Cambridge Argillite found only north of the Ponkapoag fault.

Both the rhyolite and andesite are part of the Mattapan/Brighton Volcanic Complex. The rhyolite is the most common rock found on Mann Hill. The rhyolite is predominantly red and purple, but may be brown tuffs and flows usually with small phenocrysts of quartz and microperthite. Most of the "rhyolite" is so fined grained it is probably a devitrified tuff. The andesite is the second most common lithology found on Mann Hill Beach. It is bluish to greenish grey with small phenocrysts of sericitized plagioclase, quartz and chlorite in groundmass of very fine plagioclase laths. Secondary epidote is common, giving the rock a greenish tinge.

Although only less than one percent of all the rocks on Mann Hill are recognizable as Roxbury Conglomerate, there should be many more. Many of the Dedham,andesite, rhyolite, quartzite and amphibolite pebbles are probably second generation, having first been present as pebbles in the Roxbury.

The basalt is a medium to dark grayish green, fine grained, massive rock. It frequently has subophitic texture. The basalt probably came from the diabase dikes so prevalent throughout the neighboring quadrangles to the northwest.

The first diorite is a medium to coarse grained massive dark rock, greenish-grey in color consisting predominantly of plagioclase which is almost completely altered to epidote, albite and sericite with lesser amounts of hornblende, a light pyroxene and olive green biotite. The second diorite is very similar except that it is much finer grained and even darker in color. The first diorite resembles the diorite within the Dedham. The second is a dead ringer for the Strawberry Point Diorite.

The quartzite is light grey to white, massive quartzite with minor quantities of biotite. It probably owes its origin to the "Milton Quartzite" of Billings (1929) or the quartzite pebbles in the Roxbury Conglomerate.

Least common of the rocks is the greyish green amphibolite in which the layers have fine alterations of felsic and mafic minerals producing a striped appearance. Hornblende and plagioclase are the principle components with minor amounts of chlorite, quartz, epidote, sphene and calcite. The amphibolite is common only as xenoliths in the Dedham granodiorite.

On Mann Hill Beach, all the lithologies are not admixed equally. There is more rhyolite and quartzite on the lower fringes of the shingle and more siltstone and andesite on top of the berm. What is remarkable is the size distribution of the different lithologies. Each lithology has virtually the same distribution of their intermediate axes (see table 3). The only one that is a tiny bit different is the finer grained Strawberry Point Diorite.

Shingle Shape

Anyone who has visited a shingled beach has seen the differences in the shape of the pebbles between those at mid tide and those at the top of the berm. A description of shape or geometric form involves several different but related concepts. On one side are the shape factors which depend on the relative lengths of the particle with respect to standard Cartesian coordinates. On the other hand is the sediment's angularity or roundness.

In 1935 Zingg showed that if the ratio of intermediate to the maximum lengths (B/A) of a pebble is plotted against the ratio of the shortest to the intermediate lengths (C/B), the particle may be classified according to its shape. Zingg utilized four shape classes. We have modified this to nine shapes, see figure 3.

The proportion of flat shapes at Mann Hill Beach is surprising, especially in view of the large proportion of non-layered rocks composed of equidimensional grains (granite, diorite, rhyolite and quartzite). However, the Boston Basin and surrounding areas are characterized by thrust faults, tear faults and normal faults. Especially in the southern part of the basin, the structure consists of numerous thrusted anticlines which constitute an imbricate block structure with minor thrusts and rock slices (Billings, 1929). Locally numerous shears are present, spaced at intervals between 1-100 cm apart. Moreover, two types of cleavage are prevalent. These ruptures are uniformly perpendicular to the bedding and to the imbricate blocks. Also nearly ubiquitous is a set of remarkably parallel joints spaced at intervals of 1 cm to 15 m. When two or three sets of joints are closely spaced, the rocks break into parallelopipeda 5-50 cm on a side (Billings, 1976). These, after being eroded by glacier ice and deposited in drumlins or outwash, are rounded by wave induced transport to form the flat shapes. The Wadell's (1933) operational sphericity follows the frequency of occurence pattern. The white area is bordered on the bottom by a sphericity of 0.5 on the bottom and by 0.9 on the top.

The distribution of the shapes is lithology dependent. The Cambridge Argillite has the most elliptical plates and the granites have the most spheres (see Table 2). The majority of the pebbles could not have been transported very far either by the ice or by waves. The remarkable similarity in size distribution shows there is relatively little abrasion and that the sizes are

MODIFIED ZINGG DIAGRAM

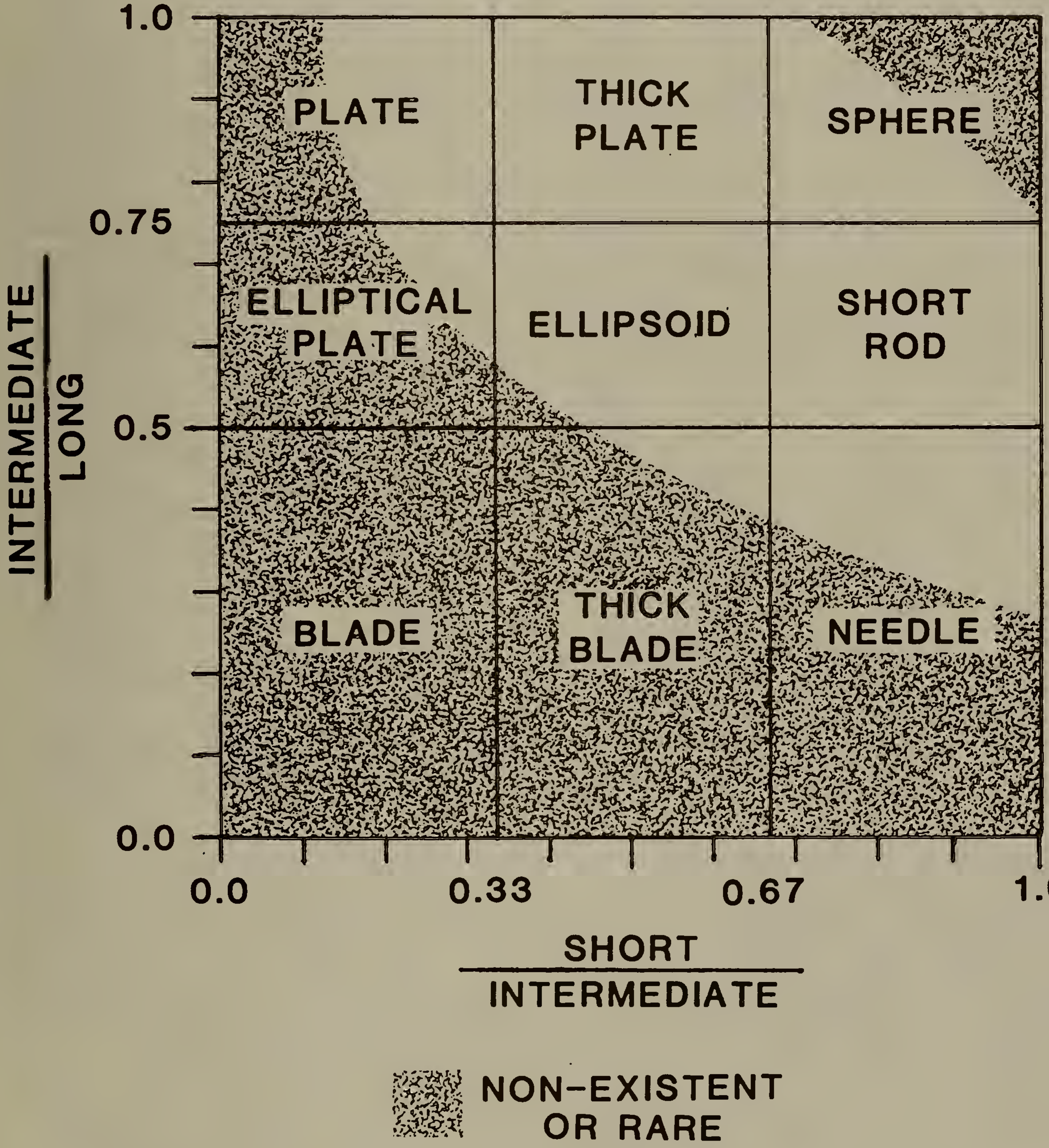

FIGURE 3

inherited from the faulting and jointing.

In the over 30,000 measured and identified pebbles, we have never seen the distinctive grey to bluish grey Quincy Granite, nor the fine grained riebeckite granite (Chute, 1969), both of which are exposed 15 km to the northwest in between the Blue Hills Thrust and Ponkapoag Fault. The closest Mattapan Volcanics are exposed 7.5 km to the north. If the faults are extended out into the ocean the closest the Mattapan could be is 6 km. The farthest from the source are the Weymouth Granite pebbles, which came from 15 km due west.

Cummulative Milage	Interval Milage	
31.5	1.2	Return to the intersection of Ganet Rd. and Hatherley Rd.
33.4	1.9	Turn left onto Ganet Rd. and return to Rt. 3A.
38.3	4.9	Turn south on 3A. Cross the intersection of 3A and 123. After going over the North River Bridge, take the first left onto Summer St., to Hummerock and Fourth Cliff.
40.8	2.5	Follow Summer St., bearing left at the intersection with Prospect St., remain on Summer St Bear right at the fork with Damon Pt. Rd., continue to the intersection of Church Rd. and Summer St.
40.9	0.1	Turn left onto Elm St, bearing right at the fork of Elm and Ferrywell Sts., remaining on Elm.
41.0	0.1	Bear right at the fork of Peabody and Elm, remaining on Elm.
41.4	0.4	Turn left onto Sea St.
41.6	0.2	Go over the bridge and turn left onto Central Ave.
43.1	1.5	Follow Central Ave to River Rd. Park off the road and follow the boardwalk to the beach. Walk north.

Stop 7 Fourth Cliff

Fourth Cliff is a 24 m high, 0.8 km long drumlin. This is the most southerly of the over 150 drumlins of the Boston area. Fourth Cliff consists of 12.30 cm of soil and 9 m of brown oxidized gray till that contain some reed-like plant remains. These have been dated at 35,000 B.P. (Chute, 1965). The typical amount of silt and clay in the till is 19% with a maximum of 40%. This is an unusually low percentage for eastern Masachusetts drumlins.

On the east side of the drumlin two lenses of sand and gravel 3-4.5 m thick separated by 3 m of till crop out. These lenses dip 10 degrees to the south and appear to pinch out at the bottom of the cliff. The composition of the pebbles and cobbles in the till is primarily Dedham, Westwood and Weymouth Granites and Mattapan Volcanics. The Cambridge Argillite is no longer present in large quantities. The sand is 60% quartz, 20% feldspars and 20% heavy minerals such as biotite, magnetite, garnet and hornblende.

On the beach are many boulders and cobbles, some of which are so large that even storm waves have difficulty moving them. Among the largest boulders, two are the Cohasset-Westwood Granites, one is from the Diorite Dike at Strawberry Point in the Cohasset, three are Dedham Granites and one is a Roxbury Conglomerate. The glacial transport of these is 6 km from the SSW and 9 km to the SE; the diorite was transported a minimum of 8 km from the SSE. The closest Roxbury is 13 km away to the NNW. The Weymouth Granite was transported just over 15 to the SE. As you walk along, make sure that you see the large (1.2m) Dedham Granite boulder with the oriented xenoliths of amphibolite. On the northeast point there is a boulder pavement. It has an exposed width of 60-100 m at low tide. Of interest is that there seems to be little or no sand movements across this boulder platform below mid-tide level. The boulders below this level show little or no abrasion; instead they are covered with barnacles and seaweed and many show evidence of abrasion. There are few weathered surfaces or flora and fauna that could survive above mid-tide level. This scouring is caused by sand and shingle which is washed back and forth at the still stand of the high tide. Above the mid-tide level the shingle extends about 4km south of Fourth Cliff. Beyond this, the beach is composed entirely of sand.

Before the 1900's there existed a barrier between Third and Fourth Cliffs. The mouth of the North River was 4.8 km south of its present location. Then on November 27th, 1898, with a high tide of 4.5 m (1.3 m above normal) and a wind of 120-130 km/hr piling waves up even higher, the ocean cut through the beach ridge between the cliffs. In a few hours a channel 45 m wide and 3 m deep had been excavated. Now the channel is 120 m wide and 4.2-4.8 m deep. The average flood tidal velocity through the gap is 24 cm/sec. The ebb tide velocity is 36 cm/sec.

It took three years of longshore drifting to fill the old river entrance. The result is that the South River now flows further north and has developed a sand bar which extends into the North River which recurves it. The bar, now, is almost a reverse mirror image of Cape Cod.

The spit behind Fourth Cliff can be divided into two distinct parts. The first part is adjacent to the drumlin. It is fronted by a low scarp 60 cm high and is composed almost entirely of boulders and cobbles. These cannot have come from the drumlin immediately behind the beach for the sizes there are much smaller. Instead, they must have been carried by storm waves around the northeast point.

About 200 m southwest of the point there is a sudden change in the size of the shingle. The particles become much smaller and sand becomes dominant. This may demark the boundary of the effect of storm waves. Also there is a scarp of old marsh grass (Spartina patens) there and only the largest storm waves can lift cobbles over this resistant scarp.

From there south westward, sand predominates. The development of the spit has dammed up the sediment coming down the South River. This is now deposited in a sand bar. This bar where it extends into the North River, recurves back to Fourth Cliff. At the north west corner, the bar has become anchored by extensive beds of mussels. The top of the bar is practically devoid of vegetation or animal life. But sand waves, current ripple marks and rhomboid ripples and rill marks are common.

Cummulative Milage	Interval Milage	
47.9	4.8	Return to Rt. 3A via the same route in reverse.
49.4	1.5	Turn right on 3A and head North to 123
55.1	5.7	Take a sharp left onto 123 and follow the signs to Rt. 3. Follow Rt. 3 North to Boston.

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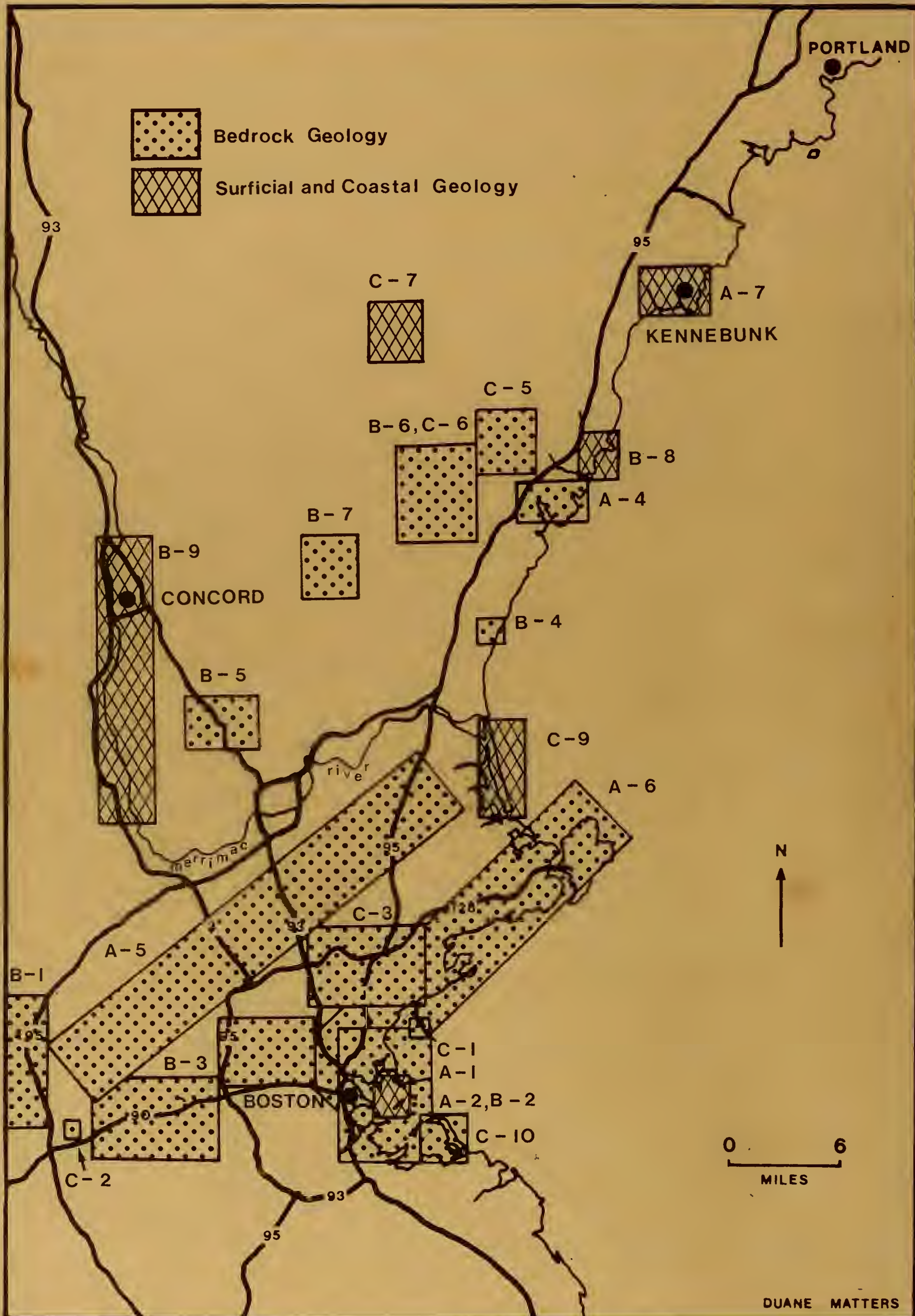
AN ABBREVIATED INDEX
OF FIELD TRIPS INCLUDED IN THIS GUIDEBOOK



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